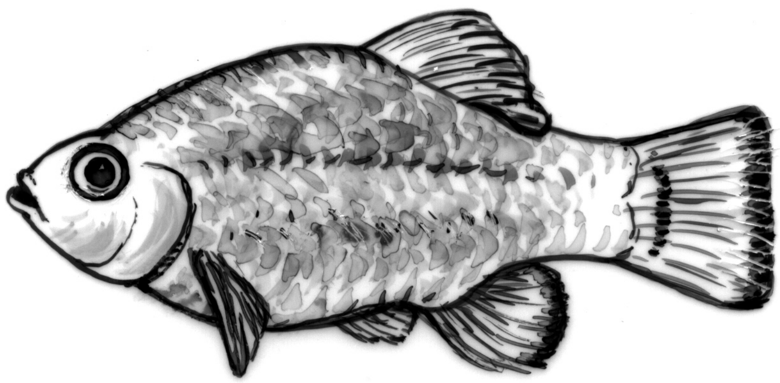
Wild, scenic and rapid:
a trip down the Colorado River trough

Robert E. Reynolds, editor



Wild, scenic & rapid—the field trip guide

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Abstracts from the 2007 Desert Symposium

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Front cover: view west across the Colorado River extensional corridor from Cottonwood Summit. R.E. Reynolds photograph.

Back cover: the Bouse Formation at Earp, California. D.V. Malmon photograph.

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Wild, scenic and rapid trip down the Colorado River trough

Desert Symposium Field Trip 2007

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This rapid trip will explore wild fluvial and tectonic events resulting in scenic and rugged topography. The extreme differences in elevation caused valleys to be choked by alluvium and incised by the Colorado River drainage system.

We will visit faulted volcanic flows and tilted plutons (Stops 1.1–1.6, 1.12) and will see the Bouse Formation of Cottonwood and Mohave valleys that marks depositional history at about 5 Ma (Stops 1.7–1.11). Be sure to look at the clasts and cobbles in the fluvial deposits. If the clasts are subangular, they were probably carried along tributaries from local sources. If clasts are well-rounded, they have probably been carried long distances. The lithology of the clast may tell you if the source is from local mountains or from the distant Paleozoic sections on the Colorado Plateau. Limestone bedrock exposed along this part of the Colorado River trough is primarily Tertiary volcanic rocks, ranging from basalt to rhyolite, and Tertiary and Precambrian granitic rocks. If you see well-rounded limestone, dark oor light-colored chert, and quartzite cobbles, they were likely transported by the Colorado River from the plateau in the past 5 million years (House, this volume; Longwell and others, 1965; Metzger and Loetz, 1973a,b).

Miocene extensional tectonics

The Colorado River extensional corridor is a 70- to 100-km-wide region of moderately to highly extended crust along the eastern margin of the Basin and Range province in southern Nevada, western Arizona, and southeastern California (stretching from the Lake Mead area southward to the Parker, Arizona area). The corridor has occupied a critical structural position in the western Cordillera since Mesozoic time. From the Cretaceous through early Tertiary, the corridor stood just east and north of major fold and thrust belts and also marked the northern end of a broad, gently (~15°) north-plunging uplift (Kingman arch) that extended southeastward through much of central Arizona. Mesozoic and Paleozoic strata were stripped from the arch by northeast-flowing streams (Faulds and others 1992, 1995, 1998, 2001a). Peraluminous 65 to 73 Ma granites were emplaced at depths of at least 10 km and were exposed in the core of the arch by earliest Miocene time.

Calc-alkaline magmatism swept northward through the northern Colorado River extensional corridor during early to middle Miocene time, beginning at ~22 Ma in the south and ~12 Ma in the north (Faulds and others, 2001a). Major east-west extension followed the initiation of magmatism by 1 to 4 m.y., progressing northward at a rate of ~3 cm/yr. The style of volcanism changed during the course of east-west extension. Eruptions of calc-alkaline to mildly alkaline mafic to intermediate magmas predated extension. Calc-alkaline to mildly alkaline mafic, intermediate, and felsic magmas were prevalent during major extension. Tholeiitic and alkalic basalts were then erupted after significant block tilting. The most voluminous volcanism occurred in early Miocene time and was accompanied by mild north-south extension. Large-magnitude east-west extension engulfed nearly the entire region in middle Miocene time, beginning about 23 Ma and ending ~18 Ma in the south and by 9 Ma in the north. Tilt rates commonly exceeded 40°/500 kyr during the early stages of east-west extension. Volcanism generally spanned the entire episode of extension south of Lake Mead. Thus, thick volcanic sections, as opposed to sedimentary rock, accumulated in many growth-fault basins.

During middle Miocene extension, strain was partitioned between a west-dipping normal-fault system in the north and an east-dipping system in the south. The two fault systems and attendant opposing tilt-block domains overlap and terminate within the generally east-northeast-trending Black Mountains accommodation zone. Major east-west extension was contemporaneous on either side of the accommodation zone. The west-dipping normal fault system in the north is kinematically linked to major strike-slip faults along the northern margin of the corridor, where a complex three-dimensional strain field, involving both east-west extension and north-south shortening, characterized the middle to late Miocene. The east-dipping system in the south linked southward with major east-dipping detachment faults that flanked several metamorphic core complexes (e.g., Chemehuevi and Whipple mountains).

The transition between the Colorado Plateau and the Basin and Range is unusually sharp along the eastern margin of the northern Colorado River extensional corridor and is marked by a single west-dipping fault zone, the Grand Wash fault zone. Subhorizontal, relatively unfaulted strata on the Colorado Plateau give way to moderately to steeply east-tilted fault blocks across the Grand Wash fault zone.

Topography and valley filling fanglomerates

Topographic and structural relief in the Colorado River trough began to develop during middle Miocene extension and was established by 9 Ma. The location and abruptness of the Colorado Plateau–Basin and Range transition in this region may have been controlled by an ancient north-trending crustal flaw, inasmuch as it follows a diffuse boundary between Early Proterozoic crustal provinces.

The Colorado River arrived in the northern Colorado River extensional corridor in latest Miocene–early Pliocene time following a long period of internal drainage (Faulds and others, 2001b). Before that, thick sections of alluvial fan, continental playa, and lacustrine deposits accumulated in major basins until ~6 Ma. Widespread late Miocene lacustrine and evaporite deposits in the Lake Mead region suggest a large influx of fresh water. However, available age constraints indicate that a through-flowing Colorado River developed between ~6.0 and 4.4 Ma (Faulds and others 2001a; House and others 2005 and this volume).

Day One

Convene at the Desert Studies Center at Zzyzx with a full tank of gas for the 212 mile trip on Day 1. Day 2 will be 166 miles and Day 3 will be 150 miles for a total of 530 miles, ending at Vidal Junction on CA Hwy 95. Wear sturdy shoes and dress for the occasion; bring water, breakfasts, lunches, snacks, hats, and sunscreen. High-clearance vehicles are mandatory; four-wheel-drive is recommended. Dine at local restaurants. Drive north to Interstate Highway 15, and enter eastbound toward Baker.

0.0 (0.0) Baker

12.6 (12.6) Halloran Springs. The composition of the Halloran volcanics to the north and the Cima volcanics to the south (Fig. 1-1) vary through time (Jessey and Reynolds, this volume). The Halloran Hills are the site of Neogene (12.1 Ma) andesitic volcanism, while the Cima volcanics are composed of older basaltic trachyandesite (7.5-3 Ma) and younger trachybasalt (l Ma–Present). This suggests an extended pattern of alkaline volcanism beginning in the initial stages of Miocene detachment that has continued to the present. Compositional variation reflects progressively deeper levels of melting, from shallow crustal rhyolite and trachydacite, to lower crustal trachyandesite, to mantle trachybasalt.

Turquoise mines of West Camp, Middle Camp and East Camp are located east and west of Turquoise Mountain to the north-northwest. The Turquoise Mountain area has been the subject of gemstone mining for 1,000 years (Reynolds, 2005).

18.5.(5.9) Halloran Summit. View southeast of Cima Dome, underlain by one of the granitic blocks tilted eastward (Fig. 19, Reynolds and others, 1996) by down-to-the-west listric normal faulting due to accompanying movement on the Kingston Range–Halloran Hills detachment fault that runs north along the western front of the Mescal Range and Clark Mountain (Reynolds and Calzia, 1996; Friedmann, 1996; Fowler and others, 1996).

25.5 (7.0) Cima Road. View north of the Pliocene and Pleistocene white-colored pond and ground water discharge sediments of the Valley Wells basin (Reynolds and others, 1991; Forester and others, 2003; Reynolds and others, 2003; Pigati, this volume). Farther north, the large cottonwood tree marks the site of the Valley Wells Copper Smelter (Rosalie post office, Vredenburgh, 1996).

34.0 (8.5) Bailey Road at Mountain Pass.

38.7 (4.7) EXIT at Nipton Road.

Fig. 1-1. The Cima volcanic field (JPL courtesy John Ford).

39.0 (0.3) Stop, TURN RIGHT (south); the road bears east.

42.5 (3.5) Continue past Ivanpah Road, running south to Cima, Lanfair, and Goffs on Route 66.

49.1 (6.6) SLOW, cross railroad tracks, and enter Nipton.

51.6 (2.5) Cross the California/Nevada state line.

53.6 (2.0) Continue past right turn to Crescent Spring and Crescent Peak.

54.5 (0.9) View at 2:00 of Crescent Peak. Lead and silver mines are associated with turquoise (Castor and Ferdock, 2004.). When rediscovered in 1889, Native American stone hammers, anvil stones, and pottery were found.

58.5 (4.0) Continue past a right turn to a microwave station.

63.3 (4.8) Continue past a right turn to Walking Box Ranch Road leading to Lanfair Valley.

70.1 (6.8) Slow as you enter Searchlight, NV. Replenish gas and supplies. Stop at Hwy 95. PROCEED EAST across Hwy 95 on AZ Hwy 164 (Cottonwood Cove Road).

70.8 (0.7) Continue past the history museum at the community center.

78.1 (7.3) Enter the Lake Mead National Recreation Area.

79.6 (1.5) Continue past the south tip of Black Mountain, on the left (north). Black Mountain consists of middle Miocene trachyandeste-trachydacite flows (~14-16 Ma).

80.3 (0.7) **STOP 1-1**. Introduction to Colorado River deposits and evolution. PULL RIGHT at turnout and park to view a major east-dipping detachment fault,with 10–15 km of slip. Stop objectives include discussion of a major west-tilted fault block that exposes 15 km-thick cross section of crust, including the nearly 10 km-thick Searchlight pluton; the Searchlight hydrothermal system; and the 18–16 Ma stratovolcano complex above the Searchlight pluton.

The prominent jagged peak to the northwest is Copper Mountain, where a splay of the Dupont Mountain detachment fault is exposed. The detachment is marked by the gently east-dipping color change between the reddish-brown brittlely deformed upper Searchlight pluton in the hanging wall and greenish-gray ductilely deformed lower Searchlight pluton and Early Proterozoic gneiss in the footwall. The Dupont Mountain fault is also exposed at Dupont Mountain, which is the low pyramidal-shaped peak in the prominent saddle to the north-northwest.

The footwall of the Dupont Mountain detachment comprises a large steeply west-tilted fault block that extends from the Copper Mountain area westward to the southern Highland Range. This fault block constitutes much of the southern Eldorado Mountains and exposes a cross-sectional view of approximately 15 km of crust (Faulds and others, 2001a). Much of this fault block is composed of new magmatic crust in the form of the ~16.5 Ma Searchlight pluton, a 10 km thick quartz monzonite to granite pluton (Bachl and others, 2001). Paleomagnetic data indicate that the pluton is tilted at least 55º to the west (Faulds and others, 1998). The Searchlight pluton invades Early Proterozoic gneiss and a 64–65 Ma peraluminous granite in its lower reaches and the lower part of the Miocene volcanic section in the upper hypabyssal part. An early Miocene brittle-ductile transition is exposed within the pluton in the southern Eldorado Mountains. The floor of the pluton is exposed on the west flank of Copper Mountain. The roof of the pluton is exposed near the town of Searchlight. The core of a steeply tilted 18–16 Ma stratovolcano overlies the pluton in the Searchlight area and is clearly genetically related to the pluton. Mineralization in the Searchlight mining district appears to be associated with a hydrothermal halo above the upper, steeply tilted part of the pluton (Faulds, 1999; Faulds and others, 2002).

The hanging wall of the Dupont Mountain fault consists of steeply to gently west-tilted early to middle Miocene volcanic and sedimentary strata. $^{40}Ar/^{39}Ar$ dating of variably tilted volcanic units within the hanging wall of the Dupont Mountain fault tightly brackets extension between ~16 and 11.3 Ma (Faulds and others, 1995). East and northeast of Copper Mountain, gently west-tilted 12.8 Ma basalt lavas cap the prominent buttes (Faulds, 1995). The Cottonwood basin to the east is a large composite basin of west-tilted half grabens situated in the hanging wall of the Dupont Mountain fault.

The Dupont Mountain fault represents the northern end of a major system of detachment faults that culminates with many tens of kilometers of displacement along the east flank of the Chemehuevi and Whipple mountains core complexes to the south. Here, it accommodates ~12–15 km of slip of correlative units in the footwall and hanging wall (Faulds and others, 2001a). Directly to the south, it links with the Newberry Mountains detachment fault. About 15 km to the north, however, it terminates in the Black Mountains accommodation zone.

RETRACE to Searchlight.

90.8 (10.5) Stop at Hwy 95. Watch for traffic and TURN LEFT (south). Proceed south on Hwy 95 toward Needles, CA. Searchlight is located on a topographic divide that separates the internally drained Eldorado Valley to the north from drainages into the Colorado River to the east and south. Piute Valley to the south drains to Needles before reaching the Colorado River.

95.1 (4.3) Continue past a power line road.

101.2 (6.1) Continue past the road east to Coast Guard Loran Station. Enter the community of Cal-Nev-Ari.

104.1 (2.9) PREPARE for a left turn across traffic! Look for a BLM kiosk on the left side of the highway.

104.3 (0.2) TURN LEFT at the BLM kiosk onto Christmas Tree Pass road. **STOP 1-2**. Discuss east/west extensional terrain differences. The Mid Hills block to the west is relatively unextended (Miller, 1996; Reynolds, 1996) compared to Cima Dome and the Halloran Hills. The Piute Range rises on the west side of Piute Valley and along the east margin of Lanfair Valley, and is bounded on both sides by Pliocene? normal faults (Nielson, 1996; Jennings, 1961). Erosional exposures on the east face of the range show dips of basalt flows fanning east, south, and north, punctuated with basalt slump blocks. Granitic gruess and arkosic sediments are exposed along the east base of the Piute Range.

Grand View Gorge on the west side of the Piute Range exposes a west-side-down fault, with basalt flows exposed at the bottom of the gorge. These may be related to the basalts to the east at the top of the narrow Piute Range, suggesting 250+ feet of uplift. Much of Lanfair Valley to the west has been covered by sediment derived from the east-tilting of the Mid Hills block (Katzenstein and others, 1996; Miller D. M., p.c. 1996; Neilson, 1996; Reynolds, 1996).

At the start of the trip, we passed sediments filling extensional basins that have been differentially tilted eastward. These include the Halloran Hills (45° east), Cima Dome (20° east) and the Mid Hills block (5° east). South-trending Piute Valley represents a structural boundary between east-tilting, westerly extended rocks to the west. In contrast, plutonic and volcanic rocks of the Colorado River extensional corridor in the Newberry Mountains (east) have been tilted westward as much as 90 degrees. PROCEED EAST on Christmas Tree Pass road.

106.3 (2.0) Continue past a power line road.

106.8 (0.5) Continue past a gas line and the Metropolitan Water District access road with a pole line, to east.

110.1 (3.3) Enter the wild, scenic granitic rocks of the Newberry Mountains.

110.5 (0.4) Continue past a reverse left turn.

110.6 (0.1) Continue past a right turn.

111.2 (0.6) Continue past a left turn.

111.4 (0.2) Continue past a left turn.

111.7 (0.3) Crossroads at summit. PULL RIGHT off the main road into the rough parking area at Christmas Tree Pass, elevation 3920'. **STOP 1-3.** . WALK to the end of the short, very rough road to the south and scramble onto the white granitic exposures.

We are on the west side, which is also the (original) top, of the west-tilted Spirit Mountain batholith (Volborth, 1973; Hopson et al,. 1994; Howard and others, 1994, 1996; Faulds and others, 1992; Haapala and others, 1996, 2005; Ramo and others, 1999; Claiborne and others, 2006; Walker and others, in press). This large (250 km^2) composite batholith is dominated by coarse-grained granite and quartz monzonite, but also contains of fine-grained granite and diorite and the miarolitic leucogranite that you see here. All of these lithologies are cut by a dike swarm (Newberry swarm) that is mostly felsic porphyry with some diabase and andesite.

U-Pb zircon geochronology using SHRIMP (*in situ* microanalysis) demonstrates a ~2 million year (17.4-15.3 Ma) history for the batholith (Walker and others, in press). Individual samples contain zircons with ages that span the lifetime of the batholith, suggesting recycling of extant zircon into new magma pulses: that is, *antecryst* entrainment. The batholith is envisioned as having been a patchwork of melt-rich, melt-poor, and entirely solid zones throughout its active life, with remobilization of crystal mush in response to magma replenishment. Only in the younger parts of the batholith are intrusive contacts well preserved.

The leucogranite exposed here forms a continuous cap for the entire 25 km north-south length of the exposed roof (Claiborne and others, 2006; Walker and others, in press). All of it is extremely felsic (76-78 wt% SiO_2). Rather than representing a single uniform zone emplaced at one time, however, it comprises innumerable sheets and dikes that accumulated over an interval of at least a million years. At this exposure, you will see that the very white rocks differ appreciably in texture (porphyry, aplite, medium-grained granite) and have fairly sharp contacts. These rocks appear to represent fractionated melts extracted from the underlying cumulate mushes over a protracted period of time. RETRACE to main road. PROCEED EAST.

111.9 (0.2) TURN RIGHT, proceed south.

112.5 (0.6) Slow for curves.

113.0 (0.5) Continue past upper Grapevine Canyon running southeasterly to Stop 1-4.

113.3 (0.3) Continue past the Lake Mead National Recreational Area (LMNRA) boundary sign.

113.4 (0.1) Pull right and PARK at the pull out.

STOP 1-4. Grapevine Lookout (elevation 3700 feet; the Spirit Mountains to the northwest rise to 5339 feet asl.) Looking to the west, we see a coarse, felsic granite (probably 73-74% SiO_2)—beneath the leucogranite—and relatively subtle features like rimmed (rapakivi) K-feldspar and schlieren. The flora in this area represents an island population of plant species common to both the Basin and Range and the Mojave Desert provinces. There is a mix of piñon, juniper, oak, mesquite, scrub oak, Nevin's barberry, nolina, Joshua tree, Mojave yucca, fleshy-fruited yucca, cottonwood, and willow (Fig. 1-2). The phainopepla (a crested, black bird with white wing bars) migrates from the coast to deserts to feed on, and spread seeds of, mistletoe that is parasitic on the mesquite, piñon and juniper. Return to vehicles, proceed down steep road.

Fig. 1-2. Christmas Tree Pass is host to uncommon plants including Nolina and Nevin's barberry as well as piñon, juniper mesquite, oak, Joshua tree, cottonwood, and desert willow (R.E. Reynolds).

114.1 (0.7) Continue past a second pull out.

114.4 (0.3) Continue past a right turn (gated) to Willow Springs.

114.6 (0.2) View southeast of Mohave Valley. Deep weathering of the Newberry Mountains in late Miocene time created granite spires and spheroidal granitic boulder piles (Oberlander, 1972) (Fig. 1-3). The Colorado River in the valley axis is about 500 feet above sea level, nearly 2000 feet lower than the axis of Paiute Valley. Note how insignificant the 10-story concrete structures at Laughlin appear when compared to the nearby granite spires. A critical question that we will consider is whether Mohave Valley is low because the Colorado River carved the valley, or the Colorado River initially occupied this valley because it was low relative to surrounding basins.

115.1 (0.5) View to the left (north-northwest) shows dark Proterozoic granite intruded by light-colored dikes. Ahead (west), the dark rocks are intruded by north-trending dikes.

116.0 (0.9) SLOW for intersection. Pipe Spring is 1.3 miles north (to the left). TURN RIGHT (south) toward Grapevine Canyon and Hwy 163.

116.6 (0.6) Continue past a pullout; note the black- and green-weathering dikes cutting granitic rocks to the left and right.

117.5 (0.9) Continue past a road on the right to Sacatone Canyon, with cottonwood and willow in thestream bottom. Proceed uphill past black-weathering, gray diabase dikes.

118.1 (0.6) Cross the divide into the Grapevine Canyon drainage.

118.2 (0.1) TURN RIGHT (west) to the Grapevine Canyon trailhead (with restrooms).

118.4 (0.2) PARK.

STOP 1-5. HIKE NORTHWEST from the trailhead for Grapevine Canyon. At the trailhead the "granite" (actually quartz monzonite) is quite dark. It's about as mafic as the granitic cumulates get in the batholith (Hopson and others, 1994). Sampling is prohibited here, but comparison with similar rocks elsewhere suggests that it has ~62-63 wt% SiO_2. Note the west-dipping foliation, defined in part by flattened mafic enclaves (these enclaves are ubiquitous except in more felsic granites toward the top, with some having entrained coarse feldspars from granite. Only at deeper levels to the southeast is there clear evidence for *in situ* mafic-felsic mingling).

Proceed up the trail. A large, extremely resistant, east-dipping (west-tilted), north-south dike supports the ridge through which narrow upper Grapevine emerges. This felsic porphyry dike is the most typical rock of the Newberry dike swarm. Elsewhere such dikes are ~15.3 Ma and have ~73 wt% SiO_2 (George and others, 2005). The host granite here is clearly more felsic than the quartz monzonite at the parking lot—true granite, ~71-72 wt% SiO_2 (again based on comparison to similar analyzed samples)—but still interpreted to be cumulate. This is one of the most characteristic rock types of the batholith.

The large dike has abundant, beautiful petroglyphs, part of the reason this canyon is sacred and protected. There is an interesting Holocene alluvial story in Grapevine Canyon. Petroglyphs support a model of aggradation (filling)

Fig. 1-3. Erosion of the Miocene granitic pluton has created scenic rock spires (R.E. Reynolds).

and degradation (cutting) in the canyon, which has one C-14 date of about 1500 yrs BP for the last major fill. The aggradation event buried a lower suite of petroglyphs and may have served as a "scaffold" for humans carving a later generation of petroglyphs. The mode of aggradation at this locality is a textbook example of a discontinuous ephemeral stream with aggrading fan-like features separated by deep arroyos. There is an intricate assemblage of steep headward cuts in the alluvial fill just downstream (east) of the Christmas Tree Pass road. Return to vehicles, RETRACE to "main" road.

118.5 (0.1) Stop, TURN RIGHT (south), and proceed to Hwy 163.

120.4 (1.9) Stop at Hwy 163 (Hiko Springs is to the right). Watch for fast downhill traffic.

TURN LEFT (east) toward Laughlin, NV.

122.5 (2.1) View northeast of Davis Dam and the Pyramid hills. Upstream from the dam is an outcrop referred to as the "Bull Head" (now partially submerged), from which Bullhead City gained its name. The dam was constructed across Pyramid Canyon in the Pyramid hills. These hills consist of distinctive, dark Proterozoic megacrystic granite. These hills were the source of the gravel of Pyramid hills (informally named "Pyramid gravel"; House and others, 2005), which occurs only downstream from the Pyramid hills in northern Mohave Valley. The Pyramid gravel is very poorly sorted and includes clasts of Proterozoic granite and light-colored Miocene granite from the Newberry Mountains. Deposits of Pyramid gravel are conformably overlain by the Bouse Formation. The Pyramid gravels were deposited by a Miocene flood when ponded waters in Cottonwood Valley breached the Pyramid hills. This flooding resulted in the deposition of coarse gravel downstream in Mohave Valley. As Mohave Valley filled with quiet water, Bouse limestone was deposited over the Pyramid gravels, followed by fine-clastic deposition. Bouse limestone outcrops in Cottonwood Valley range from 1160 to 1800 feet asl. In Mohave Valley, outcrop elevations of Bouse limestone range from 500 feet to 1800 feet. Based on these relationships, we infer that the floor of Cottonwood Valley prior to development of the Colorado River was at least 600 feet higher than the floor of Mohave Valley. This difference in elevation facilitated the downstream spilling of incipient Colorado River water from Cottonwood to Mohave Valley.

123.0 (0.5) TURN RIGHT (south) on Needles/River Road, CA.

125.3 (2.3) Continue past Bruce Woodbury Drive.

126.1 (0.8) Continue past the light at El Mirage Way.

127.4 (1.3) Continue past East Casino Drive and proceed straight ahead (south).

127.9 (0.5) Drop into the pre-dam Colorado River flood plain.

128.1 (0.2) **STOP 1.6**. The Newberry detachment fault is visible to the west at 2:00. PARK and WALK to the smooth detachment surface. The upper plate consists of red megacrystic mid-Proterozoic granite of Davis Dam. The gray granodiorite in the lower plate is the middle Miocene granite of Mirage (House and others, 2004).

129.7 (1.6) Proceed south and ascend bluffs.

130.4 (0.7) TURN RIGHT (west) onto the gas line road.

131.2 (0.8) Continue past the first pole line.

131.4 (0.2) TURN RIGHT (north) on the pole line road.

131.8 (0.4) **STOP 1-7**, deep gully. Examine the basal white marl of the Bouse Formation at elevation 800 feet interbedded with fanglomerate derived from the Newberry Mountains to the west at elevations 920–960 feet asl (House and others, 2004). The thinly bedded Bouse marl is up to 1 m thick and apparently was draped over a gently undulating surface formed on relatively fine, oxidized fanglomerate. Bouse deposits are overlain by gray Newberry fanglomerate.(Fig. 1-4). RETRACE to gas line.

132.1 (0.3) TURN RIGHT (west) at the gas line road.

133.5 (1.4) Cross the buried gas line.

134.1 (0.6) **STOP 1-8**. At the intersection, WALK NORTHWEST to outcrops of Bouse marl cementing granitic clasts. This represents a slope transgression up the pediment by the Bouse Formation from 760 feet (Stop 1.6) to 1560 feet at this location. To the southwest (stop 1.9), basal Bouse limestone reaches 1720 feet asl (House and others, 2004). This is one of several local Bouse outcrops that suggest that the late Miocene paleotopography of Mohave Valley was similar to that of the modern valley (House and others, 2005). When Mohave Valley was filled by Bouse waters to ~1800 feet, the downstream paleodivide might have been the Mohave Mountains south of Topock and east of the Chemehuevi Mountains, an area that has been deeply eroded by the Colorado River (Stop 2-13). PROCEED LEFT (southwest) at the line of double poles.

135.1 (1.0) TURN LEFT (southeast) on a dirt track.

135.5 (0.4) **STOP 1-9.** PARK before reaching the granitic (quartz monzonite) outcrops. WALK NORTH to observe basal limestone (micritic limestone, Poulson and John, 2003) of the Bouse Formation with granitic clasts deposited on the limonite-stained erosional surface. The rounded, locally derived clasts here (elevation 1600 feet) have been interpreted as beach gravels (House and others, 2004, 2005). RETRACE to gas line road.

135.9 (0.4) TURN RIGHT at the double pole road.

136.9 (1.0) TURN RIGHT (east) on the gas line road.

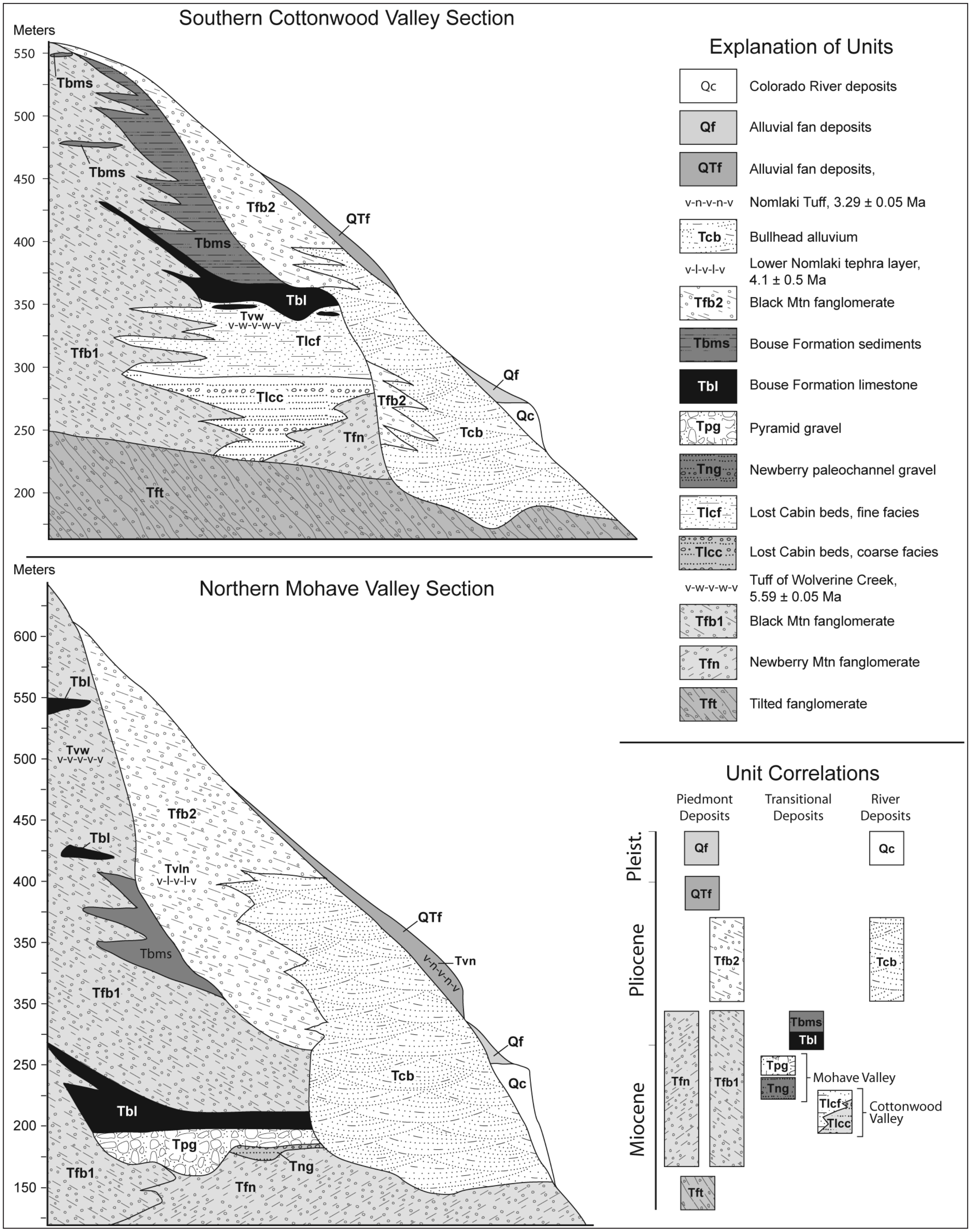


Fig. 1-4. Stratigraphic sections, Cottonwood and Mohave valleys (P.K. House).

138.9 (2.0) Continue past the pole line road.

139.1 (0.2) **STOP 1.10**. Pull to the right near a low terrace. Examine rounded cobbles of Pliocene Colorado River deposits (alluvium of Bullhead City; House et al, 2005; informally called "Bullhead alluvium") at 840´ elevation asl. The rounded clasts are lithologically diverse, including quartzite, siliceous Paleozoic limestone, dark Paleozoic chert, minor granite (quartz monzonite), and fractured Miocene porphyritic volcanics. PROCEED EAST to Needles Highway.

139.9 (0.8) Stop, TURN LEFT (north) onto Needles Hwy.

142.6 (2.7) TURN RIGHT (east) on Casino Drive.

143.7 (1.1) This route passes through a series of exposures of various Quaternary Colorado River deposits (look for rounded gravels) and related piedmont tributary units as it climbs from the pre-dam floodplain to Pleistocene terrace remnants (elevation 540´).

145.9 (2.2) Top of bluff.

146.3 (0.4) **STOP 1-11**. Panda Gulch. WALK EAST to a deep gully on the east side of Casino Drive. The purpose of this stop is to observe an exposure of the stratigraphic evidence of the local transition from closed/restricted basin, to lake, to the through-flowing Colorado River in northern Mohave Valley. Each member of the sequence is clearly visible on the northwest-facing slope above the bed of the wash. Late Miocene, untilted Newberry fanglomerates form the base of the section and are overlain by a cross-stratified channel gravel deposit containing subangular clasts of local rock types—the "axial gravel" (Tag). Paleochannels are generally oriented north-south. We interpret this gravel as marking the position of an axial channel near the head of Mohave Valley at a time when it was not connected with drainage through Cottonwood Valley.

The axial gravel and underlying fanglomerate are unconformably overlain by the Pyramid gravel (Tpg), a conspicuous, immature conglomerate. Deposition of the Pyramid gravel was accompanied by erosive enlargement of the axial channels and locally deep scour into the underlying fanglomerate. The Pyramid gravel is predominantly a cobble-boulder conglomerate with a maximum thickness of about 20 m. Many clasts are sub-rounded to rounded. Some exposures contain sparse, light-colored granitic gravel and sand (reworked Newberry fanglomerate) and others are dominated by thick, stratified sequences of monomictic grussy sand and subrounded pebble-gravel possibly derived from regolith stripped from slopes of the Pyramid hills. Cross-stratification is evident in some intervals of the Tpg (e.g., trough cross-bedding). Many exposures show clast-supported structure, whereas some are matrix-supported and slurry-like.

Most outcrops of the Pyramid gravel are composed almost entirely of cobbles and boulders of megacrystic granite from the Pyramid hills and reworked local fanglomerate from the Newberry Mountains. House and others (2005) interpreted the Pyramid unit as a catastrophic flood deposit from a clear-water breach through a paleodivide in the Pyramid hills. The lowest Bouse deposits found in the axis of Cottonwood Valley are at about 1160 ft asl, about 500 ft higher than the Pyramid gravel deposits. These relationships are consistent with filling of a perched lake in Cottonwood Valley prior to spill-over through the Pyramid hills.

The Pyramid gravel is conformably overlain by the Bouse Formation, indicating inundation of this area by standing water following the deposition of the gravel. Exposures of Bouse deposits in the Laughlin bluffs include beds of marl, mud, and minor sand and range from 0.5 m to approximately 3 m thick. The Bouse outcrops in the Laughlin bluffs occur only 140 feet above the surface of the modern Colorado River. The sequence of Bouse deposits over Pyramid gravel suggests that a catastrophic flood from an upstream source, likely a lake, was followed by quiescent deposition in a large body of standing water. As noted earlier, presence of a deep lake in this valley would require a paleodivide at the southern end of Mohave Valley near Topock (Stop 2-13).

The preceding sequence is unconformably overlain and deeply incised by gravels and sands of the early Colorado River—the Bullhead alluvium (Tbh). The Bullhead alluvium is generally equivalent to units A and B of Metzger and others (1973).The gravel at the base of the Bullhead unit is largely composed of locally derived sand and gravel reworked from Newberry fanglomerates. It is also peppered with well-rounded pebbles of chert and well-rounded cobbles of diverse lithologies from far upstream. The mix of light-colored local sand and gravel with the exotic, mostly dark-colored pebble component imparts a distinctive black and white speckled appearance to the unit, hence our informal name "Panda gravel" (Tbhl).

The Panda gravel has a cobble-rich basal conglomerate containing many clasts of reworked fanglomerate and Pyramid gravel. Look on the slopes for examples of fluted (stream-worn) boulders. The base of the Panda unit dives down-section into the late Miocene fanglomerate down the wash and then climbs up toward the east and the modern course of the Colorado River. The base of the Panda gravel here thus defines a large paleochannel. Evidently, the arrival of the river in northern Mohave Valley immediately preceded or was accompanied by an interval of erosion following the recession of the Bouse water body. The step-like geometry in the Panda gravel channel here, the occurrence of stacks of laterally continuous layers of coarse gravel, and the abundance of locally derived clasts in the body of the deposit suggests lateral erosion and reworking of the Miocene substrate by a rapidly aggrading, greatly over-fit river. The Bullhead unit is discontinuously exposed from this point (~20m above the river) up to gravel lags and


Fig. 1-5. Road cut exposes severely folded Proterozoic granitic gneiss and cross-cutting dikes (R.E. Reynolds).

river sands interfingered with Black Mountain alluvial fan gravels at levels as high as 760 feet above the river at 1,300 feet asl. PROCEED NORTH on Casino Drive.

146.7 (0.4) Continue past Harrah's on the right.

146.8 (0.1)TURN LEFT (northwest) at Thomas Edison.

147.8 (1.0) Continue past the signal at Bruce Woodbury.

148.4 (0.6) Stop sign at Civic Drive.

148.7 (0.3) Stop at Highway 163. TURN right (east) toward Arizona.

149.2 (0.5) Stop light at Casino Drive. Move left and prepare for a left turn on AZ Hwy 68.

149.5 (0.3) Stop at the junction of Bullhead Parkway and AZ Hwys 163 & 68 on the east side of the river. TURN LEFT (north) on Highway 68, which bears east.

150.3 (0.8) Bear right at McCormick Blvd.

151.0 (0.7) Continue past a left (north) turn toward Davis Dam. Drive through a section of Colorado River gravels with rounded cobbles.

156.3 (5.3) Continue past Katherine Mine Road to the left.

156.8 (0.5) Enter Proterozoic megacrystic granite bedrock.

158.2 (1.4) Enter the Black Mountains, with red-brown early to middle Miocene volcanic and intrusive rocks over gray granitic rocks.

159.3 (1.1) Curve to the left. Note the dark and light volcanic flows on the ridge to the east over dark gray granite.

161.5 (2.2) Union Pass, elevation 3625´. Proceed east into Golden Valley toward Kingman and Hwy 93.

167.4 (5.9) Continue past Estrella Street.

169.3 (1.9) Continue past Roosevelt Road—two gas stations are ahead.

176.0 (6.7) Prepare to enter Hwy 93 northbound toward Hoover Dam and Las Vegas. Stay in the left lane.

177.0 (1.0) ENTER AZ Hwy 93. The onramp loops 270° south, then north. Watch for merging traffic.

178.9 (1.9) The road cut on the right exposes severely folded Proterozoic granitic gneiss and cross-cutting dikes (Fig. 1-5).

179.1 (0.2) Sundown Drive (Agua Fria Drive is to the west). Mineral outlets at this intersection offer a good selection of turquoise from local mines. These mines are located to the northeast on Turquoise Mountain and on Ithaca Peak in the Cerbat Mountains (Anthony and others, 1995).

182.2 (3.1) Historical marker for the town and mining district of Cerbat.

183.0 (0.8) The Duval mine dumps are to the northeast at 2:00.

185.5 (2.5) Mineral Park Road.

191.3 (5.8) The town of Chloride is to the right on Mohave County Road 125. Mining started in1864, producing gold,


Fig. 1-6. Cottonwood Summit. View of west-tilted Mt. Perkins fault block, which exposes rotated crust and a large stratovolcano with a superimposed rhyolite dome complex. The Mt. Perkins stratovolcano is probably the eastern half of the Searchlight stratovolcano, displaced eastward on the Dupont Mountain detachment across the Colorado River to Stop 1-1 (R.E. Reynolds).

copper, and silver. A tour of Chloride and its active reuse of historic buildings is well worth your time. The valley to the west marks the drainage divide between Detrital Valley draining north to Lake Mead, and Sacramento Valley draining south to Topock (Stop 2.15).

192.9 (1.6) Big Wash Road.

198.3 (5.4) Move to the left lane for a left turn across traffic. You must stop in the median.

198.5 (0.2) TURN LEFT onto Cottonwood Road.

201.5 (3.0) Cross Detrital Wash. The road bends west-northwest.

204.5 (3.0) Cross a cattle guard.

205.5 (1.0) Continue past Jones Road on the left.

205.9 (0.4) Continue past the southwest diagonal road to Lost Cabin Spring.

207.6 (1.7) Slow for a cattle guard at the road south to Squaw Pocket Well. TURN RIGHT in 100 feet.

207.7 (0.1) **STOP 1-12**. Cottonwood Summit. The crest of the range provides a view of the large west-tilted Mt. Perkins fault block, which exposes a 9 km thick section of crust on end, including a large 18–16 Ma stratovolcano and superimposed 16–15 Ma rhyolite dome complex, all exposed in cross-section (Faulds and others, 1995). The Mt. Perkins stratovolcano is probably the eastern half of the Searchlight stratovolcano but has been displaced eastward on the Dupont Mountain detachment. View across the Colorado River to related faults at Stop 1-1 (Fig. 1-6).

SUMMARY. Day One stops have visited wild and scenic terrain created during the Miocene along the Colorado River extensional corridor. The resultant trough was filled by fluvial deposits, followed by spillover creating Bouse lakes with distinctive white basal limestone. The Bouse Formation is overlain by Colorado River deposits including the basal alluvium of Bullhead City.

207.9 (0.2) RETRACE to the cattle guard and PROCEED EAST on Cottonwood Road.

209.5 (1.6) TURN OBLIQUE RIGHT (west-southwest) on the road to Lost Cabin Spring.

211.1 (1.6) Continue through crossroad to Squaw Pocket Wells,

212–212.8 (1.0-1.7) Juniper Flats.

End of Day One

Day Two

This trip speeds rapidly along the east side of the Colorado River through wild and rugged tectonic and fluvial topography (Stops 2.1, 2.6). The extreme differences in elevation caused valleys to be partially filled by the Lost Cabin beds, the Bouse Formation sediments, Bullhead alluvium and the Black Mountain fanglomerate (stops 2.2 – 2.5), and subsequently incised by the Colorado River drainage system (Stops 2.5, 2.7, 2.8). The Pleistocene Colorado River has left Chemehuevi overbank deposits, the Chemehuevi Formation of Longwell (1936). Recent dams provide scenic back-bay wildlife habitat.

Look for the difference between fluvial deposits derived from ranges lateral to the Colorado River trough that contain angular clasts and deposits from the axial drainage of the Colorado River with rounded clasts of far-traveled lithologies. In general, the source of gravel on the east side of the Colorado River is the volcanic terrain of the Black Mountains. The ranges on the west side of the Colorado River are predominantly granitic and metamorphic rocks. When we get to Chemehuevi Wash (Day 3) the source is the Turtle Mountains volcanics in the west.

In the afternoon of Day 2 we will visit Bouse Formation outcrops where time of deposition is constrained by dated tuffs. We will also inspect Colorado River gravels and outcrops of the Chemehuevi Formation that have been interpreted as lacustrine and overbank deposits.

0.0 (0.0) Juniper Flats. Proceed southwest down Lost Cabin Wash. High clearance and FWD vehicles are recommended.

0.1 (0.1) Lost Cabin Springs. Proceed south, down-canyon.

0.2 (0.1) Miocene volcanics to the east contain white ash and breccia.

0.8 (0.6) A sharp left "S" curve in the canyon. The road

Fig. 2-1 The weight of dark volcanic breccias has compressed (load cast) the upper surface of the white ash (R.E. Reynolds).

Fig. 2-2. White, thin beds of basal Bouse limestone (R.E. Reynolds).

bears southward past two water tanks.

1.1 (0.3) Look west at sediments with dark volcanic boulders. These are typical of dark sediments derived from the Black Mountains.

1.4 (0.3) View east of dark volcanic flows and breccias. The upper surface of the white ash has been compressed by load casting (Fig. 2-1).

2.2 (0.8) The road in Lost Cabin Wash passes through upper narrows.

3.1 (0.9) Continue past an aluminum water tank. The road bears to the west.

3.4 (0.3) Continue past a right turn.

3.8 (0.4) **STOP 2-1**. A major east-dipping, low-angle normal fault (the Mockingbird Mine fault) that bounds the Mt. Perkins block on the east cuts through the canyon in this area, juxtaposing upper parts of the Miocene section to the east against lower parts of the section to the west. A few km to the north in the footwall of the fault, rhyolite domes are tilted 90° to the west. Displacement across the Mockingbird Mine fault decreases to the south, but the fault probably continues southward to Union Pass.

4.0 (0.2) Middle narrows. Continue past a road to the left.

5.5 (1.5) Enter the lower narrows.

5.8 (0.3) Junction. BEAR RIGHT (west) on the road to Jeep Cove. We will proceed 5.4 miles. The steep road to the south leads to the Portland mine.

6.0 (0.2) Exit the lower narrows and view Cottonwood Valley and the waters of Lake Mohave.

6.8 (0.8) Leave volcanic terrain and enter canyon walls composed of moderately indurated west-dipping fanglomerate.

9.2 (2.4) Continue past white limestone of the Bouse Formation. We will discuss these on our return.

11.6 (2.4) **STOP 2-2.** Ahead are the sparkling blue waters of Lake Mohave at Jeep Cove. Spirit Mountain and the Newberry Mountains (Stops 1-1, 1-3, 1-4) can be seen to the southwest. At this stop, examine the fluvial gravels in the canyon walls. They are light colored because they are composed of pale granite, quartz monzonite, and gray diabase. This is the same suite of rocks that we saw in the Christmas Tree Pass area 3 km west (Stop 1-3). When these gravels were deposited, there was no through-flowing Colorado River, and the fans from the Newberry Mountains extended 10 miles across the trough to this area. Walk southwest up-canyon. The brown and greenish gray silts inset into the white fanglomerates are Quaternary Colorado River sediments, probably the Chemehuevi beds/formation. RETRACE east up Lost Cabin Wash.

11.9 (0.3) **STOP 2-3**. Dark, volcanic clasts from the Black Mountains mix with and begin to dominate the granitic clasts in the fanglomerate. This mixing of source lithologies suggests that this was approximately the valley axis when these sediments were deposited. We infer that the axial wash(es) drained to the north at that time (House and others, 2005).

12.6 (0.7) Enter cliffs exposing a section of Lost Cabin beds

Fig. 2-3. A thin bed of Bouse limestone on north canyon wall contains impressions of water reeds and charophytes, both requiring fresh water (R.E. Reynolds).

that are distinct from the light colored Newberry Mountains fanglomerate and the dark-colored Black Mountain fanglomerate.

13.2 (0.6) **STOP 2-4**. The Lost Cabin beds (House and others 2005) are tan to pinkish, flat bedded silts with gravel stringers supported by the silt matrix. The existence of minor unconformities and weakly to moderately developed buried soils in the Lost Cabin beds implies that they represent subaerial deposition, possibly on the southern margin of the primary basin depocenter. They are disconformably overlain by Black Mountain conglomerates with a coarse volcaniclastic groundmass. In several exposures north of this site, the 5.5 Ma tuff of Wolverine Creek lies below a paleosol in the upper third of the Lost Cabin beds (House and others, this volume).

13.7 (0.5) Continue past white, thin beds of Bouse limestone, one of the three facies of the Bouse Formation. (Fig. 2-2). The white Bouse marl (limestone) is generally conformable on the Lost Cabin beds, although there are local unconformities between Bouse and Lost Cabin beds. One example in Lost Cabin Wash is an erosional channel in the Lost Cabin beds that is filled with green-gray mud. Elsewhere in this area, Bouse deposits (reworked in some cases) unconformably overlie the Lost Cabin beds (House and others, this volume). The lowest Bouse outcrops in Cottonwood Valley rest on the Lost Cabin beds at about 1160 feet asl, substantially higher than the Bouse outcrop at Stop 1.7 in Mohave Valley. Bouse and Lost Cabin beds are disconformably overlain by the Black Mountain fanglomerate, and in many outcrops Bouse beds have been cut out by erosion.

14.5 (0.8) **STOP 2-5**. A thin bed of Bouse Formation limestone is exposed on north canyon walls (Fig. 2-3). Below the Bouse Formation is deeply weathered red-brown sandy gravel with limonite cemented clasts. The association of Bouse limestone over oxidized sediment is very common. We saw similar red oxidized gruss at Stop 1.9. The thin fossiliferous layer of calcareous sand contains impressions of water reeds and charophytes, the latter requiring fresh water. The fossiliferous layer is overlain by the basal white limestone of the Bouse Formation, overlain by brown and greenish silts typical of the clastic interbedded basin-filling member of the Bouse Formation (Metzger and Loeltz, 1973). These facies suggest a fresh water, transgressive, onlapping lacustrine sequence. Proceed up Lost Cabin Wash.

16.6 (2.1) Volcanic outcrops begin.

17.4 (0.8) Bear left at the junction with the road south to the Portland mine.

18.2 (0.8) Continue past a tank.

19.2 (1.0) Upper narrows.

22.2 (3.0) Continue past Lost Cabin Spring.

24.1 (1.9) Continue through cross roads.

25.5 (1.5) TURN RIGHT onto Cottonwood Road.

27.0 (1.4) Cattle guard. Cottonwood Road bends east-southeast.

30.2 (3.2) Cottonwood Road bears east at Detrital Wash.

33.2 (3.0) Stop at AZ Hwy 93. TURN RIGHT (south).

40.4 (7.2) Continue past Chloride, to the east on Hwy 125.

53.5 (13.1) Prepare to exit Hwy 93.

53.9 (0.4) EXIT Hwy 93 onto Hwy 68

54.3 (0.4) BEAR RIGHT and PROCEED WEST on Hwy 68.

Fig. 2-4. Union Pass. The west-dipping Frisco Mine fault, exposed at the Frisco Mine, cuts the east-dipping Union Pass fault (R.E. Reynolds).

63.1 (8.8) Cross Sacramento Wash, which we will visit late in the day at Topock.

63.5 (0.4) Estrella Street.

69.4 (5.9) Union Pass.

71.9 (2.5) Sugarloaf Mountain.

72.7 (0.8) Prepare to turn right.

72.9 (0.2) TURN RIGHT (north) on old Route 66.

76.3 (3.4) Slow past an outcrop of ~1.4 Ga “megacrystic” granite with large (2´´) K-feldspar phenocrysts.

76.7 (0.4) Continue past the rock yard on the left at the Frisco mine.

77.3 (0.6) Wild and scenic Route 66 curves right, then left.

77.5 (0.2) **STOP 2-6**. Union Pass and Route 66. Stop at the berm blocking old Route 66. The ridge to the east is largely composed of ~6-17.4 Ma rhyolite flows and intrusions, punctuated by several rhyolite domes that form the large buttes. The east-dipping Union Pass fault juxtaposes the rhyolites to the east against Proterozoic granite and gneiss exposed in the more subdued lower terrain. A large rhyolite dike is intruded along the Union Pass fault. The west-dipping Frisco Mine fault, exposed just to our west at the Frisco mine (Fig. 2-4), cuts the east-dipping Union Pass fault. It therefore appears that an early east-dipping normal fault system was sealed by rhyolite dikes and subsequently cut by a later west-dipping fault system (Murphy, 2004; Murphy and others, 2004). The rhyolites in this area may be genetically related to the Spirit Mountain or Mirage plutons in the Newberry Mountains and were displaced eastward relative to their plutonic roots by displacement on the Newberry Mountains detachment fault. RETRACE to Hwy 68.

79.7 (2.2) Stop at Hwy 68. Wash for traffic and TURN RIGHT (west) toward Laughlin.

81.3 (1.6) Katherine Mine Road.

83.3 (2.0) View west and south of Mohave Valley (Cottonwood Valley is northwest, out of sight).

86.5 (3.2) Continue past Davis Dam Road.

87.3 (0.8) Stop light at McCormick Blvd. Slow for steep downhill.

87.7 (0.4) The Colorado River museum is on the right. Prepare to stop at Bullhead Parkway and AZ Hwy 163.

88.0 (0.3) Stop light. TURN LEFT on Bullhead Parkway and proceed east.

89.1 (1.1) Continue past a right turn to the airport.

90.3 (1.2) Continue past a light at Desert Foothills Parkway.

91.1 (0.8) Continue past a light at Laughlin Ranch.

The timing of deposition of the Bouse Formation north of Topock Gorge is constrained between two dated tuffs (see Fig. 1-4: Tvln=Nomlaki Tuff; Tvw=Tuff of Wolverine Creek). They appear in stratigraphic superposition (see Table 1).

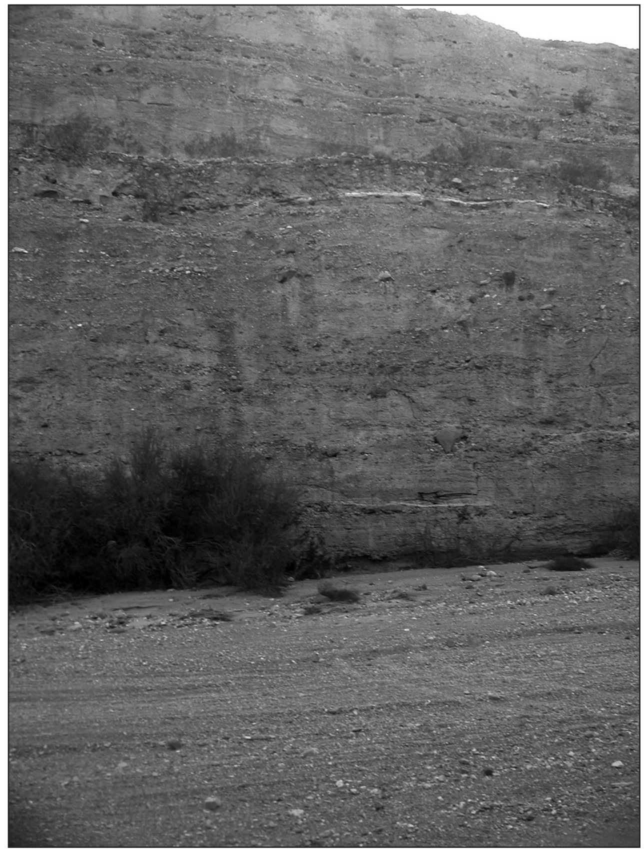


Fig. 2-5. Secret Pass Canyon. The tuff of Wolverine Creek (5.59 Ma) is exposed about 2 meters above the canyon floor below the basal Bouse limestone. The basal Bouse limestone (Tb) sits at the top of Black Mountain fanglomerate Unit Tfb1, and is overlain by Tfb2, a second pulse or unit of fanglomerate from the Black Mountains. Dated tephra deposits indicate that the Bouse formation was deposited between 5.5–3.3 Ma (R.E. Reynolds).

91.3 (0.2) Continue past the contractor yard at the entrance to Secret Pass Canyon.

93.0 (1.7) Stop at the light for Silver Creek Road, the east turn to Oatman. TURN LEFT (east) and proceed on the rough, graded dirt road.

94.1 (1.1) Continue past a road north into Silver Creek wash.

94.5 (0.4) Continue past a second road north into Silver Creek wash.

94.7 (0.2) **STOP 2-7**. Nomlaki Tuff at Silver Creek. Climb down-canyon and look south at exposures of the Nomlaki Tuff (Ttn; 3.3 Ma) in tributary fan gravel over an erosion surface cut across the older Colorado River deposits of the alluvium of Bullhead City (Tbh). The Bouse Formation (Tb) is not exposed at this locality, but elsewhere sits stratigraphically below the Bullhead alluvium. View west across the river of the detach-

Table 1. Stratigraphic position of dated tuffs.

Location	Unit	Feature	Age/Age Range
Boundary Cone powerline	Black Mtn fanglomerate	Lower Nomlaki tephra layer	~4.1 Ma
Mohave Valley	Bouse Formation	Basal limestone	5 – 4.1 Ma
Secret Pass Canyon	Lost Cabin Beds	Tuff of Wolverine Creek	5.59 Ma

ment, lower plate, and gas line stops from Day 1.RETRACE to Bullhead Parkway.

96.5 (1.8) Stoplight at Bullhead Parkway. TURN RIGHT (north) and stay in the right lane.

97.7 (1.2) Prepare to turn right at the contractor yard.

98.2 (0.5) TURN RIGHT (east) through the contractor yard (FWD is strongly recommended!) and proceed east upstream into Secret Pass Canyon.

100.5 (2.3) The canyon widens into two branches: STAY LEFT (northeast). Continue past two roads ascending terraces to the south.

101.0 (0.5) Clastic Bouse deposits are exposed on both sides of Secret Pass Canyon, but are most obvious on the north bank. Large blocks of indurated Black Mountain fanglomerate have slumped down onto the much finer-grained Bouse deposits.

101.5 (0.5) **STOP 2-8**. Secret Pass Canyon. The tuff of Wolverine Creek (5.59 Ma) is exposed about 2 meters above the canyon floor (Fig. 2-5). This thin white layer of tephra is overlain by a thick sequence of fanglomerate. The basal Bouse limestone (Tb, Fig. 2-5) sits at the top of Black Mountain fanglomerate Unit Tfb1, and is overlain by Tfb2, a second pulse or unit of fanglomerate from the Black Mountains. Therefore, exposures in northern Mohave Valley containing dated tephra deposits indicate that the Bouse Formation was deposited between 5.5–3.3 Ma.

At this site, the Bouse Formation and an underlying tephra bed are exposed in a thick package of tributary fanglomerate. A fairly continuous, up to 0.5-m-thick tephra bed exposed about 2 meters above the canyon floor has been identified via geochemical analysis and correlation as the 5.59 Ma tuff of Wolverine Creek (House and others, this volume). This tephra bed is overlain by a thick sequence of fanglomerate. The very thin basal Bouse limestone (Tb) is intercalated in Black Mountain fanglomerate deposits about 10 m above the tephra bed. In some exposures upstream, the Bouse Formation consists primarily of massive or cross-bedded sand deposits over a very thin limestone layer, but in contrast to a few hundred meters to the west, the total thickness of the deposit is 3 m or less. The Bouse deposits here seem to represent a brief incursion of quiet water into an alluvial fan environment, but erosion prior to renewed tributary fan deposition likely removed some unknown amount of overlying Bouse deposits. The Bouse limestone in Secret Pass Canyon of Mohave Valley is at 1400–1450 ft asl; other Bouse exposures in this area range up to 1800 ft asl. In this same general area, Colorado River gravel and sand of the alluvium of Bullhead City is found up to 1320 ft asl. At two locations in this area, the 4 Ma Lower Nomlaki tephra is exposed at elevations of 1200–1300 ft asl in association with high Colorado River deposits that post-date the Bouse deposits. Thus, filling of the valley with deep water and deposition of the Bouse deposits in this area occurred after 5.6 Ma, and the major phase of Colorado River aggradation culiminated about 4 Ma. RETRACE to Bullhead Parkway.

104.8 (3.3) Continue through the contractor yard. Stop at paved Bullhead Parkway, watch for oncoming traffic, TURN LEFT (south) and proceed toward AZ Hwy 95 toward Golden Shores.

106.5 (1.7) Continue past the Silver Creek Road stoplight.

107.8 (1.3) Bullhead Parkway bears right.

113.9 (6.1) Continue past Aztec and El Rodeo roads.

117.2 (3.3) Continue past Boundary Cone Road.

120.2 (3.0) Continue past Willow Road.

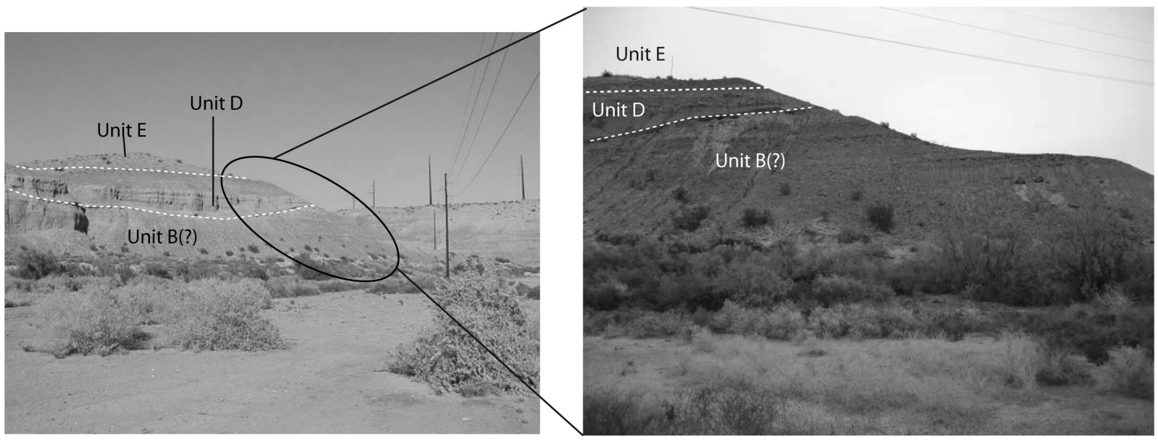
124.1 (3.9) TURN LEFT (east) on Courtwright Road, Highway 227

126.0 (1.9) Continue past Vanderslice Road.

127.5 (1.5) Continue past View Lane on the right. Calpine

Fig. 2-6. The Chemehuevi Formation near Stop 2-9 (D.V. Malmon).

Panoramic photograph of powerplant outcrop, looking south. Dotted white lines delineate apparent erosional unconformities.



Panoramic photograph of nearby outcrop, looking southeast.

Fig. 2-7. The Bouse Formation near Stop 2-9 (D.V. Malmon).

Southpoint Power Plant is on the left.

127.7 (0.2) Courtwright Road bears right (south). PROCEED EAST on the dirt road.

127.9 (0.2) **STOP 2-9**. Southpoint Power Plant. PARK east of the Southpoint Power Plant. The Quaternary Colorado River deposits here contain multiple erosional unconformities, illustrating some unresolved issues about the evolution of the Quaternary Colorado River. In the outcrop that faces the floodplain, flat-lying beds that appear to be the lower, mud subunit (Unit D; Metzger and others, 1973) of the late Pleistocene Chemehuevi Formation of Longwell (1936) overlie similar-looking beds above a clear erosional unconformity. This erosional surface, which can be traced for several kilometers to the north and south, contains tens of meters of relief. In a wash a mile southeast, this erosional surface is overlain by a bed of reworked tephra that has been geochemically correlated to two ash beds found farther upstream in Unit D-like beds (D. Wahl, USGS Tephrochronology Laboratory, p.c. 2007). These ash beds have been assigned to the late Pleistocene, though their precise age remains uncertain (see Malmon and Howard, this volume). At the nearby tephra locality, the beds underlying the erosional surface are similar to the Bullhead alluvium seen earlier on this trip, suggesting that this particular unconformity represents the erosional interval between the "Bullhead" deposits and the "Chemehuevi" deposits. If this is true, then the similarity of the beds above and below the erosional surface in this outcrop demonstrates that the two units can look very similar, and that the river during "Bullhead" time contained at least some floodplain lake and overbank environments.

An interbedded sand-dominated sequence is visible in the north-facing outcrop around the corner (the mesa top with the electrical lines leading east) from the power plant. This outcrop contains more than one erosional unconformity (Figs. 2-6, 2-7), overlain by sands of Unit E (Metzger and others, 1973), the upper sand subunit of the Chemehuevi Formation. At the northern end of Mohave Valley, erosional unconformities are in the fine-grained late Pleistocene beds (House et al, 2005). Major unconformities in deposits commonly assigned to the Chemehuevi Formation may represent erosion between different fill episodes. Alternatively, they may represent minor erosional intervals within a generally aggrading river system. Additional geochronologic constraints are required to understand this stratigraphy.

Recent results from optically stimulated luminescence (OSL) samples collected from the Chemehuevi Formation suggest that a major late-Pleistocene aggradational pulse occurred sometime between 60–50 Kyr ago, following the transition from OIS 4 to OIS 3, with downcutting occurring about 45–30 Kyr ago. The data indicate that most of the fine-grained "Chemehuevi"-like outcrops that remain in the valley are from this alluviation episode, and that similar conditions did not occur again until Holocene time.

RETRACE to pavement.

128.1 (0.2) Stop at the pavement. TURN LEFT (south) onto Highway 1.

129.5 (1.4) Continue past a right turn to the Pintail Slough north dike.

132.0 (2.5) Prepare to turn left.

132.2(0.2) Past a sandy wash, TURN LEFT up an obscure jeep road and proceed 0.2 mi to buried pipelines.

132.4 (0.2) **STOP 2-10.** WALK up the wash to examine well-exposed stratigraphy. High banks on the right expose >20 m of cross-bedded fluvial sandstone and interbeds of fluvial roundstone conglomerate (unit B of Metzger and Loeltz, 1973; the Pliocene-age alluvium of Bullhead City, House et al, 2005; units A and B of Metzger and others, 1973).

Look for well-rounded pebbles of chert, quartzite, and Paleozoic limestone derived from far upstream. Locally derived angular or subangular clasts of volcanic rocks and gneiss make up at least half of the larger clasts. Well-rounded quartz sand grains in the sandstone may be reworked from Permian and Jurassic sandstones on the Colorado Plateau. Rusty zones in the deposits along this wash are commonly associated with clay balls, wood casts, or rare bones of fossil vertebrates. Clay balls may derive from bank erosion of rare clay beds in the sequence, or from the Bouse Formation. Note the planar crossbeds in both sandstone and conglomerate on both banks of the wash.

Note the planar crossbeds in both sandstone and conglomerate on both banks of the wash. Cross-bedded fluvial sandstone is also exposed in the bed of the modern wash. This indicates that the modern flat-floored drainage is graded to Topock Marsh on an erosion surface, and by inference this particular wash postdates the 20–30 m of Holocene aggradation known for the nearby Colorado River floodplain downstream (House et al, 2005).

Capping the alluvium of Bullhead City is a paleosol overlain by dark, coarse-grained alluvial fan deposits (fanglomerates) dominated by basalt derived from the Black Mountains. Metzger and Loeltz (1973) described a series of such fanglomerates (their unit C, piedmont gravels) as alluvial fans that prograded into the valley as the thick underlying Colorado River aggradation package underwent incision. Several ages and terrace levels of such fans are preserved in the landscape to the east. The alluvium of Bullhead City into which these fans are inset can be traced intermittently to the east from here to valley-flank elevations as high as 1100 ft above the river, indicating a 656-foot thick valley fill represented by the unit.

Walk northwest up a narrow tributary through sandstone exposures of the Bullhead City unit to the head of the gully. Overlying units include, from the lowest: a thin alluvial-fan deposit of basalt boulders (unit C, Metzger and Loeltz, 1973) overlain by a thick sequence of cliff-forming pale orange layered mud and very fine sand making up the lower part of the Chemehuevi Formation (Longwell, 1936; called unit D by Metzger and Loeltz, 1973), in turn overlain on the hilltops by 20 m of slope-forming well sorted, unconsolidated, light-toned sand and minor gravel of the upper part of the Chemehuevi Formation (unit E of Metzger and others, 1973; Metzger and Loeltz, 1973).

The Chemehuevi beds (pinkish silts and sands) represent one or more cycles of aggradation (unit D) and subsequent degradation (unit E) of the river valley (Metzger and others, 1973). The best existing dates for these beds are late Pleistocene, between 35 and about 100 Ka (Bell and others, 1978; Blair, 1996; Lundstrom and others, 2004; Malmon and others, 2006, See discussion at Stop 2-12). However, these dates are not consistent with the presence of the early Pleistocene *Mammuthus meridionalis.* Sedimentary deposits that contain *M. meridionalis* are significantly older (1.7 and 1 Ma) than Chemehuevi deposits and may represent a different sequence of fluvial events.

RETRACE to pavement

132.7 (0.3) Stop at highway; TURN LEFT (south).

133.4 (0.7) Continue past Five-Mile Landing.

134.1 (0.7) Flat-bedded sandstones crop out on the east side of the road.

135.2 (1.1) TURN LEFT (east) on Powell Lake Road.

135.3 (0.1) Stop at Route 66, Oatman Highway; TURN LEFT (north).

136.3 (1.0) Continue past Casa Grande Road. Prepare for curves and dips ahead.

137.6 (1.3) Note the red clays of the lower part of the Chemehuevi Formation on the left (west).

138.2 (0.6) Continue past a BLM kiosk.

139.3 (1.1) Cross a major gully.

139.4 (0.1) PARK on the right side of the road at the left bend. **STOP 2-11**. Plio-Pleistocene Deformation. View east across the gully shows south-dipping alluvial fan deposits cut by a pair of conjugate west-northwest-striking normal faults. An undeformed basalt-clast fanglomerate unconformably overlying the deformed rocks has a two meter-thick pedogenic carbonate soil. The deformed sediments are at least as old as early Pleistocene. The undeformed capping fanglomerate is one of a downward succession of post-Bullhead fans from the Black Mountains graded toward the central part of the valley (see also Metzger and Loeltz, 1973).

A mile east-southeast of this location, the Needles graben (Purcell and Miller, 1980; Pearthree et al, 1983) and a monocline and small transverse thrust fault deform at least two younger Pleistocene alluvial surfaces, including one capped by thick calcic soil (House and others, 2005). The graben also deforms middle and upper Quaternary surfaces (Pearthree and others, 1983). The Needles graben and related monocline and small thrust directly overlie a gravity low indicating thick low-density sedimentary fill (Gray and others, 1990). Sedimentary compaction of this fill may explain the graben and monocline. The shape of the gravity low, however, does not offer an obvious explanation for the southwestern downthrow or for possible sagging of sediments under the river valley as suggested by Metzger and Loeltz (1973). Quaternary graben structures have been reported in alluvial fans near the river 30 km south in Chemehuevi Valley and 100 km south near Blythe (Purcell and Miller, 1980). The cause of the deformation recorded by these grabens is uncertain. Return to vehicles. Use caution entering Route 66. RETRACE south to Golden Shores and Polaris Road.

143.4 (4.0) Continue past Powell Lake Road.

143.5 (0.1) Stop at the Hwy 1/227 junction with Hwy 66. Proceed south on Route 66.

144.1 (0.6) TURN LEFT (east) on Polaris Road.

144.6 (0.5) Proceed on Polaris Road as it bears right.

145.1 (0.5) TURN LEFT at the asphalt batch plant.

145.2 (0.1) Cross a wash.

145.7 (0.5) **STOP 2-12** on top of terrace. Golden Shores Mammoth. These sediments at one time were the source of a nearly complete skeleton of a young mammoth. The mammoth settled on its back in shallow water and was covered with fine clay. Imprints of reeds and sedges suggest a still, fresh water marshy environment (Agenbroad et al, 1992). Mammoth teeth were not available for examination. However, Agenbroad (1995 and 2005, written comm.) made a tentative identification and age evaluation based on a tooth photograph as *Mammuthus meridionalis*, dating as old as 1.7 Ma. Dates of *M. meridionalis* from the Ocotillo and Hueso Formations at Anza-Borrego Desert State Park to the west are as recent as 0.9 Ma (Jefferson, p.c. 2007). Therefore, deposits that contain *M. meridionalis* probably fall within the first half of the Pleistocene. Mammoth fossils along the Colorado River are known from the Lake Mead–Las Vegas area south to El Golfo, Sonora. Three localities of *M. meridionalis* are known from Needles to Parker. Younger mammoth forms include *M. imperator* and *M. columbi*, which are reported from Las Vegas, Nevada and El Golfo, Sonora (Johnson and Miller, 1980; Agenbroad and others, 1992).

Sediments containing the fossil mammoth lie at an elevation of 680 feet (207 m). The large, dense mammoth bones occur within a section containing locally derived angular gravel with reworked roundstone pebbles that interfinger with Colorado River mudstone, sandstone, and gravel. Nearby within the deposits at about the same stratigraphic level as the mammoth, root casts below a minor paleosol indicate that deposition of the fluvial section was discontinuous.

Mammoths are not known in North American before the Pleistocene (1.75 Ma, Agenbroad, p.c. 1995; Woodburne and Swisher, 1995). The age of the facies suggest deposition during an early Pleistocene aggradational episode of the river. If so, this postdates the alluvium of Bullhead City and predates or includes early deposition of the Chemehuevi Formation.

The working model is that the beds containing the mammoth are part of a Pleistocene sequence inset into the Pliocene alluvium of Bullhead City. Similar beds containing roundstone gravels, interbedded muds and sands, and root casts are exposed to the south in Sacramento Wash.

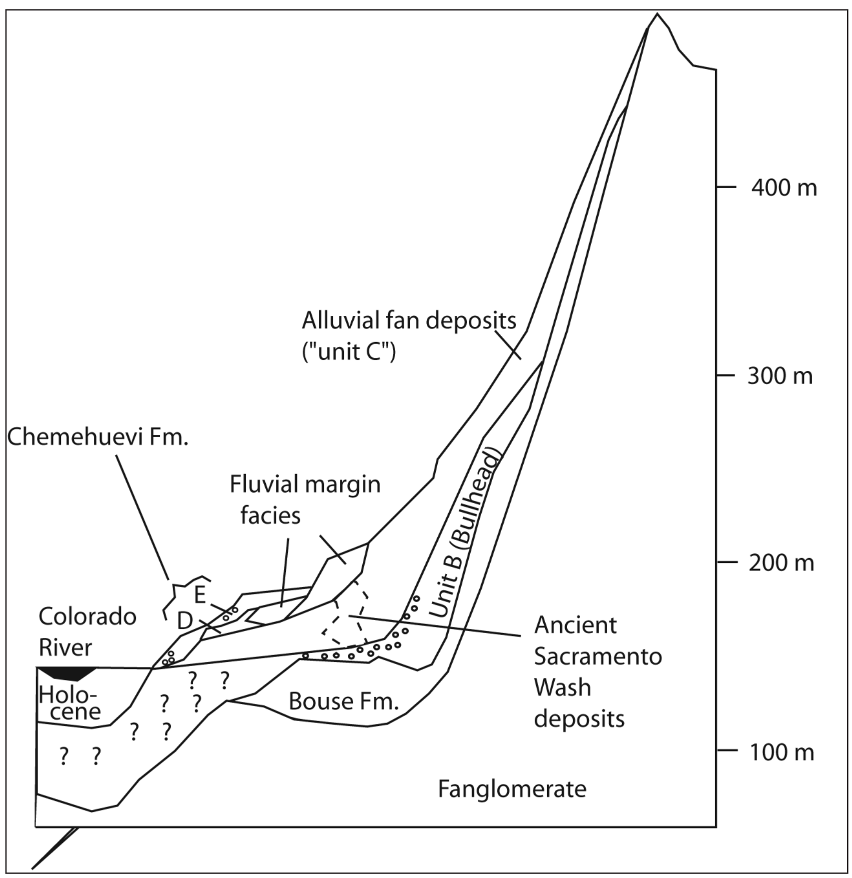


Fig. 2-8. Stratigraphic section near Stop 2-13 (K.A. Howard). Circles indicate boulder deposits.

RETRACE to Hwy 66.

147.3 (1.6) Stop, TURN LEFT (south) on Hwy 66 toward I-40.

148.5 (1.2) Drop into Warm Springs Wash.

149.1 (0.6) Catfish Paradise.

149.9 (0.8) Sacramento Wash.

150.7 (0.8) Railroad tracks on the left.

151.0 (0.3) Slow.

151.1 (0.1) Cross under railroad trestle. The road bears east, parallel to I-40. Prepare to turn left.

151.4 (0.3) TURN LEFT off the frontage road and drive northeast alongside the railroad tracks.

152.1 (0.7) **STOP 2-13**. An unconformity in the railroad cut exposes pre-Bouse Miocene fanglomerate of angular pebbles derived from the Chemehuevi Mountains, imbricated northward indicating that the ancestral Mohave Valley depocenter was to the northeast, so there was no stream exit though the site of Topock Gorge. Many of the pebbles are green from chlorite and epidote alteration, typical of rocks exposed in the footwall of the Miocene Chemehuevi detachment fault in the Chemehuevi Mountains (John, 1987). The fanglomerate is overlain unconformably here by fluvial gravel and sand containing round pebbles reworked from the Pliocene alluvium of Bullhead City and deposited in an ancestral Sacramento Wash (Fig. 2-8). PROCEED NORTHEAST along tracks.

153.1 (1.0) **STOP 2-14**. Pliocene sandstone and conglomerate of the alluvium of Bullhead City. Look northwestward half a mile across the Sacramento Wash at light-toned gypsiferous mudstone, a facies of the Chemehuevi Formation. We interpret this as being deposited in a marshy environment formed as the aggrading main stem of the Colorado River raised the local base level for Sacramento Wash. Fossil snail shells have been found here that indicate a freshwater marsh environment. PROCEED EASTWARD along tracks.

153.9 (0.8) **STOP 2-15**."T" intersection near a railroad trestle with a road from the south. View north across Sacramento Wash at folded Colorado River deposits (Fig. 2-9) that were pictured by Lee (1908) and Metzger and Loeltz (1973). Metzger and Loeltz assigned folded beds to their unit B. Unit B includes conglomerates and mudstone, overlying typical sandstone of the alluvium of Bullhead City, and resembling the beds that host the mammoth at Stop 2.12. Outcrops west of the T-junction are alluvium of Bullhead City. The road cut east of the junction exposes younger fluvial sand and gravel deposited in ancestral Sacramento Wash. From this "T" intersection, we will reach I-40 by driving south, uphill and following roads to the Needles Mountain Road overpass.

155.1 (1.2) Stop at pavement at Y-junction. Do not take the road east; proceed southeast.

155.30 (0.2) Stop, TURN RIGHT at Needles Mountain Road, and proceed through the gates. Ahead are The Needles, pinnacles of synextensional, steeply west-tilted lower Miocene volcanic, intrusive, and sedimentary rocks (Howard and John, 1997). The elevation at the base of the Needles is around 1800 feet. To the left (northeast) of The Needles, the upturned metamorphic basement substrate of this section is exposed to paleodepths of 8 km below the Miocene rocks. All of these rocks were tilted and structurally superposed by eastward fault slip on the Chemehuevi detachment fault over the Chemehuevi Mountains metamorphic core complex in Miocene time (John, 1987; Miller and John, 1999).

Fig. 2-9. "T" intersection near the railroad trestle. The view north across Sacramento Wash shows folded Colorado River deposits (R.E. Reynolds).

155.6 (0.3) ENTER I-40 WEST BOUND toward the Colorado River (Topock Gorge is to the south) Look southwest across the river to gently east-sloping rock surfaces on the California side. These surfaces approximate the exhumed Chemehuevi detachment fault, which projects eastward under the allochthonous rocks of The Needles. The river's path tracks this structural boundary southward through Topock Gorge.

157.8 (2.2) Continue past Exit 1, Golden Shores, Oatman exit (Topock Road).

158.4 (0.6) Cross the Colorado River.

159.6 (1.2) EXIT for Park Moabi Road

160.2 (0.6) STOP at overpass. TURN RIGHT (north) at Park Moabi and proceed down the hill ~0.5 miles past Chemehuevi Formation silts and interbedded locally derived gravel, which rest on mudstone of the Bouse Formation. Well-bedded mudstones visible 1/4 mile to the left (northwest) are the interbedded member of the Bouse Formation. TURN RIGHT (east) at the first intersection and follow this paved road one mile.

161.7 (1.5) **STOP 2-16**. Continue past a gravel road to the right, which leads to unrestored parts of historic Route 66, and PARK at the left shoulder near the trailers before reaching the overhead railway trestle. Road cuts here expose the pre-Bouse fanglomerate (under the trestle) overlain unconformably by both Pliocene (Bullhead City unit) river gravel, and inset pinkish fine sands of the Chemehuevi Formation. Half a mile to the northwest, and 3 km to the ESE across the river, the Bullhead City unit contains boulders as large a 1 m (Metzger and Loeltz, 1973b). The latter exposures are in the fill of a paleovalley, the base of which is a few meters above the modern river. One-half mile south, beyond the trestle and highway overpass, dipping red Miocene conglomerates in the upper plate of the Chemehuevi detachment fault sit on gray-green gneisses of the lower plate in the Chemehuevi Mountains. The Miocene conglomerate was deposited during extensional faulting and tilting (Miller and John, 1999).

Topock remediation effort

Recent drilling related to groundwater remediation efforts has provided new information on the subsurface stratigraphy, including identification of tens of meters of river-laid sand below the west bank of the Colorado River. Radiocarbon dates from buried wood fragments recovered from cores beneath the floodplain near Topock suggest that, prior to the construction of the dams, the river was aggrading about 3 mm/yr since the mid-Holocene. The upper part of this sand section is believed to be historical, as the river aggraded about 8 m in the mid 20th century in the backwaters of Lake Havasu (Metzger and Loeltz, 1973). Wood fragments collected from fluvial sand at depths of 60–90 feet recently yielded mid-Holocene C-14 ages. This supports the conclusions of Metzger and Loeltz (1973) and Metzger and others (1973) that the pre-dam river aggraded about 20–30 m in the Holocene. They based that conclusion on Holocene-age wood fragments recovered from drilling in fine-grained fluvial sediments beneath the floodplain 100 km downstream (near Blythe), and from a similar but undated section drilled 15 km north of this spot (Peter Martin: pmmartin@usgs.gov). RETRACE to I-40.

163.2 (1.5) Stop at Park Moabi Road, TURN LEFT (south).

163.7 (0.5) Proceed over the freeway.

164.0 (0.3) Pavement ends. TURN LEFT on the graded road.

164.8 (0.4) Continue past a left turn to an evaporation pond.

164.9 (0.1) The road bears left.

165.1 (0.2) Continue past a right turn to a quarry. The road cut exposes Chemehuevi overlain by a well developed soil profile that includes ed silts and underlying B-horizon in which carbonate has been remobilized into carbonate kernels.

165.3 (0.2) TURN LEFT: do NOT pass “Not a Through Road” sign. TURN NORTH and PARK. **STOP 2-17**. This is the Mystic Maze or Topock Maze (Fig. 2-10) (D.G. Thompson, 1929:517 and plate 30; Haenszel, 1978). “The maze has always been here” are key words to remember when viewing the maze. The river narrows were first called Red Rock (after the red upper plate Miocene sediments), then named Mellen (Mendenhall, 1909) after a river boat captain, and finally Topock (Thompson, 1921). The Topock Maze consists of three components. A curvilinear 100´ x 60´ anthropomorphic figure was destroyed by construction of the California Southern/Atlantic and Pacific/Santa Fe Red Rock cantilever bridge in 1888 (Myrick, 1963) during rip-rap collection for berm and roadbed for tracks laid in 1893. A curvilinear “eye”

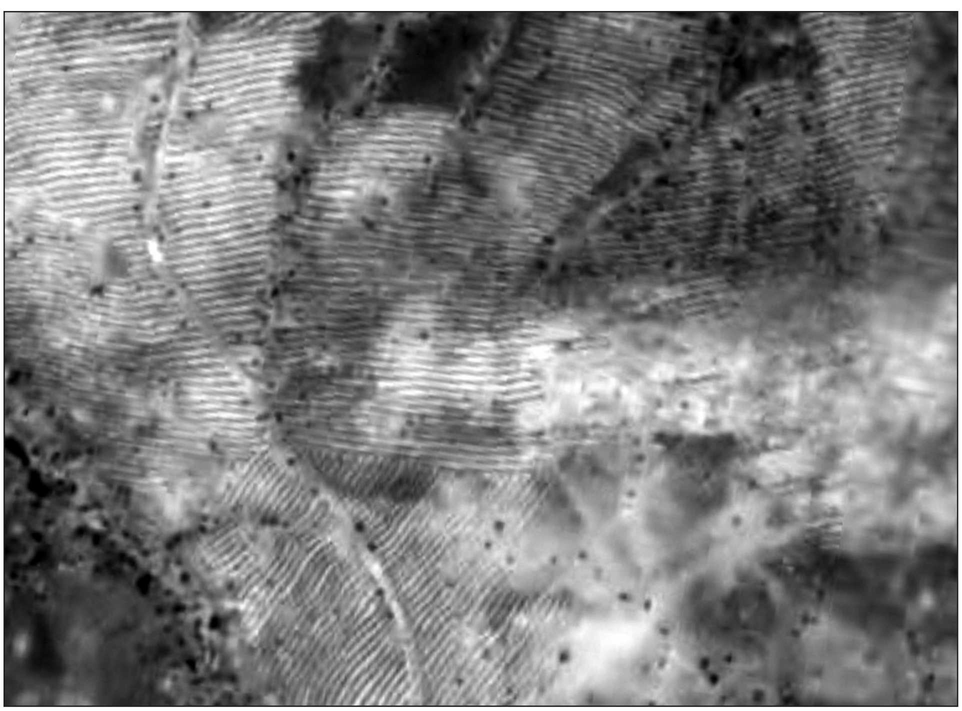


Fig. 2-10. Aerial photograph of the Topock Maze.

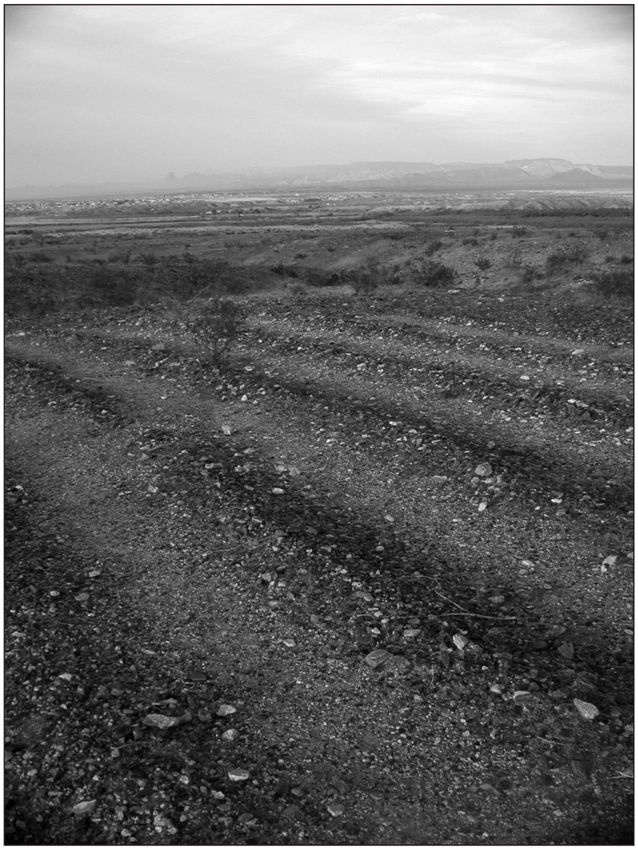


Fig. 2-11. Windrows cover vanished desert pavement on terrace lobes that stop before the terrace margin. This is not a true "maze" since the ends of windrows are not closed. The rows do not follow contours, but transgress slopes and surfaces (R.E. Reynolds).

figure (from "hook and eye") remains intact. The other component of the maze consists of parallel gravel windrows (Fig. 2-11). Dark vanished desert pavement on terrace lobes is covered with windrows that stop before the margin of the terrace (Fig. 2-12) and rarely cover all of the terrace surface area. These geometric rectilinear windrows are referred to as the "maze," but the ends of windrows are not closed except where they approach other windrows. The rows do not follow contours, but transgress slopes and surfaces (Haenszel, 1978). Another "maze" occurs at Topock on the Arizona side of the river. "In 1932 I flew over…the maze…on the California side…and a smaller maze on the Arizona side near the railroad bridge. The latter had been made by scrapers when the bridge was built…" (Woodward, p. 28 in Haenszel, 1978).

Fig. 2-13. Ground figures (intaglios) near Blythe (R.E. Reynolds).

Both curvilinear figures associated with the Mystic Maze are similar to other ground figures (intaglios, Fig. 2-13) on terraces along the lower Colorado River. However, no other examples of parallel windrows of proven aboriginal

Fig. 2-12. Windrows at the Topock Maze rarely cover all the surface area of the terraces, ending before the terrace margins (R.E. Reynolds).

Fig. 2-14. Low, parallel rows at the Topock Maze (R.E. Reynolds).

origin are known (Haenszel, 1978). The use of Mohave Indian labor to scrape desert pavement gravel into rows to be collected for construction of bridge abutments, berms, and road grade is well documented (Haenszel, 1978). Interviews with Mohave Indians indicate that the maze (windrows) was not used by the Mohave tribe, but was made by the "Old Ones" (Thompson, 1929; Haenszel, 1978). Haenszel (1978) notes that windrows were probably made by scraping or raking desert pavement. However, she presents documentation from observers that the maze was present in 1886, prior to construction of the Red Rock cantilever bridge. Haenszel relies heavily on Mohave statements that the maze was in place prior to their arrival (A. D. 1200), although used for religious purposes by the Mohaves.

Questions arise: Were the Mohaves referring to the intaglios or to the windrows? The windrows avoid the curvilinear figures. Did Mohave laborers avoid the intaglios for religious reasons? Amassing gravel in windrows over 18 acres involves a great deal of effort. Who paid for this labor and why was the gravel not collected or used?

Historic photographs (Pl. 30-B, Thompson, 1929; photos from 1897, 1922, 1931 in Haenszel, 1978) show gravel rows as steep-sided ridges. Today, they are deflated to low, rounded rows. If rows were steep-walled in 1897 and almost flat in 2007, this author (RER) suggests that the rate of deflation is finite, and doesn't allow for construction of windrows prior to the mid nineteenth century. Read the archaeological description by the Mohave Nation mounted at the viewpoint. Examine the low, parallel gravel rows (Fig. 2-14). Look for black desert varnish (upper surface) and red oxide (under surface) that indicate the cobbles were disturbed. Have a great discussion at the end of Day 2.

RETRACE to the I-40 Park Moabi exit.

166.6 (1.2) Cross over I-40 at the Park Moabi overpass.

Summary of Day 2. This rapid trip has followed the distinctive white, basal Bouse limestone in the early Pliocene paleotopography of Cottonwood and Mohave Valleys. The deposition of the Bouse Formation dates between 4.1–5.6 Ma. The Bouse is overlain by Colorado River deposits including the basal alluvium of Bullhead City and pink, silty overbank deposits referred to as the Chemehuevi beds. Mammoth fossils indicate that local deposition dates to early Pleistocene, between about 1.7 and 1 Ma (Stop 2-10).

End of Day Two

Day Three

Start Day 3 from the Park Moabi exit. We will follow the event-marking Bouse Formation with white limestone through Chemehuevi Valley, Parker Valley and the La Posa Plain (Stops 3.1–3.6). Wild roads take us rapidly to other members of the Bouse Formation. In Chemehuevi Wash we will see colorful siltstone of the basin-filling middle member. Bouse tufa, the upper member of the Bouse Formation southwest of Vidal, marks the high stand of the body of water that filled Parker Valley. Bouse limestone in Chemehuevi Wash west of Lake Havasu is at 560 feet elevation. Bouse limestone in Osborne Wash southeast of Parker is at 600 feet. Upper Bouse tufa south of Vidal in the Riverside Mountains is at least 650 feet, suggesting inundation of Parker Valley by a body of water at least fifty feet deep. We will visit Osborne Wash, pass through Parker, Arizona and end at Vidal Junction on CA Hwy 95.

0.0 (0.0) ENTER I-40 WEST BOUND, toward Needles (excerpts from Reynolds, Buising and Beratan, 1992).

0.7 (0.7) Pass through pink Chemehuevi Formation sediments.

0.9 (0.2) View west of green mudstones of the Bouse Formation. Pink Chemehuevi sediments are inset into the mudstones.

4.0 (3.1) Pass through the Agricultural Inspection Station.

4.6 (0.6) EXIT at Five Mile Road/U.S. 95 south.

5.0 (0.4) Stop at end of off ramp, TURN LEFT, and proceed west.

6.7 (1.7) Stop, TURN LEFT (south) on CA Hwy 95.

8.9 (2.2) Note the south-dipping, red Miocene sediments in the upper plate of the Chemehuevi detachment fault.

9.9 (1.0) Pass porphyritic granitic rocks of the lower plate of the detachment.

11.8 (1.9) Continue past the South Needles Compressor Station.

12.6 (0.8) Slow through curves.

16.7 (4.1) Approach summit.

19.3 (2.6) Snaggletooth Summit. Snaggletooth pinnacle is Miocene porphyritic dacite in a 700 m thick section of Tertiary volcanics dipping 40–90° southwest. The volcanics make up the hanging wall to the regionally developed detachment fault system (Howard and others, 1993; Howard and others, 1994). The 18.5-Ma Peach Springs Tuff is at the top of the section. The Tertiary rocks noncomformably overlie Proterozoic gneisses, granite and diabase dikes. To the left in the Chemehuevi Mountains, the dark Proterozoic rocks are superposed in fault contact along the west dipping Chemehuevi detachment fault (John, 1987) on the light colored Chemehuevi Mountains plutonic suite (Late

Cretaceous, John and Mukasa, 1990; John and Wooden, 1990). Proceed downhill toward Chemehuevi Valley.

21.8 (2.5) Prepare for a left turn.

22.0 (0.2) TURN LEFT (southwest) onto Lake Havasu Road.

26.4 (4.4) Granitic outcrop on left is cut by dark diabase dikes. Gray diabase contains dark green pyroxene spheroids that are hollow; the hollows now filled with calcite.

28.0 (1.6) Cross the Needles–Vidal–Blythe Road (Thompson, 1921, 1929).

31.8 (3.8) Continue past a power line road running south-southeast to Chemehuevi Wash and marked by stockpile of white Bouse marl. West Well in Chemehuevi Wash was a stop on the Needles–Vidal–Blythe Road (Thompson, 1929). Automobile technology improved and drivers became less dependent upon water holes, wells, and springs, and the Needles–Vidal–Blythe route was straightened and upgraded in the 1930s to become CA Hwy 95.

36.1 (4.3) Havasu Lake Road bends northeast.

36.7 (0.6) Havasu Lake Road bends northeast.

37.4 (0.7) Havasu Lake Road bends northeast. Look for the approaching pole line at a left (northeast) curve in Havasu Lake Road.

37.6 (0.2) TURN RIGHT (south) on a pole line road into Chemehuevi Wash. Follow the pole line road south, up the next terrace

38.3 (0.7) Top of terrace—watch your odometer carefully. Pass through three dips.

38.5 (0.2) The road bears right at a BLM post on top of the terrace, take the RIGHT FORK. Proceed south-southwest on BLM bypass road built on firm substrate.

39.0 (0.5) The road bears south-southeast through a cut. Proceed to the south side of the wash.

39.5 (0.5) TURN RIGHT (west) at mid-wash, and proceed west to the basal part of the Bouse Formation. We can see orange and green silts of the middle member of the Bouse Formation. The Bouse Formation is unconformably overlain by fluvial gravels with well-rounded cobbles of basalt, tuff, gneiss, and granitic rocks, suggesting a source in the Turtle Mountains to the west. Rocks from the Colorado Plateau appear to be absent. Proceed west through multicolored Bouse exposures.

40.6 (1.1) **STOP 3-1**. On the north side of the wash, faults disrupt tan sandstone in greenish gray clays (Fig. 3-1). Sense of movement appears down-to-the-east. PROCEED southwest.

41.2 (0.6) **STOP 3-2**. WALK WEST into canyon and examine basal limestone (Elevation 560´) of the Bouse on

Fig. 3-1. Down-to-the-east faults disrupt tan sandstone in greenish gray clays of the interbedded member of the Bouse Formation (R.E. Reynolds).

Miocene volcanic rocks. Return to vehicles and PROCEED WEST up-drainage.

41.5 (0.3) The road bears west up the wash.

41.7 (0.2) Where the road narrows, examine angular conglomerates under Bouse marl.

41.9 (0.2) **STOP 3-3.** End of the road. The Bouse sits on fanglomerate and Miocene breccia. RETRACE to pole line road.

44.2 (2.3) Pole line road. TURN LEFT (north) at pole line. TURN LEFT (north-northwest) up the BLM trail.

44.7 (0.5) The road bears northeast.

45.1 (0.4) TURN LEFT (north) on the pole line road.

46.0 (0.9) Stop at Havasu Lake Road. Watch for traffic. TURN LEFT (west) and proceed toward Hwy 95.

51.7 (5.7) Continue past a pole line road with a white stockpile.

61.1 (9.4) Stop at Hwy 95. TURN LEFT (south) and proceed toward Vidal Junction.

62.9 (1.8) Continue past a right turn to Mohawk Spring, Ward Valley, and the Old Woman Mountains.

69.6 (6.7) Cross Chemehuevi Wash.

74.1 (4.5) View west of the Turtle Mountains, which consist of Proterozoic granite and gneiss (Howard and others, 1982, 1988). The Mopah Range and Turtle Mountains, both 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 of the Whipple detachment fault. View at 2:00 (southwest) of Mopah Peaks at the north end of the Mopah Range on the east side of the Turtle Mountains. The Mopah Range consists of Tertiary volcanic rocks, including the remains of volcanic vents (Hazlett, 1993; Nielson and Hazlett, 1993). 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 Whipple Mountains consist of a metamorphic core including dikes and mylonites, overlain structurally above the domed Whipple detachment fault by tilted Tertiary volcanic and sedimentary rocks and their depositional basement of gneiss and granitoids (Davis and others, 1980; Carr, 1991). View at 10:00 (southeast) of Pyramid Butte in front of Savahia Peak, located west of Chambers Well and the Whipple Mountains. Savahia Peak consists of tilted Miocene rocks.

91.6 (17.5) Stop at Vidal Junction, where CA Hwy 95 meets Hwy 62. PROCEED SOUTH toward Vidal.

97.1 (5.5) Continue through the community of Vidal.

97.3 (0.2) Cross railroad tracks.

98.1 (0.8) CA Hwy 95 bends southeast.

100.1 (2.0) CA Hwy 95 bends south. Prepare for a right turn.

100.4 (0.3) TURN RIGHT (west) on the dirt pole line road and proceed west across the terrace.

101.1 (0.8) BEAR LEFT at the intersection.

101.2 (0.1) Continue past a reverse turn. Proceed south across a wash, then southwest across the terrace on the north side of the Riverside Mountains.

101.4 (0.3) **STOP 3-4**. Bouse Formation Shoreline Tufa. Walk south to gneissic basement rocks (black and dark red-brown) with a discontinuous veneer of light gray, spongy, upper Bouse Formation shoreline tufa (Fig. 3-2). This is the first locality where we have seen the tufa that is the upper unit of the Bouse Formation (Metzger and others, 1973a). Outcrops of tufa continue south along the Colorado River trough. Tufa is gray, dendritic, and spongy in contrast to the white basal limestone/marl of the Bouse Formation. Tufa forms a rind on basement exposures; locally, the base of the tufa includes a monolithologic breccia of angular basement clasts cemented by gray carbonate. 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. The elevation of this outcrop above Bouse basal limestone at our next stop in Osborne Wash suggests that Parker Valley was inundated by a body of water at least fifty feet deep. More than one layer of tufa is present in some places. Explore the shoreline to find exposures of clastic limestone and coarse clastic strata interbedded with the tufa. The view northeast shows the Whipple Mountain detachment marked by dark volcanic rocks over the light-colored metamorphic core complex. RETRACE to CA Hwy 95.

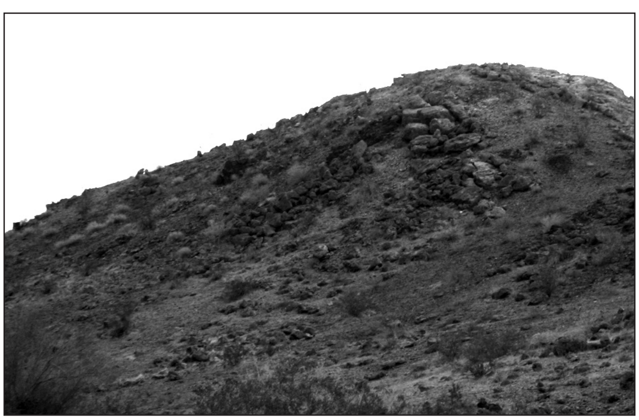


Fig. 3-2. Light gray, spongy tufa member of upper Bouse Formation caps Paleozoic metamorphic rocks (R.E. Reynolds).

102.6 (1.2) Stop at CA Hwy 95, watch for cross traffic, andTURN RIGHT (south).

103.0 (0.4) Enter Riverside County.

105.5 (2.5) TURN LEFT (east) onto Agnes Wilson Road (Indian Hwy 18).

106.0 (0.5) The road cuts through Quaternary Colorado River sediments.

108.0 (2.0) Cross the Colorado River into Arizona.

110.6 (2.6) TURN LEFT (north) on Mojave Drive.

113.6 (3.0) TURN RIGHT (east) on Burns Road.

115.5 (2.9) Cross 4th Avenue.

117.5 (2.0) Cross 1st Avenue. The pavement ends and the road bears northeast.

119.2 (1.7) Stop at AZ Hwy 95. TURN LEFT (north) and proceed to Shea Road.

121.8 (2.6) TURN RIGHT (east) on Shea Road.

122.0 (0.2) Stop at the railroad tracks; proceed on Shea Road.

122.7 (0.7) Cactus Plain to the right (south) is covered by semi-stabilized dunes which overlie the Bouse Formation. The fine-grained sand may have been derived by reworking the Bouse Formation.

124.4 (1.7) Curve northeast. Black Peak is at 3:00. Note the dark varnished basalt debris covering steep slopes, indicating that the slope surfaces have been stable for a relatively long time.

126.8 (2.4) Drop off the terrace. Cross, but do not drive up, the sandy wash.

127.1 (0.3) **STOP 3-5**. Osborne Wash. PARK beside the road or on gravel north of the wash. The western half of Black Peak is part of the Colorado River Indian Tribes reservation. This mountain is sacred to CRIT—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 to the northeast. The middle 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 Peach Springs Tuff. The red beds are conformably overlain by olivine basalts at the top of Black Peak. The basalt stack contains an angular unconformity within it. The flat-lying Pliocene Bouse Formation and immediately underlying fanglomerates are separated from older units by a buttress unconformity.

Walk west along Shea Road to the point where it enters Osborne Wash. Walk southeast along the wash to outcrops in the west wall of Osborne Wash (Fig. 3-3) and in smaller tributaries entering Osborne Wash from the southwest. The buff to pale pink unit is an Upper Miocene fanglomerate (Osborne Wash strata of Buising, 1990; Osborne Wash Formation of Davis and others, 1980) capped by 1–2 m of yellow sandstone. The yellow sandstone locally interfingers with the bright white, well-bedded basal carbonate of the Bouse Formation. The carbonate is in turn overlain by up to 10 m of poorly exposed green silt of the Bouse Formation basin fill association (Buising, 1988, 1990; Smith, 1960, 1970).

Things to notice in these exposures include:

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.

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

White carbonate: Basal Bouse limestone at 600´ elevation contains locally abundant ostracodes and cm-scale gastropods; well-developed bedding, locally draped over cobble- and boulder-grade clasts in a basal lag horizon; also note stromatolitic or biohermal structures developed on some of the lag clasts. This is a good place to discuss fossil evidence for the environment of deposition of Bouse sediments from Parker Valley and Danby trough and to the south.

RETRACE west on Shea Road to AZ Hwy 95.

132.3 (5.2) Stop, cross the railroad tracks.

132.5 (0.2) TURN RIGHT (north) on AZ 95.

133.0 (0.5) Enter Parker, Arizona on California Avenue (AZ Hwy 95).

134.0 (1.0) Continue past Riverside Drive.

135.0 (1.0) Cross the Colorado River into California. In the road cut, the Bouse Formation overlies fluvial cross-bedded gravel and sand. The matrix contains minor amounts of clasts derived from the Colorado Plateau (Buising, 1990). These lithologies may mark the upper Colorado River's entrance into the Parker basin.

Fig. 3-3. Osborne Wash, exposures of Bouse limestone overlying cross-bedded yellow sandstone. D.V. Malmon photograph.

135.5 (0.5) Stop at CA Hwy 62. PULL RIGHT and PARK on the south side of the highway. **STOP 3-6**. Earp. This town is named after Wyatt Earp, who hired on to guard stage coaches transporting gold, and eventually stayed to prospect and live in this area.

Walk back to the cut and inspect white Bouse marl structure and relationships. Note the excellent large-scale crossbeds in the yellow sands and gravels below the Bouse (Fig. 3-4 and back cover). Farther south, historic cabins were excavated below the resistant marl.

Bouse questions remain unanswered:

• Why is the base of the Bouse Formation in Cottonwood, Mohave, Chemehuevi, and Parker valleys always marked by white limestone? What conditions caused the limestone to precipitate when water entered a new basin?

• Elevations of the basal Bouse limestone on the Nevada

Fig. 3-4. Bouse Formation basal limestone and overlying basin-filling silts at Earp, California. Photo by D.V. Malmon.

pediment west of Bullhead City range from 780 to 1720 feet and provide an indication of paleotopography in the Mohave Valley basin. Bouse limestone elevations in other basins vary (560´ in Chemehuevi; 600´ in Parker; 1000´ in Bristol basin north of Amboy; 1000´ in Milpitas Wash). Can these elevations be attributed to transgression up topography in closed basins? At what point can basal Bouse limestone be considered constrained by sea level, with elevations being caused tectonically?

• Can the time of Bouse deposition in each basin along the sequence be better constrained? We have seen Bouse in Cottonwood and Mohave valleys constrained by ashes dating between 5.59 and 4.1 Ma. Outcrops of Bouse limestone facies to the south in Milpitas Wash (1000 feet in the SW ¼ Sec 34, T 11 S, R 20 E, SBBM) and north of Amboy in the Bristol Basin contain the 4.83 Ma Lawlor Tuff (A. Sarna-Wojcicki, p.c. 2007; M. Perkins, p.c. 2007). Does this occurrence help constrain the Bouse in Mohave Valley between 4.83–5.59 Ma?

Fossils may help answer some of these questions. Fossils in the Bouse Formation include microfossils (foraminifera, diatoms and ostracodes), megafossils (mollusks and barnacles) and vertebrate fossils (fish). Microfossils were once used as evidence extending a marine incursion as far north in the Colorado River Trough as the Grand Wash Cliffs, where the Hualapai Limestone would be, in part, a marine deposit (Blair, 1978). Microfossils in cores from Danby Basin have been used to correlate sediments at depth with the Bouse Formation (Brown and Rosen, 1992). The Cadiz Basin to the northwest also contains foraminifera indicating brackish water, but Brown and Rosen (1992) caution that studies of deposits in the Owens River basins containing brackish water foraminifera (Smith and others, 1983) have not supported a marine embayment into that drainage. Those authors suggest distribution of saline-tolerant microfossils by birds and other mechanisms.

Macrofossils have been noted in well cuttings from Parker to Blythe and south to Cibola Lake (Metzger, 1965, 1968, Metzger and others, 1973a, 1973b). These include marine clams, marine, brackish and fresh water snails, and *chara*, the latter supporting brackish to fresh water conditions (Metzger, 1973a). "Fossils are not common in the Needles area…" and are too poorly preserved for identification (Metzger, 1973b). Barnacles are noted in the Bouse Formation in the Parker area, some with unusual morphology attributed to the stress of a low saline environment (Reynolds and others, 1992: Buising, 1992). An inshore marine fish (*Colpichthys regis*) related to the jacksmelt and grunion was recovered from a quarry of Bouse marl north of Cibola Lake, Arizona (Todd, 1976). Certainly, the outcrops of Bouse Formation along the Colorado River Trough warrant more prospecting to recover additional fossils that would help determine the environment of deposition. The summary of available macrofossil and vertebrate fossil data suggests that marine to brackish water outcrops of the early Pliocene Bouse Formation occur from the vicinity of Parker south past Yuma, and possibly westward into Danby Basin. Subsurface data also suggest that Bouse-like sediments may be found along the Gila River as far as Gila Bend (Metzger, 1968).

ENTER HWY 62 WESTBOUND toward Vidal Junction.

136.2 (0.7) On the north side of the road, green sediments of the Bouse Formation (Metzger, 1968) are overlain by pinkish Colorado River sediments.

139.3 (3.1) Rio Mesa Road and Big River Development. The low, dark hills north of the road are made up of Tertiary andesite flows from the Whipple Mountains. On the southern flank of the range, the volcanics have been subjected to a type of alteration known as potassium metasomatism. Large volumes of potassium have been added to these rocks, with removal of sodium. K_2O values in typical andesites are commonly less than 1%; K_2O values as high as 16% have been measured from these andesites. Silica and iron were also added to both volcanic and sedimentary rocks, turning them red and resistant (Reynolds and others, 1992).

139.7 (0.4) Continue past a deep gully with dark Tertiary volcanics and red sandstone of the Turk Mine and Twin Lode Mine(?) formations, which predate the18.5 Ma. Peach Springs Tuff (Nielson, 1993).

150.9 (11.2) **STOP 3-7**. Chambers Well Road. Drive north and PARK on the elevated berm of the Colorado River Aqueduct. This road eventually leads to Turk mine and Chambers Well in the saddle between Savahia Peak and the Whipple Mountains. The latter is a well on the Needles–Vidal–Blythe Road running south from West Well in

Chemehuevi Wash (Thompson, 1921, 1929). The Whipple Mountains to the northeast consist of dark-colored Tertiary sedimentary and volcanic rocks in the upper plate of the Whipple detachment system; the light-colored rocks are mylonitic gneisses in the lower plate (Davis and others, 1980). The Whipple detachment fault is at the dark-light contact. The broad, low dome of the Whipple Mountains is characteristic of elevated metamorphic core complexes. Climatic changes during the Quaternary caused profound erosion, developing alluvial fans, pediments, and soils of the eastern Mojave Desert along the Colorado River (Bull, 1991). RETRACE to Hwy 62 and PROCEED WEST to Vidal Junction.

154.6 (3.7) STOP at Vidal Junction.

Summary

In three rapid days we have visited wild and scenic terrain created during Miocene time along the Colorado River extensional corridor. The resultant trough was filled by fluvial deposits, hypothetically followed by a "spillover model" that may have created Bouse lakes with distinctive white basal limestone. This trip has followed the distinctive white, basal Bouse limestone that constrains early Pliocene paleotopography between 4.1–5.6 Ma. In Cottonwood and Mohave valleys, the Bouse is overlain by Colorado River deposits including the alluvium of Bullhead City. Fossil mammoth indicates that local deposition of unnamed fluvial sediments were deposited in the early Pleistocene, between 1.7 and 1 Ma. Deposition of Chemehuevi beds dates between 80–30 Ka, in the late Pleistocene.

The Bouse deposits south of Parker contain barnacles and marine mollusks, suggesting estuarian, brackish water conditions or marine embayment. Near Parker, the Bouse section contains the white, basal marker, the colorful silts of the basin-filling member, and the upper "bath tub ring" of porous tufa. Dated Lawlor Tuff in Bouse outcrops of Milpitas Wash suggest deposition was as early as 4.83 Ma.

End of Day Three

CA Hwy 95 runs north to Needles (48 miles) and joins Interstate 40. CA Hwy 95 runs south to Blythe (48 miles) and joins Interstate 10. CA Hwy 62 runs west through Twentynine Palms (93 miles) and Morongo, or turns south and becomes Hwy 177 to Desert Center (68 miles). Both these routes join I-10.

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Major and trace element geochemistry of the Neogene Halloran Hills andesites, San Bernardino County, California: Implications to tectonic evolution of the eastern Mojave

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Abstract

The Halloran Hills, northeast of Baker, CA, were the site of Neogene (12.1 Ma) andesitic volcanism. Field sampling reveals that andesite outcrops consist of both brecciated and non-brecciated flows. Large phenocrysts of altered clinopyroxene are present in most samples. A smaller number contain euhedral, phenocrysts of orthopyroxene. XRF whole rock analyses indicate that Halloran Hills andesites are High-K trachyandesites, emplaced during continental, crustal extension. Two distinct sample populations were noted; High silica (SiO2 ≈ 62%) and Low silica (SiO2 ≈ 56%). Trace element data suggest both High and Low silica andesites are derived from a common source, the former representing a smaller melt fraction.

Cenozoic volcanic rocks in nearby Mesquite Pass and the Cima volcanic field were also examined for this study. The 13 Ma Mesquite Pass volcanics are a suite of alkaline rocks ranging in composition from rhyolite to trachydacite. The Cima volcanics are comprised of older basaltic trachyandesite (7.5 - 3 Ma) and younger trachybasalt (l.5 Ma - Pres.). When combined with data from the Halloran Hills, an extended pattern of alkaline volcanism emerges. It began during the initial stages of Miocene detachment and has continued to the Present, as upwelling asthenosphere occupied the void created by a thinning lithosphere. The compositional variations of the rocks reflect progressively deeper levels of melting, from shallow crustal rhyolite and trachydacite, to lower crustal trachyandesite to mantle trachybasalt. Assuming a stationary heat source for the volcanism, the eastern Mojave has undergone clockwise rotation at an average rate of 5 mm/year during the Neogene. This rotation involved two distinct events; a period of east-west extension, related to detachment, and a younger north-south directed event, a consequence of dextral shear.

Introduction

The Halloran Hills (HH) are 18 kilometers northeast of Baker, California (Fig. 1). They are bounded to the south by the Cima volcanic field and to the east and west respectively, by Shadow Valley and Silurian Valley.

The oldest rocks in the HH are 1.7 Ga metamorphic rocks, informally termed Fenner Gneiss. These have been intruded by Mesozoic granitoids of the Teutonia batholith (Beckerman, et. al., 1982). Teutonia intrusives range in composition from diorite to granite with monzogranites predominating (Wilshire, 2002).

Cenozoic stratigraphy is complex, influenced by Miocene detachment faults (Reynolds, 1991; Reynolds and Calzia, 1996) and subsequent dextral shear Reynolds, 1997). Peach Springs Tuff (18.5 Ma) unconformably overlies rocks of the Teutonia batholith (Nielson et. al., 1990). Freshwater carbonates, fine-grained clastics, gravels and andesitic volcanics (12.1 Ma) (Wilshire, 2002) lie above the Peach Springs Tuff. Where the tuff is absent, they are in direct contact with the Mesozoic intrusives. Detached blocks, debris flows and avalanche deposits of Proterozoic to Lower Paleozoic sedimentary rocks are often juxtaposed with the Tertiary units. Brecciated debris has moved westward from the Mountain Pass area to the east (Reynolds and Nance, 1988; Reynolds, 1991). Post-extensional erosion was followed by "glide" blocks moving great distances from source areas in the Avawatz Mountains to the west (Reynolds, 1991).

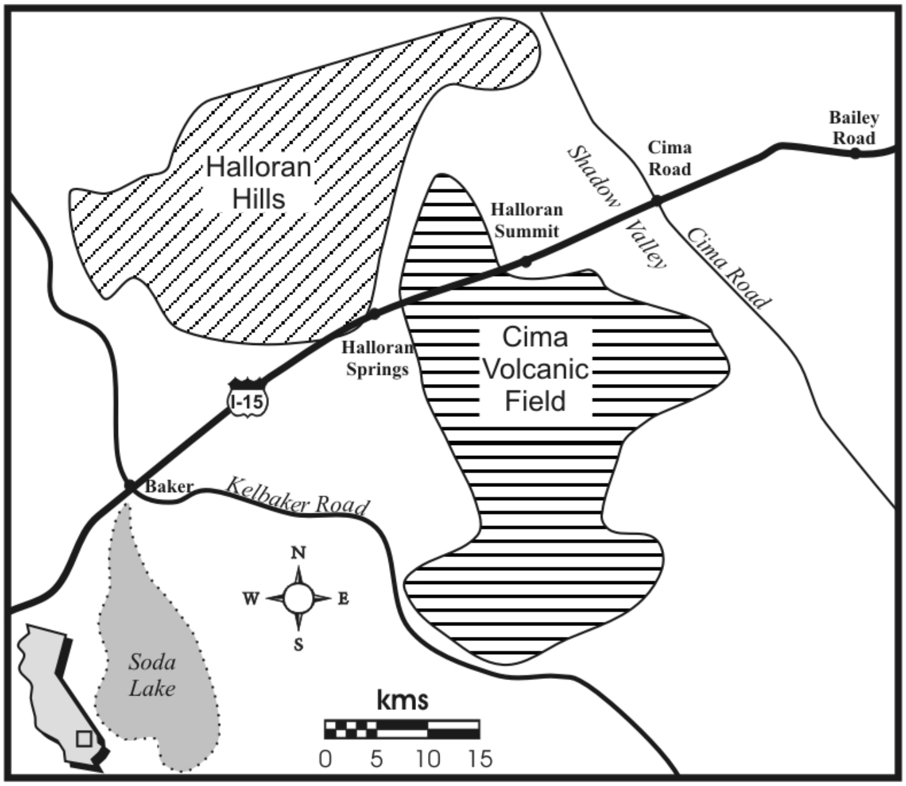

Figure 1. Index map showing the location of the Halloran Hills.

Sampling and Analytical Procedures

Rocks of andesitic composition outcrop at a number of localities in the Halloran Hills. No detailed geologic map exists, thus, stratigraphic and structural relationships are often uncertain. Two andesite age dates have been reported. One, from a plagioclase separate of a highly altered andesite, yielded an age of 12.8 Ma (Wilshire, 2002). The second, a "fresh" pyroxene andesite, was dated at 12.1 Ma (Wilshire, 2002).

Twenty-two samples of andesite were collected from the Halloran Hills. Samples varied widely in character. Those from the northern portion of the HH were often extensively brecciated. Breccias consisted of andesite clasts in a groundmass of andesitic composition, suggesting autofragmentation of flows. Some breccias samples, however, contained clasts derived from basement rocks. These breccias were extensively altered and silicified, perhaps due to emplacement along faults. Andesite samples from the southern HH were not brecciated.

Color and phenocryst mineralogy also varied. Many samples were medium to dark gray, but a small number from the southeastern HH were red-purple to red-brown.

Following erosion and tilting of the Miocene sedimentary rocks, basaltic volcanism began around 7.5 Ma (Turrin, et. al., 1984). While much of this volcanism occurred in the Cima volcanic field south of the HH, scattered 4.5-5.5 Ma flows are present in the southern Halloran Hills. After a hiatus of 1.5 to 2 million years volcanism recommenced at 1.5 Ma in the southern portion of the Cima field. Erosion has generated Quaternary alluvial deposits ranging from fine-grained sediments in the pluvial basins to coarse grained fanglomerates mantling the resistant ridges created by basalt flows.

The purpose of this research is to reexamine the "pyroxene" andesites of the Halloran Hills. It seeks to provide a more quantitative characterization of the rocks and answer fundamental questions regarding their petrogenesis. Specifically, do the andesites represent a single eruptive event or multiple events? Are the andesites related in time and space to other Cenozoic volcanism in the eastern Mojave? Do these relationships fit into a comprehensive tectonic model?

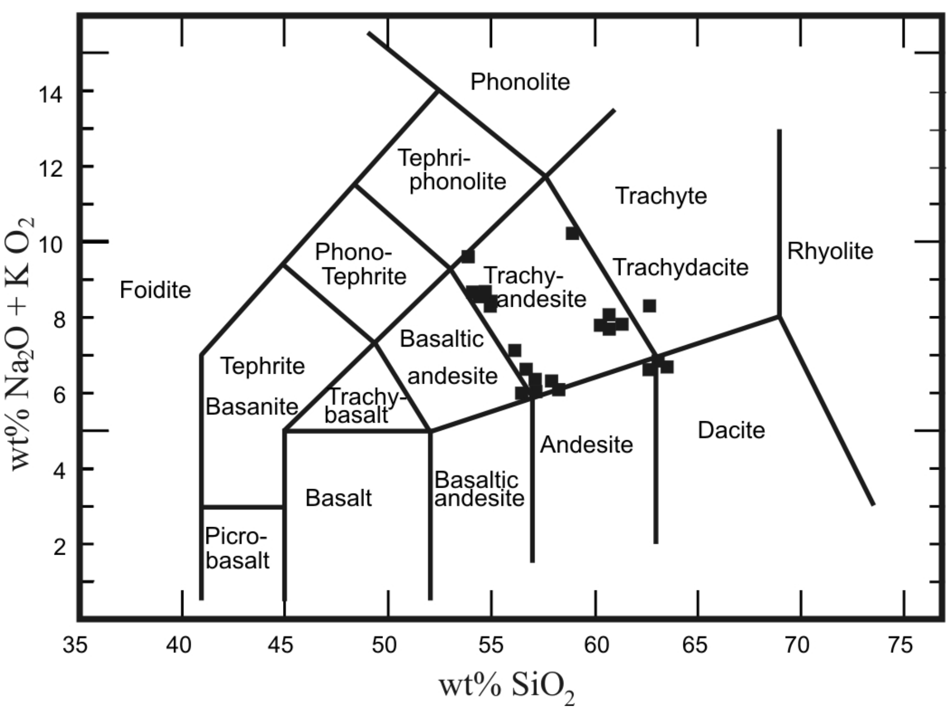

Figure 2. TAS diagram for the Halloran Hills andesites.

Table 1. Halloran Hills Andesites - CIPW Norm

	Q	or	ab	an	di	hy	mt	il	ap
High silica	13.27	23.22	30.77	15.51	8.27	3.96	2.49	1.21	1.12
Low silica	2.77	23.32	33.38	17.71	8.83	7.95	2.91	1.78	1.34

Trace Elements (ppm)

	Rb	Ba	Sr	Cr	Zr	Sc	La	Ce	Nd	Sm	Y
High silica	161	1203	1857	50	463	45	21	71	22	2	72
Low silica	163	809	1032	53	412	29	19	61	24	4	67

Seriate textured plagioclase was ubiquitous and large, euhedral k-feldspars common. Very large (up to 5 mm), phenocrysts of green to black clinopyroxene were also common. The clinopyroxene was slightly altered and embayed. The "red" andesites contain smaller (1-2 mm), euhedral, phenocrysts of wine-colored orthopyroxene. The orthopyroxene appeared largely unaltered. Biotite and hornblende are rare accessory phases.

Field samples were crushed, sieved and pressed into aluminum mounts for XRF analysis. Major elements were analyzed with software provided by PANanalytical Inc. Trace element analyses utilized "basaltrace" a software routine developed in-house by the Cal Poly Geological Sciences Department.

Geochemistry

Figure 2 is a TAS (Le Bas) diagram for the andesite samples from the HH. The samples span a diverse range of composition from dacite and trachydacite to basaltic trachyandesite, with the overwhelming majority within the trachyandesite field. The Le Bas diagram assigns rock name strictly on the basis of chemistry. No formal relationship to modal mineralogy has been proposed. Informally, however, pyroxene andesite and trachyandesite are likely synonymous. The andesites appear to define two distinct populations; termed High (9 samples) and Low silica (13 samples). The former group is composed of samples with >60 wt% SiO2, and the latter <58 wt% SiO2. To facilitate comparison, the CIPW norm was calculated for each group (Table 1), and a spider diagram constructed from minor and trace element data.

CIPW normative mineralogy shows little variation between groups, with normative Q and hy the exceptions. The disparity in normative Q results from a 6 wt% difference in average SiO2 content of the High (62%) and Low (56%) silica andesites. The normative hy anomaly is explained by the presence of both orthopyroxene and clinopyroxene phenocrysts in some samples. Samples with modal orthopyroxene would be expected to have significantly greater normative hy.

Figure 3 compares trace element data for High and Low silica andesites. Trace elements are normalized against the MORB standard of Pearce (1983). In general, both High and Low silica andesites show similar patterns. This argues for a common source rock or parental magma. However, High silica andesites are enriched in the incompatible elements Sr and Ba and depleted in the compatible elements Sm and Ti relative to Low silica andesites. This is consistent with derivation of High silica andesite from a smaller volume of partial melt and/or greater fractionation from a mafic parent.

The variation in color, texture and phenocryst mineralogy along with grouping of andesites on the TAS diagram suggests two or more episodes of andesitic volcanism. The spider diagram supports this hypothesis.

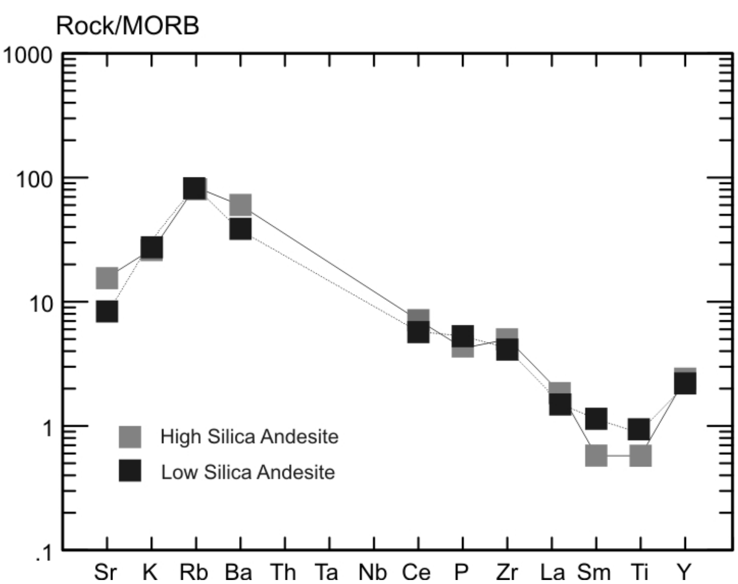

Figure 3. MORB normalized spider diagram for Low and High silica Halloran Hills andesites.

Andesite Petrogenesis

Gill (1981) argued persuasively for the importance of K2O, terming it "the significant variable in major element composition between andesites for tectonics". He believed that most andesites are produced at collisional plate boundaries and that K2O content was a measure of the extent to which continental lithosphere is involved in orogenesis. Low-K andesites are produced by limited interaction with continental lithosphere while High-K andesites indicate that continental crust played a significant role. He further stated that shoshonites were exceedingly rare in convergent settings, implying that the "shoshonite" andesites were not the products of convergence.

Figure 4 is a Gill diagram for the HH andesites. Twenty of the 22 andesite samples plot on

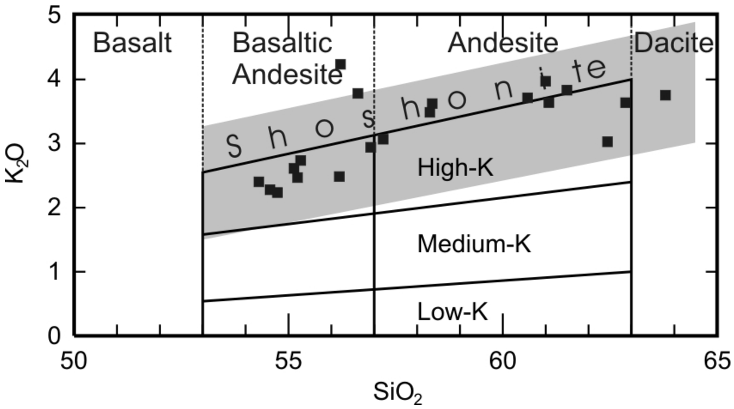

Figure 4. K2O-SiO2 diagram for the Halloran Hills andesites (Gill, 1981). Shaded rectangle represents the approximate compositional variation for the Conejos volcanics of Parker, et. al., 2005.

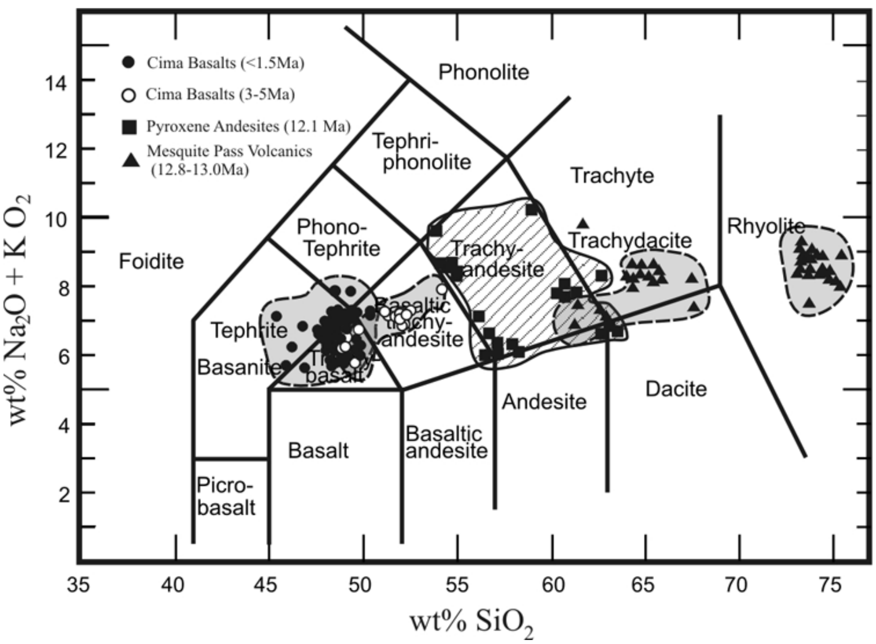

Figure 5. TAS diagram for the eastern Mojave volcanics. Dashed line approximates the path of magma evolution.

Figure 6. Basalt flows in the Halloran Hills. R.E. Reynolds photograph.

the diagram, the remainder containing greater than 5% K2O. All HH andesites lie within either the High-K or Shoshonite fields. Parker, et. al., (2005) studied the Oligocene Conejos volcanics of southern Colorado. These rocks bear remarkable petrographic and geochemical similarities to the HH andesites. Emplacement of the Conejos Volcanics was related to evolution of the Rio Grande Rift.

While it is unlikely the Halloran Hills andesites are a consequence of continental rifting, certain parallels exist. The High-K andesites and shoshonites of both the Halloran Hills and Rio Grande Rift were produced from a continental lithospheric source. Furthermore, the Conejos volcanics are a product of extension, as are the HH andesites. In the latter case, detachment rather than rifting is the driving mechanism.

Recent work by Tarman, et. al., (2002) on the petrology and geochemistry of the Mesquite Pass volcanics, 15 kilometers northeast of the Halloran Hills, has documented a suite of rhyolitic to trachydacitic rocks characterized by high alkali content. Samples yielded Ar/Ar ages of 12.8 to 13.0 Ma (Tarman, et. al., 2002). Field mapping has shown that the Mesquite Pass volcanics were emplaced at the apex of the breakaway zone for the 12.2 Ma (Davis, 1992) Kingston Range detachment fault.

Jessey, et. al., (2007) undertook an extensive sampling program in the Cima volcanic field, south of the Halloran Hills. This research demonstrated that the Cima volcanics underwent significant compositional changes from early mugearites (basaltic trachyandesite) emplaced from 7.5 to 3 Ma, to hawaiites (trachybasalt) extruded from 1.5 Ma to the Present.

Figure 5 summarizes the geochemical trends for eastern Mojave volcanics. Clearly, alkali enrichment characterizes the entire spectrum of Cenozoic rocks. The 13 Ma Mesquite Pass volcanics are of rhyolitic to trachydacitic composition, followed by the Halloran Hills trachyandesite at 12 Ma. Basaltic trachyandesite (mugearite) and trachybasalt (hawaiite) of the Cima volcanic field represent the most recent periods of alkali magmatism.

Discussion and Conclusions

Cenozoic volcanism in the eastern Mojave has evolved over time. The earliest volcanic rocks were rhyolites and trachydacites, followed closely (less than 1 million years) by a period of trachyandesitic magmatism. After a gap of three to five million years, basaltic trachyandesite (mugearite) and trachybasalt (hawaiite) were erupted. The older basalts (7.5-3 Ma) were dominantly mugearites; and the most recent activity (<1.5 Ma) hawaiite. This defines a trend of decreasing silica content (75% to 48%) accompanied by a modest decrease in total alkalis (about 2%).

Greg Davis (1992) speculated that volcanic rocks outcropping in the Mesquite Pass area might be related to the Kingston Range detachment fault. Ar/Ar geochronology has demonstrated that ages for the Mesquite Pass volcanics are compatible with the hypothesized age (≈12 Ma) for detachment. The heat source generating the melts is problematic. One theory suggests that frictional energy associated with large scale crustal dislocations. A second model postulates crustal extension and thinning that results in upwelling of asthenospheric mantle; the heat from the mantle generating partial melts (Tarman, et. al., 2001).

The high-K andesites of the Halloran Hills have chemical similarities to rocks from the Rio Grande rift (Conejos Volcanics, Parker, et. al., 2005). It is proposed that the Halloran Hills andesites, like those of the Rio Grande rift, are a product of crustal extension. Their postulated age (12.1 Ma) is close to that for the Kingston Range detachment, and they lie in close proximity to the Mesquite Pass volcanics. As extension thinned the crust, deeper partial melts generated from more mafic portions of the lower crust were able to reach the surface. As these magmas rose they interacted with the crust, digesting upper crustal rocks and increasing alkali content while adding crustal contaminants.

The most recent episode of basaltic volcanism began at 7.5 Ma. It resulted in the early mugearites followed by the more recent (<1.5 Ma) hawaiites. The role of extension in the emplacement of Cima basalts is more enigmatic. Emplacement begins after a hiatus of 3-4 million years and is coincident with a transformation from extension to dextral shear (Reynolds, 1996). Perhaps, shearing generated faults that penetrated the thinned lithospheric crust enabling mantle-derived magmas to ascend to the surface.

Farmer, et. al., (1995) stated that the locus of volcanism in the eastern Mojave has moved little during the past 12 million years and that there is no relationship to a mantle hot spot. However, if one assumes a genetic relationship between the volcanics of the Cima field, Halloran Hills and Mesquite Pass then a fixed heat source is more likely. If so, a pattern of clockwise rotation or northeastward migration relative to a fixed heat source at a rate of approximately 5 mm/yr would result. Furthermore, this motion appears to have two distinct components; an older phase (13 - 10 Ma) of dominantly east-west motion, and a younger (7.5 Ma - Present) period of nearly north-south movement; the former related to detachment and the latter, dextral shear.

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Excerpt from:

Stratigraphic evidence for the role of lake-spillover in the inception of the lower Colorado River in southern Nevada

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Abstract

The stratigraphy of Late Miocene and early Pliocene sediments exposed along the lower Colorado River near Laughlin, Nevada contains evidence that the establishment of this reach of the river after 5.6 Ma involved flooding from lake spillover through a bedrock divide between Cottonwood Valley to the north and Mohave Valley to the south. An early phase of intermittent lacustrine inundation restricted to the Cottonwood Valley is recorded by lacustrine marl interfingered with and overlying an 80-100 m thick sequence of fine-grained valley-fill deposits. Outcrops of limestone, mud, sand, and minor gravel of the Bouse Formation at elevations up to 550 m a.s.l. record a subsequent interval of deep aqueous inundation in both valleys. A coarse-grained and lithologically distinct fluvial conglomerate at the north end of the Mohave Valley separates subaerial, locally derived fluvial deposits from subaqueous deposits of the Bouse Formation. We interpret this key unit as evidence for the breaching of the paleodivide immediately before deep lacustrine inundation. Sedimentary strata in the upstream valley host a corresponding erosional unconformity. The second lacustrine phase was followed by basin drainage and the arrival of the through-going, sediment-laden Colorado River. Voluminous river aggradation filled the valleys with more than 300 m of alluvium and culminated in the Pliocene between 4.1 and 3.3 Ma. The stratigraphic associations and timing of this major transition in regional drainage are consistent with geochemical evidence linking lacustrine conditions to the early Colorado River, the timings of drainage integration and canyon incision on the Colorado Plateau, the arrival of the river at its terminus in the Salton Trough, and with a mode of river integration that is typical in areas of crustal extension.

Introduction and background

The development of the lower Colorado River is closely linked with incision of Grand Canyon, and both topics have been debated for most of the past century (e.g., Blackwelder, 1934; Longwell, 1936, 1947; Hunt, 1969; Lucchitta, 1972, 1979; Spencer and Patchett., 1997; Lucchitta et al., 2001). The Colorado River drains a large interior highland (including the Colorado Plateau and large parts of the southern and central Rocky Mountains) then passes through the arid lowlands of the southern Basin and Range province before emptying into the Gulf of California (Fig. 1). The transition between highland and lowland reaches is Grand Canyon, where the river is incised more than 1000 m below the surrounding plateau. Downstream from Grand Canyon, the Colorado River follows a peculiar course through the rugged terrain of the Basin and Range province. The river's course initially tracks westward through a series of alternating alluvial basins and bedrock canyons in the Lake Mead area, and then abruptly turns south through more canyons and basins before reaching the Gulf of California.

Models for the inception and development of the lower Colorado River

Three primary mechanisms have been proposed to explain the existence of and the particular features associated with the lower course of the Colorado River: (1) antecedence; (2) combinations of regional subsidence, marine incursion, headward erosion, drainage capture, and river progradation; and (3) downstream integration via lake-spillover. The simplest explanation is antecedence, in which the river existed along its current course before uplift of the Colorado Plateau and adjacent areas and simply incised in response to the uplift (Powell, 1875; Dutton, 1882; Blackwelder, 1934). Abundant evidence that the lower Colorado River did not follow its present course before 6 Ma rules this possibility out, however (e.g., Lucchitta, 1979).

Fundamental differences in the latter two explanations stem primarily from alternate interpretations of the depositional environment of the Bouse Formation, a distinctive suite of sediments that occupy a key stratigraphic position between pre-integration and post-integration deposits in a string of alluvial basins between Hoover Dam and the

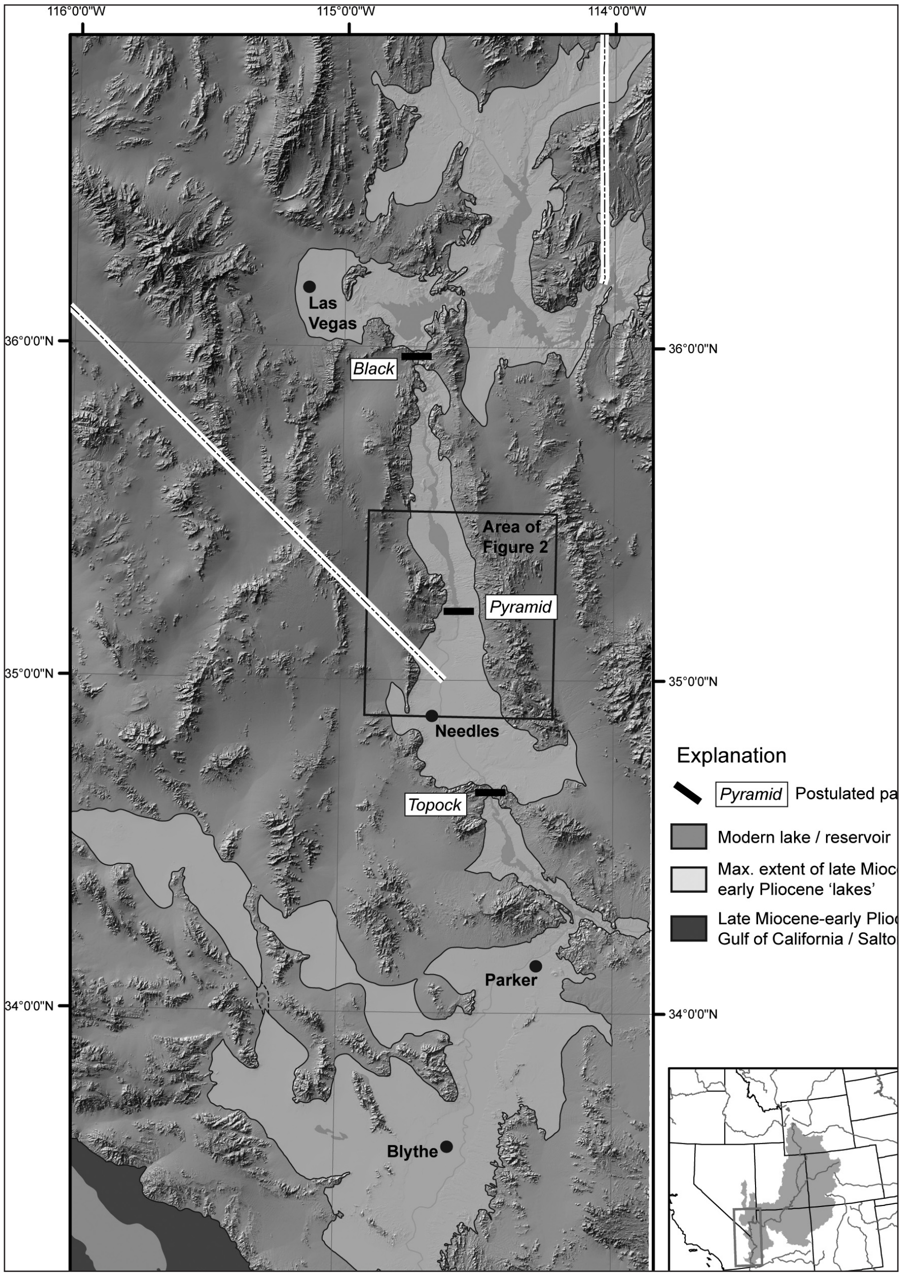

Figure 1. Overview of the lower Colorado River region from the western edge of Grand Canyon to the Salton Trough. Maximum extent of late Miocene lakes indicated on basis of extrapolations of highest outcrops of Hualapai Limestone and Bouse Formation in relation to modern topography. Inset map shows the extent of the Colorado River Basin north of the confluence of the Colorado and Gila Rivers near Yuma, Arizona.

Salton Trough (Fig. 1). The Bouse Formation includes a sequence of basal limestone and associated tufa overlain by mud, sand, and minor gravel of varying thicknesses (Metzger, 1968; Buising, 1990). It has been interpreted as a marine-estuarine deposit because it contains a limited assemblage of marine fossils in exposures as far north as Parker, Arizona (Metzger, 1968; Smith, 1970; McDougall, 2005). In contrast, strontium isotope ratios in Bouse carbonates throughout the extent of the deposit are more similar to modern Colorado River water than to sea water, which supports the hypothesis that Bouse sediments were deposited into a lake fed by the Colorado River (Spencer and Patchett, 1997; Poulson and John, 2003).

Critical evaluation of the competing models has been hindered by a lack of stratigraphic evidence documenting the geologic circumstances and timing of the arrival of the Colorado River into the region. In this paper, we report on new stratigraphic and tephrochronologic evidence from Mohave and Cottonwood Valleys (Fig. 2) that is consistent with the lake-spillover model. Our evidence comes from geologic mapping of late Miocene and early Pliocene sediments on both sides of an inferred bedrock paleodivide that would have separated the valleys. Stratigraphy of the deposits in each valley records a concurrent series of changes in depositional conditions that are consonant with short-lived lacustrine inundation in the upstream valley, flooding through the valley-separating divide, followed by an episode of deeper lacustrine inundation of both valleys. Tephrochronologic data indicates that these events occurred after 5.6 Ma. The deep lacustrine episode was followed by a period of erosion and then thick aggradation of sand and gravel associated with the Colorado River. Additional tephrochronologic data indicate that river aggradation culminated soon after 4.1 Ma and that net downcutting has dominated the history of the river since at least 3.3 Ma.

Study area

Cottonwood and Mohave Valleys are elongate, north-trending alluvial valleys along the lower Colorado River that lie roughly halfway between the mouth of Grand Canyon and the Gulf of California. The valleys are structural basins produced by major crustal extension in the middle Miocene (Howard and John, 1987; Spencer and Reynolds, 1989; Faulds et al., 1990). Our primary study area is approximately the southern half of Cottonwood Valley and the northern third of Mohave Valley. Light-colored Tertiary granitic rocks in the lower plate of a major detachment fault form the bulk of the Newberry Mountains on the west side of the study area, whereas upper-plate Tertiary volcanic and minor sedimentary rocks predominate in the Black Mountains on the east side. Alluvial deposits derived from each side of the valleys reflect this bedrock

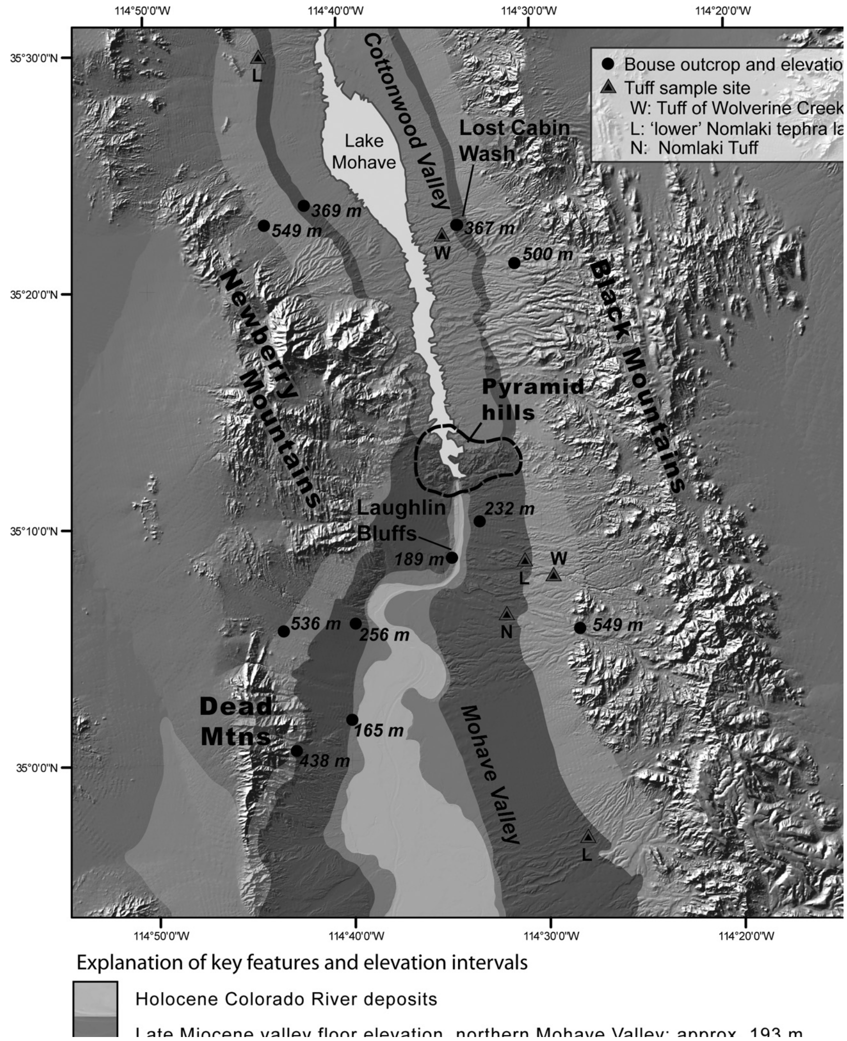

Figure 2. The study area in southern Cottonwood Valley and northern Mohave Valley showing tephra sample locations and selected Bouse Formation outcrops that indicate the vertical range of the unit in both valleys with corresponding elevations in meters noted. Relief base is not a geologic map, but is instead color-coded to indicate general geologic and topographic relations associated with pre-Bouse valley geometries, maximum Bouse water levels, and maximum extent of Pliocene Colorado River deposits.

source dichotomy and help constrain pre-river valley geometry. The Pyramid hills (informal name) form the boundary between Cottonwood and Mohave valleys and are composed almost entirely of Proterozoic megacrystic granite (Faulds et al., 2004). The Colorado River has carved Pyramid Canyon through this divide. Topock Gorge, another rugged bedrock canyon, defines the southern margin of Mohave Valley.

Key stratigraphic relations

Correlative deposits from each side of the Pyramid hills paleodivide document changes in depositional conditions associated with the arrival and early development of the Colorado River (Fig. 3). Important late Miocene to early Pliocene stratigraphic units exposed within about 20 km of the Pyramid hills paleodivide, from oldest to youngest, include: (1) post-extensional alluvial fan deposits derived from the valley-bounding mountains (Tfn and Tfb1; Newberry and Black Mountain fanglomerate); (2) axial valley facies including fine gravel from the Black and Newberry Mountains interfingered with valley-filling silt and sand deposits (Tlcc and Tlcf; the Lost Cabin beds, found only in Cottonwood Valley); (3) gravelly alluvial fills in paleochannels cut in fanglomerate (Tng; Newberry channel gravel, found only in northern Mohave Valley; (4) coarse axial valley deposits dominated by clasts of Precambrian granite (Pyramid gravel, in northern Mohave Valley only); (5) fine-grained deposits of the Bouse Formation typically including a thin basal limestone, locally grading upward into clay, silt and sand beds (Tbl and Tbms; found in both valleys); (6) elaborately cross-stratified early Colorado River deposits of medium to coarse sand and gravel with abundant exotic, well-rounded clasts (Tcb; Bullhead alluvium, both valleys; generally correlative to unit B of Metzger and Loeltz, 1973); (7) local fan deposits interfingered with unit Tcb (Tfb2, both valleys); and (8) local fan deposits deposited above erosional surfaces cut across all older units (QTf, both valleys). We have identified three different tephra layers in this sequence that provide new temporal constraints on the development of the river in this reach. They include the 5.6 Ma tuff of Wolverine Creek; the 4.1 Ma lower Nomlaki tephra layer (informal name); and the 3.3 Ma Nomlaki Tuff.

The Cottonwood Valley section

Late Miocene sedimentary strata in southern Cottonwood Valley (Fig. 3) suggest that deposition was in an enclosed basin with large alluvial fan complexes extending to the valley axis from the Newberry and Black Mountains (units Tfn and Tfb1, Figure 3). Exposures in the area of Lost Cabin Wash along the eastern shore of Lake Mohave reveal indurated, tilted fanglomerate deposits (Tft, Fig. 3) that date to the period of active normal faulting in the middle Miocene (Spencer and Reynolds, 1989). Gently eastward-dipping fanglomerate dominated by granitic clasts derived from the Newberry Mountains unconformably overlies the tilted deposits at about 240 m above sea level (a.s.l.). This Newberry fanglomerate grades into flat-lying axial valley gravel deposits containing mixed clasts from the Newberry and Black Mountains (coarse-grained facies of the Lost Cabin beds, Tlcc). The axial gravel deposits grade upward into a sequence of flat-lying sandstone, siltstone and mudstone beds with minor gravel (fine-grained facies of the Lost Cabin beds, Tlcf; Fig. 4a).

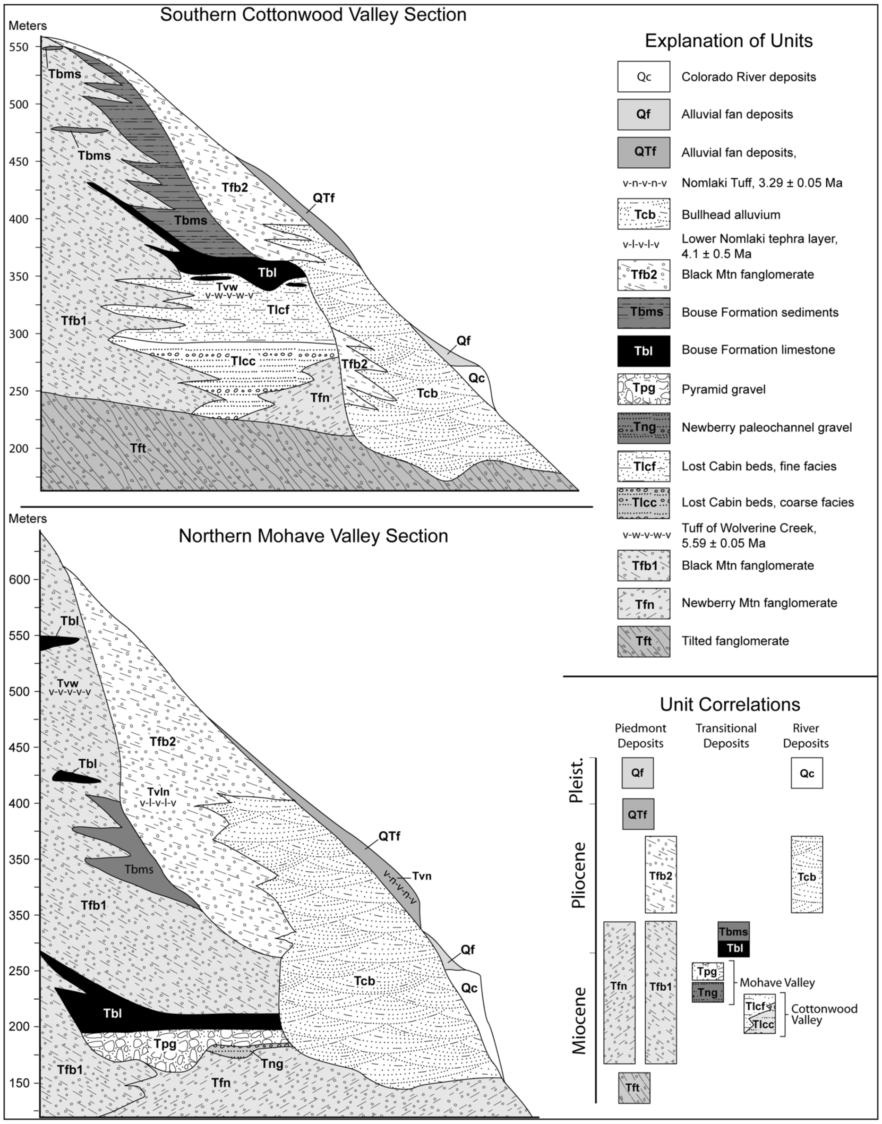

Figure 3. Schematic diagrams of key late Cenozoic stratigraphic relations in Cottonwood and Mohave Valleys. Vertical exaggeration approximately 10x. Upper section: Southern Cottonwood Valley. Note that the contact between the Lost Cabin beds (Tlc) and the Bouse limestone (Tbl) indicates the approximate valley axis elevation prior to river integration; Lower section: Northern Mohave Valley. Note that the contact between the Newberry paleochannel gravel and the Pyramid gravel indicates the approximate valley axis elevation prior to river integration. Unit explanations and correlation diagram on right side of figure apply to both sections. Transitional units exclusive to each valley indicated on correlation diagram.

The Lost Cabin beds grade laterally into local fanglomerate, and also contain several weak to moderately developed paleosols (Fig. 4b). We infer that these stratigraphic relations preclude the presence of a through-going Colorado River. Our interpretation is that the valley was an enclosed basin during the deposition of the Lost Cabin beds and the axial drainage in southern Cottonwood Valley fed a depocenter to the north, the direction in which the valley widens and the fine-grained Lost Cabin Beds thicken. The 5.59 ± 0.05 Ma tuff of Wolverine Creek (Fig. 4c) is in the upper third of the fine Lost Cabin beds.

Fine Lost Cabin beds typically are overlain by Black Mountain fanglomerate (Tfb2 and upper part of Tfb1) along an erosional unconformity at an elevation of 350 m a.s.l. or less. In some locations, however, thin beds of Bouse limestone and calcareous mud interfinger with the upper few meters of the Lost Cabin beds (Fig. 5a). Thick and extensive Bouse limestone, sand, and mud deposits overlie a minor, locally erosional unconformity above this key interval (Fig. 5b). In places, the unconformable contact is characterized by small channels filled with local gravel and reworked mud (Fig. 5c). This stratigraphic relationship suggests a phase of intermittent lacustrine and subaerial sedimentation separated from a more extensive and prolonged period of lacustrine deposition by a period of erosion. At higher elevations to the east, Bouse limestone rests on gently west-dipping fan paleosurfaces underlain by weathered Black Mountain fanglomerate. In a few localities, up to 10 m of mudstone, siltstone, and sandstone are preserved above the basal limestone (Fig. 5d). We have found Bouse limestone, tufa, and related clastic shoreline sediments at elevations up to 550 m a.s.l. in central Cottonwood Valley, indicating a local water depth of at least 200 m (Fig. 2). The upper Bouse contact typically is erosional and Bouse deposits and some underlying Lost Cabin beds were removed prior to renewed fanglomerate deposition.

Early Colorado River sand and gravel deposits (Bullhead alluvium) consisting of a mix of well-rounded, exotic gravel and subangular local gravel rest on an erosional unconformity carved deeply into all of the older deposits. In southern Cottonwood Valley, the lowest exposed outcrops of the Bullhead alluvium are at the level of Lake Mohave (approximately 195 m a.s.l.), and they have been found as high as 330 m in the valley. Gently west-dipping Black Mountain gravel deposits (QTf) with massive duripans up to 3 m thick cap the highest ridges on the piedmont. These piedmont deposits rest unconformably on older Black Mountain fanglomerate (Tfb) and Bullhead alluvium and were graded to a paleo-river level of about 300 to 330 m a.s.l.

The Northern Mohave Valley section

Stratigraphic relations in northern Mohave Valley (Fig. 3) also record a late Miocene to early Pliocene transition from local drainage to deep inundation to through-going Colorado River, and they comprise an interesting comple-

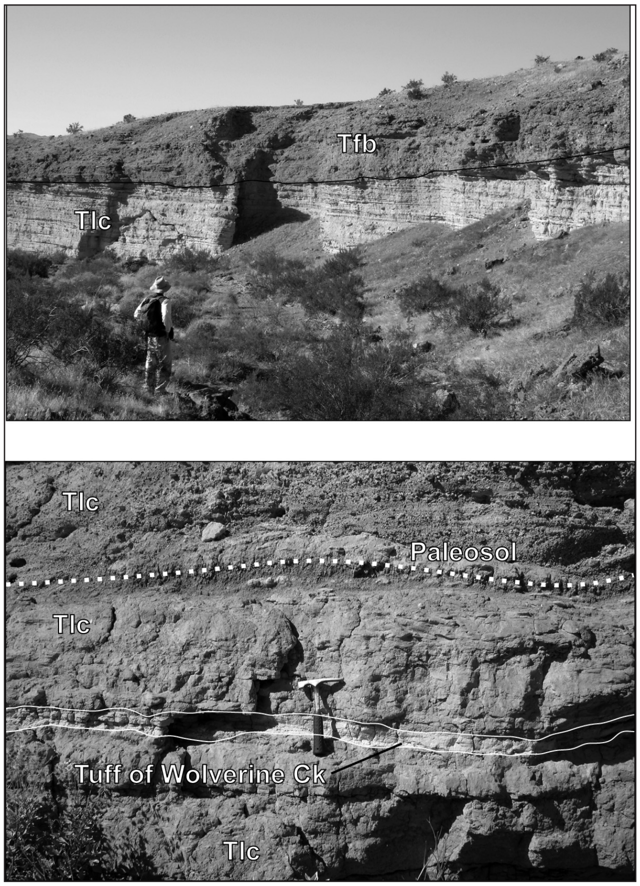

Figure 4. Photographs of the Lost Cabin beds in the Lost Cabin Wash area. A. Thick section of fine-grained Lost Cabin beds (Tlc) overlain by Black Mountain fanglomerate (Tfb); B. Paleosol in the Lost Cabin beds overlying the tuff of Wolverine Creek. Note rock hammer for scale. Site is in Odyssey wash (informal name) in vicinity of Lost Cabin Wash.

ment to the Cottonwood Valley sequence. Conspicuous differences between the late Miocene to early Pliocene strata in each valley exist and are interpreted to have resulted from their locations relative to the inferred paleodivide between the valleys. The lowest strata exposed in northern Mohave Valley consists of local fanglomerate, which at various locations contains gravel primarily derived from the Newberry Mountains, the Black Mountains, or the Pyramid hills (Faulds et al., 2004). This is consistent with alluvial fan deposition from the western, eastern, and northern flanks of the valley and drainage to a depocenter to the south.

A particularly illustrative example of the northern Mohave Valley section is exposed in the bluffs astride the

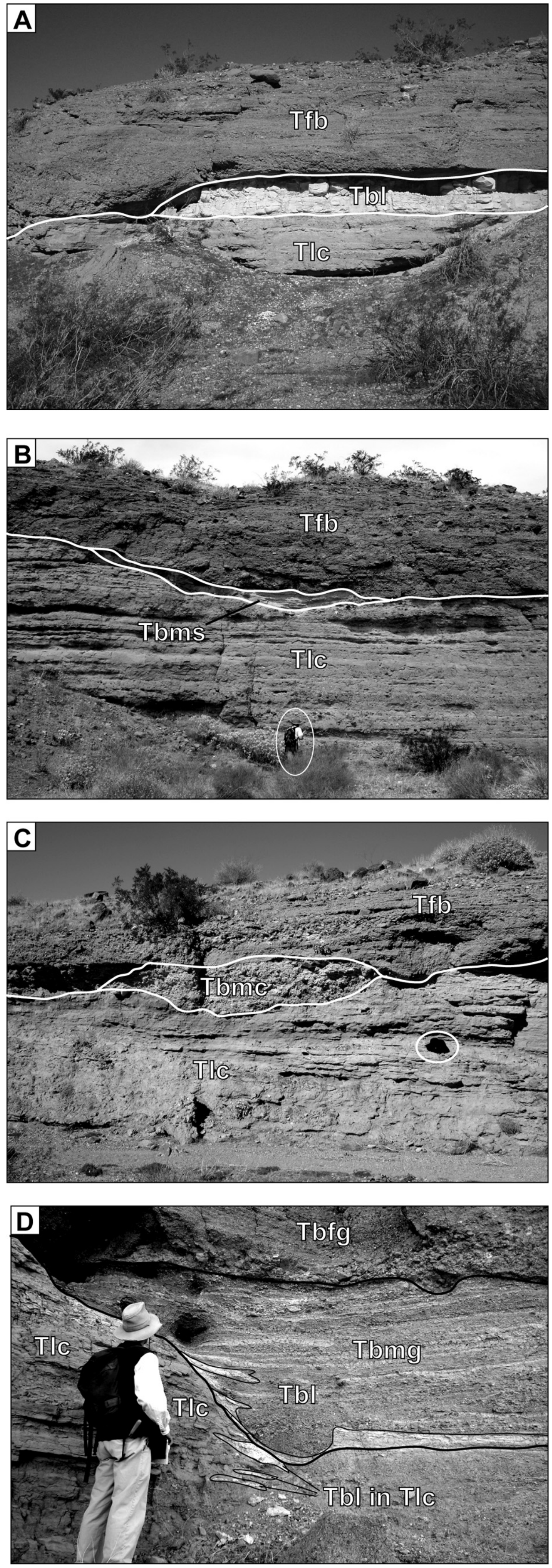

Figure 5. Stratigraphic relations between the Lost Cabin beds and the Bouse Formation. A. Coarse lost Cabin beds (Tlc) overlain by Bouse limestone (Tbl), which is, in turn, unconformably overlain by Black Mountain fanglomerate (Tfb). Thickness of Tlc in photo approximately 1 meter. B. Paleochannel at top of Lost Cabin beds (Tlc) filled with calcareous Bouse mud and sand (Tbms). Note person for scale; C. Paleochannel in Lost Cabin beds (Tlc) filled with pebbly Bouse mud-ball conglomerate (Tbmc). Note hat (circled) for scale; D. Contact between Lost Cabin beds (Tlc) with interfingered marl (Tbl) overlain along erosional unconformity by muddy gravel (Tbmg) with interbedded marl (Tbl). Disconformable upper beds are possible wave-worked fanglomerate (Tbfg).

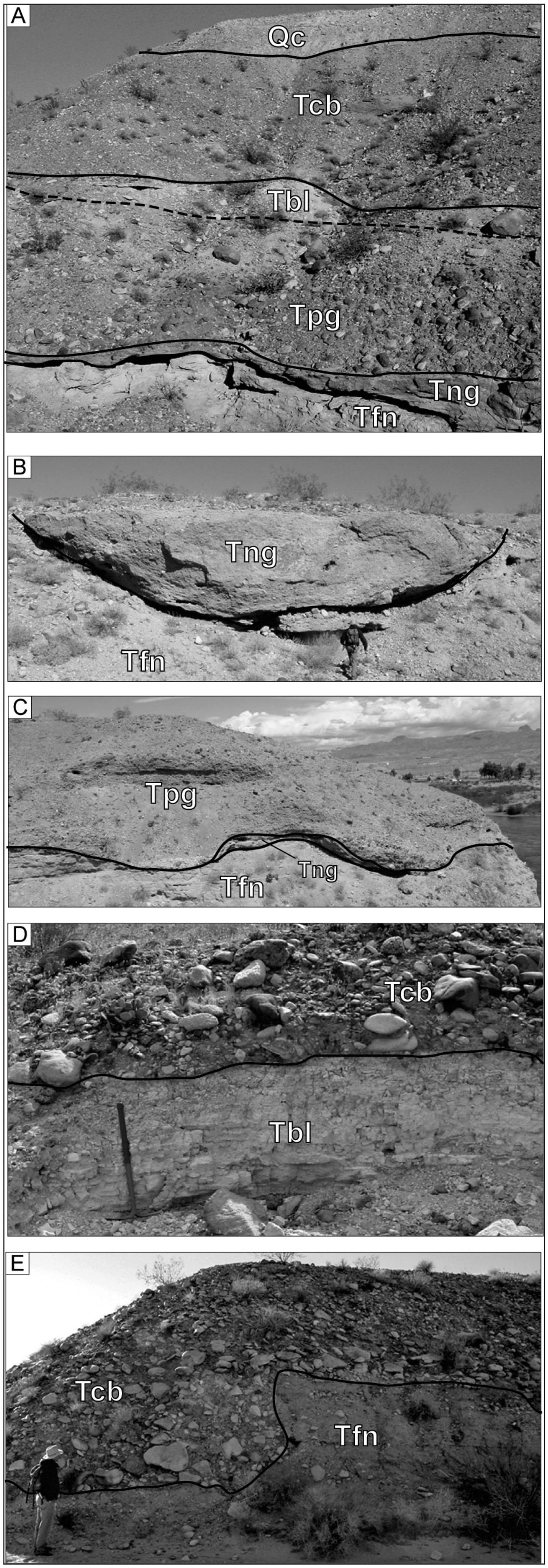

Figure 6. Photographs of the Laughlin Bluffs section. A. Complete section (approx. 10 m thick) showing stratigraphic relation between each key unit (Tfn: Newberry Mtn fanglomerate; Tng: Newberry paleochannel gravel; Tpg: Pyramid gravel; Tbl: Bouse limestone; Tcb: Bullhead alluvium; B. Newberry gravel (Tng) filling paleochannel in fanglomerate (Tfn); C. Pyramid gravel flood deposit (Tpg) filling paleochannels in Newberry fanglomerate (Tfn); D. Bouse marl (Tbl) disconformably overlain by Bullhead alluvium (Tcb); E. Base of the Bullhead alluvium incised into Newberry fanglomerate.

Colorado River south of Laughlin, Nevada (Figs. 2 and 6a). There, a deposit of cross-stratified, locally-derived gravel (Newberry paleochannel gravel) fills relatively small, roughly south-trending paleochannels cut into Newberry fanglomerate (Fig. 6b). The fanglomerate and paleochannels are overlain along an erosional unconformity by a distinctive coarse-grained fluvial conglomerate with a composition dominated by clasts of dark-colored megacrystic granite with lesser amounts of gravel eroded from local fanglomerate deposits (Fig. 6c). We have not identified any diagnostic Colorado River sediments in this gravel unit. The nearest source of the dominant lithology (megacrystic granite) is in the Pyramid hills paleodivide about 7 km to the north, and we have informally named the unit the Pyramid gravel. The Pyramid gravel is a broadly tabular deposit up to 20 m thick, is crudely to moderately stratified, and has clast-supported and matrix-supported beds. Clast imbrication and trough cross-stratification are evident in several channel-filling exposures. Boulders up to 1 m in intermediate axis diameter are common in the lower part of the deposit, but it typically is a pebble and cobble conglomerate. The basal limestone of the Bouse Formation

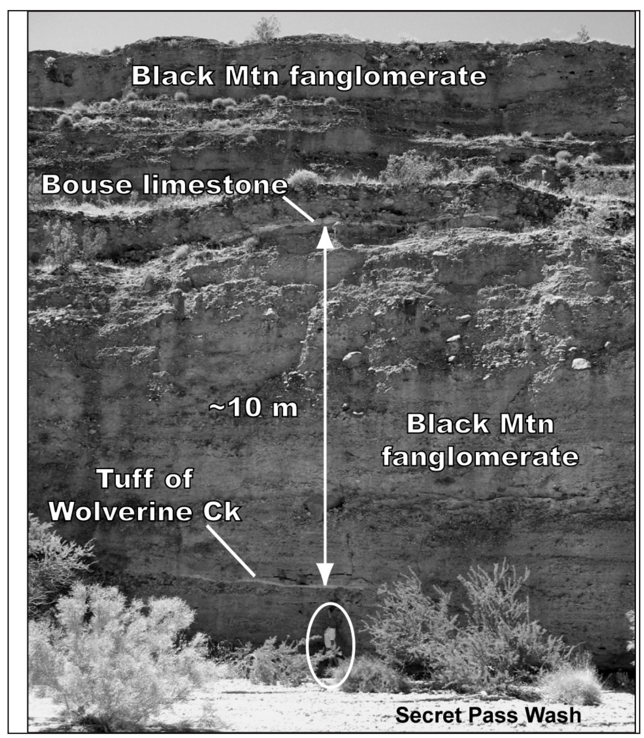

Figure 7. Photograph of the tuff of Wolverine Creek and overlying, thin bed of Bouse limestone in Secret Pass Wash (modified from House et al., 2005a). Person indicated in lower part of photo for scale.

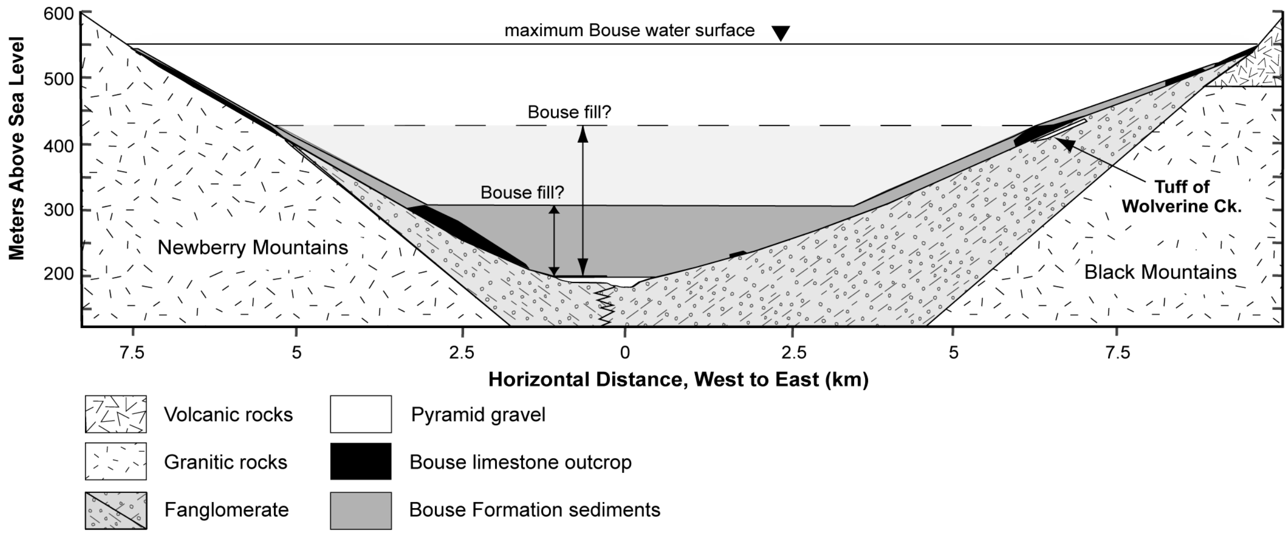

Figure 8. Schematic diagram showing distribution of Bouse limestone, tufa, and clastic sediments in Mohave Valley and speculations on extent of Bouse fill. Extant outcrops of Bouse formation are shown in black (adapted from House et al., 2005a).

overlies the Pyramid gravel in the Laughlin bluff section along a sharp, relatively flat contact. This stratigraphic pairing indicates an abrupt change from high- to low-energy depositional conditions (Fig. 6d). The entire foregoing sequence is overlain along a high-relief erosional unconformity by the Bullhead alluvium (Fig. 6e).

At other locales in Mohave Valley, basal Bouse deposits overlie paleo-alluvial fan surfaces and bedrock slopes and are interbedded with fanglomerate deposits up to elevations of 550 m a.s.l. (Fig. 7). High on the east side of the valley, the tuff of Wolverine Creek is interbedded with fanglomerate deposits 10 m below the Bouse limestone, providing a maximum age constraint of 5.6 Ma for deep inundation of Mohave Valley (Fig. 7). The basal limestone beds are everywhere less than a few meters thick, but are locally overlain by up to 30 m of mud and sand. The distribution of outcrops of the Bouse Formation suggests that water depth in northern Mohave Valley may have exceeded 400 m, and the general form of the valley at that time was quite similar to its present form (Fig. 8).

As in Cottonwood Valley, a thick deposit of Bullhead alluvium rests on an unconformity that cuts across all older basin deposits in Mohave Valley. The lowest exposures of Bullhead alluvium in both valleys contain abundant locally derived gravel mixed with well-rounded quartzite, chert, and other exotic pebbles and small cobbles (Figs. 9a and 9b). Exposures of Bullhead alluvium higher in the sec-

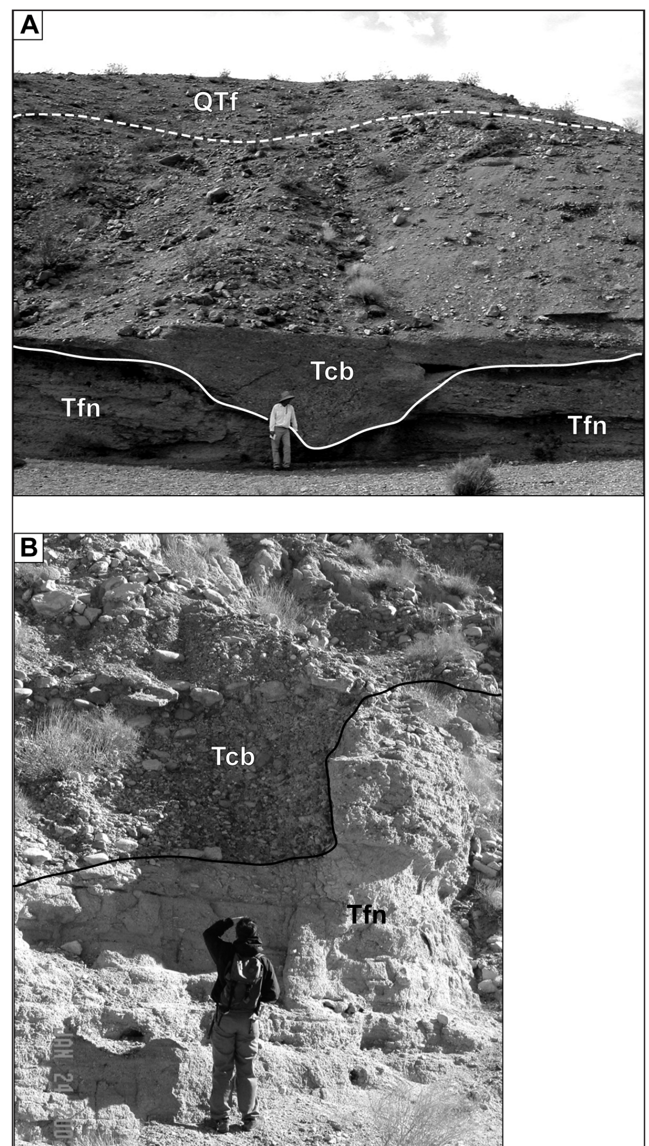

Figure 9. Examples of the lower, erosive contact at the base of Bullhead alluvium in Cottonwood and Mohave Valleys. A. Paleochannel in Granite Wash, Cottonwood Valley, approximately 25 meters above Lake Mohave surface (195 m at mouth of wash). Deposit is poorly sorted, massive to cross-stratified pebbly conglomerate with mix of dominantly local gravel and sparse, exotic, well-rounded pea gravel. Bullhead paleochannel is cut into east-dipping Newberry Mountain fanglomerate; B. Base of gravelly Bullhead alluvium incised in Newberry fanglomerate in the Laughlin bluffs area (Panda gulch of House et al., 2005a). Note conspicuous contact. Deposit is mix of locally derived coarse, subrounded gravel and exotic, well-rounded pea gravel and cobbles. Lowest exposure of Bullhead alluvium in Panda gulch is 18 m above modern Colorado River level (152 m at mouth of gulch).

tion on the Black Mountain piedmont in Mohave and Cottonwood valleys contain laterally extensive, tabular beds of trough cross-stratified gravel, thick beds of complexly cross-stratified sand, and minor flat-lying mud. Overall, the Bullhead alluvium is a very complex fluvial deposit with numerous stratigraphic discontinuities (e.g., Fig. 12 in Metzger and Loeltz, 1973); however, to date we have not recognized paleosols or erosion surfaces suggestive of major hiatuses or cut-and-fill episodes. Bullhead alluvium is extensively interstratified with piedmont fanglomerate deposits, particularly in the upper part of the section. Two outcrops of the 4.1 ± 0.5 Ma lower Nomlaki tephra layer rest in fine tributary fan gravel at elevations of 365 to 390 m a.s.l., but Bullhead alluvium is found at stratigraphically higher positions up to 400 m a.s.l. in the immediate vicinity of each tephra layer exposure in Mohave Valley (Fig. 10). Thus, the lower Nomlaki tephra is very near the top of the Bullhead aggradational sequence.

Relatively thin, west-dipping Black Mountain piedmont gravels with very strongly developed duripans (Herriman and Hendricks, 1984) lie above an erosion surface cut on Bullhead alluvium and Black Mountain fanglomerate. The 3.3 Ma Nomlaki Tuff is intercalated in these piedmont deposits at elevations of 395 and 350 m a.s.l., so they must have been graded to a former, post-Bullhead level of the Colorado River below 350 m a.s.l. In summary, the tephrochronologic and stratigraphic evidence support an interpretation that the Bullhead aggradation event in the study area culminated sometime around ~4.1 Ma and incision into the deposit by at least 50 m had occurred by ~3.3 Ma.

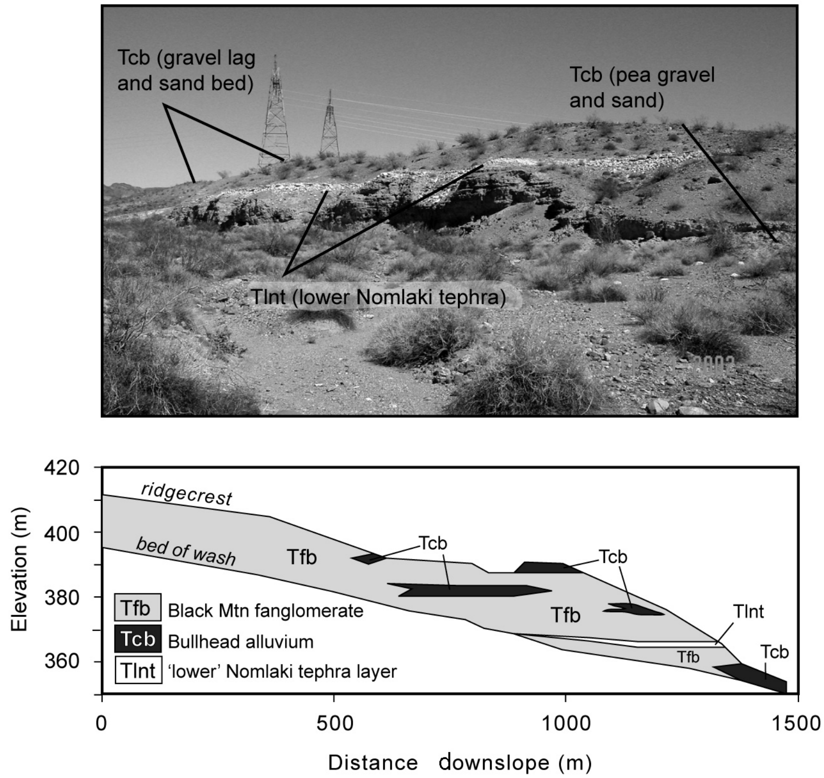

Figure 10. The 'lower' Nomlaki tephra layer interfingered with Black Mountain fanglomerate and Bullhead alluvium, Powerline outcrop, Mohave Valley, Arizona. Upper image is photo of tephra layer outcrop in fanglomerate with strata of Bullhead sediment and surface gravel lags indicated; Lower image is a cross-section of the same site showing interpreted stratigraphic relations.

The development of the Colorado River in Cottonwood and Mohave Valleys

Any proposed scenario for the development of the Colorado River in this region must accommodate several key stratigraphic relationships and temporal constraints in Mohave and Cottonwood Valleys. No through-going regional drainage existed in these valleys before 5.6 Ma. In Cottonwood Valley, an interval of fine-grained clastic deposition by local streams was ongoing by 5.6 Ma, and terminated with an erosional interval followed by quiet-water limestone deposition. There is strong evidence for a southward-directed drainage divide failure between the valleys before deep inundation of both valleys. A period of erosion followed the deep inundation and preceded the arrival of voluminous coarse Colorado River sediment. After its arrival, the river aggraded dramatically until shortly after 4.1 Ma. The river incised well below its level of maximum aggradation by 3.3 Ma, and has remained far below that level to the present. The entire series of transitional events from enclosed basins to deep-water inundation followed by the arrival of the Colorado River and thick aggradation, and finally the initial river incision appears to have occurred in less than 2 million years.

We propose the following scenario for the early development of the Colorado River based on interpretation of the evidence that we have compiled from this area (principal events summarized in Fig. 11):

Through much of the late Miocene, the Cottonwood and Mohave valleys had separate, closed drainage systems (Fig. 11a) that were relicts of middle Miocene extensional faulting as shown by local fanglomerates, axial valley gravel deposits, and inferred playa deposits north and south of the primary study area.

Exotic water and sediment, possibly associated with the Colorado River, began to enter Cottonwood Valley from the north in the late Miocene (Fig. 11b). This resulted in the expansion of fine-grained, primarily subaerial deposition into southern Cottonwood Valley from the north. This interval of deposition was ongoing by 5.6 Ma as shown and dated by the Lost Cabin beds and the tuff of Wolverine Creek.

The influx of water eventually formed a lake in Cottonwood Valley (Fig. 11c), represented by thin beds of limestone interfingered with upper Lost Cabin beds.

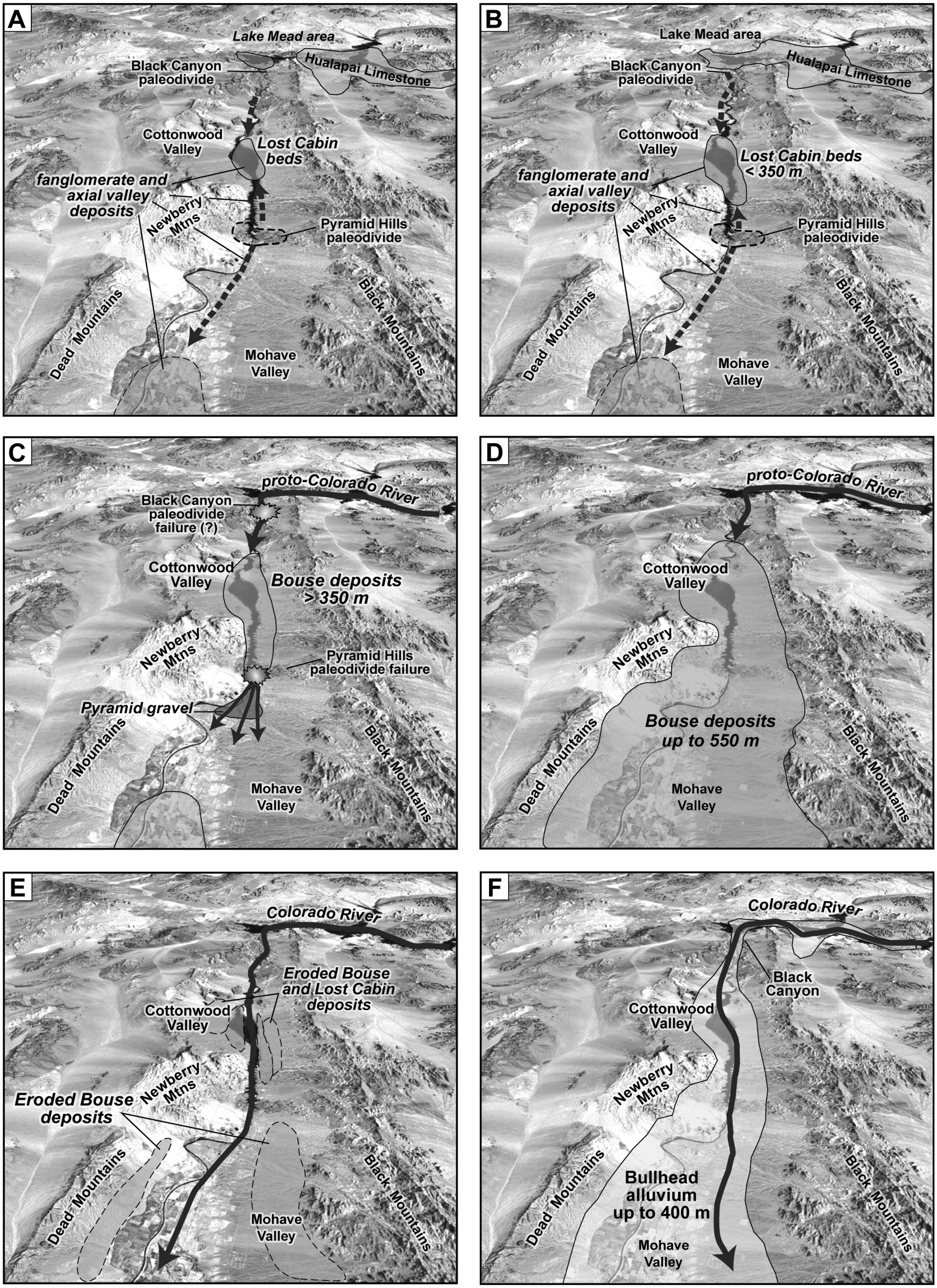

Figure 11. Proposed stages in the evolution of the Lower Colorado River based on evidence and interpretations presented in this paper. Base map provided courtesy of William A. Bowen. See text for discussion of each panel.

Water began to flow over or through extensively fractured bedrock in the Pyramid hills, as indicated by the Newberry paleochannel gravel. Eventually, the divide was completely breached as shown by the Pyramid gravel (Fig. 11c). In Cottonwood Valley, this event is recorded as an erosional interval separating the Lost Cabin beds and the bulk of the Bouse Formation.

Persistently inflowing Colorado River water was dammed, possibly at the south end of Mohave Valley, resulting in deep inundation of both Mohave and Cottonwood Valleys (Fig. 11d). The maximum water surface elevation in both valleys was ~550 m a.s.l. as inferred from the highest outcrops of Bouse limestone, sandstone, and tufa in both valleys. During this time, concurrent influx of fine sediment and reworking of local sediments resulted in deposition of clastic sediments of the Bouse formation, possibly in the form of a delta extending from northern Cottonwood Valley.

The paleodivide to the south was ultimately overtopped and eroded in some manner, resulting in lake drainage and erosion of Bouse and older deposits (Fig. 11e).

Voluminous Colorado River bedload sediment arrived into subaerially exposed valleys (Fig. 11e) and was first deposited with abundant locally derived sediments at levels near the modern river about 150-200 m a.s.l. This is indicated by the lower part of the Bullhead alluvium and its contact with underlying fanglomerates.

Massive, presumably integration-driven aggradation of the Colorado River and tributary fanglomerates filled Mohave and Cottonwood Valleys with sediment to a level of about 400 m a.s.l. (Fig. 11f). This aggradational interval culminated shortly after 4.1 Ma as indicated by stratigraphic relations among the Bullhead alluvium, local fanglomerate, and the lower Nomlaki tephra layer.

Incision of the Colorado River below the maximum level of aggradation began before 3.3 Ma (Black Mountain piedmont gravel, Nomlaki Tuff) and ultimately continued to a level near modern river grade, resulting in the removal of vast amounts of Bullhead alluvium and local fanglomerate (not specifically shown on Fig. 11, but similar to Fig. 11e).

Discussion

The postulated sequence of events outlined above is consistent with recent geochemical studies that have tied the development of the lower Colorado River to lacustrine deposits elsewhere along the river's course (Spencer and Patchett, 1997; Poulson and John, 2003). It is also consistent with stratigraphic and geochronologic evidence for development of the Colorado River in areas upstream (e.g., Howard and Bohannon, 2001) and downstream (e.g., Buising, 1990; Dorsey et al., 2005) as described below.

Bouse associations between Laughlin, Nevada and Parker, Arizona

Deposits of the Bouse Formation are relatively extensive along the lower Colorado River downstream from Mohave Valley, and they have stratigraphic and geographic characteristics that can be reconciled with a downstream directed lake-spilling model of integration. For example, in the Parker, Arizona area Buising (1988, 1990) described a stratigraphic relation in which a distinctive cross-stratified deposit of fluvial gravel that is interfingered with the base of the Bouse Formation overlies locally derived fanglomerate. The cross-stratified gravels contain far traveled sediment and have sedimentary structures indicative of generally southward transport (Buising, 1990). These relations suggest that an influx of water from the north preceded and accompanied basal Bouse deposition in this area. Buising (1990) interpreted this association as evidence for a fluvial-marine interface (estuary) at this location; however, it is also consistent with a fluvial-lacustrine interface that is roughly similar to, but more well-developed than, what we have described in the Laughlin area.

If the Bouse formation at Parker records an interface between the early Colorado River and the sea, then basal Bouse and interfingered Colorado River deposits there would have to be older than deposits in the Laughlin area. Additionally, there should be some evidence for northward transgression of the fluvial-marine interface through the intervening areas. However, at Laughlin the (younger in the case of transgression) Bouse Formation overlies a south-directed divide-breach deposit composed of locally derived alluvium. There is no evidence of a through-flowing Colorado River below the lowest Bouse outcrops at Laughlin. This argues against the presence of the river in the Parker area before the divide between Mohave Valley and Cottonwood Valley was breached.

The simplest, but not necessarily the only, explanation of the Bouse stratigraphy in each location is that divide overtopping or seepage from a lake in Mohave and Cottonwood Valleys delivered southward flowing water into the Parker area, ultimately forming a large lake into which the Bouse Formation and early Colorado River sediment were deposited.

Conclusions

Stratigraphic evidence and tephrochronologic data from Cottonwood and Mohave Valleys reported here are consistent with the lake-spillover model of Colorado River integration. The data, their implications, and their relations to other constraints on river evolution pose some challenges to models of river inception and early evolution that invoke combinations of subsidence, headward erosion, marine transgression, marine regression, and regional uplift. Our interpretation of the field evidence reported here is that a series of lakes developed after 5.6 Ma along the course of the lower Colorado River below the mouth of the Grand Canyon to at least Mohave Valley, and then drained in succession as divides were breached and lowered. Large volumes of coarse Colorado River sediment eventually arrived in these valleys, resulting in massive river aggradation between 5.6 and 4.1 Ma. The period of thick aggradation was concurrent with drainage integration and canyon enlargement upstream and with the arrival of Colorado

River sediment in basins downstream. The proposed mode of basin interconnection and river integration through lake spillover is a well-documented phenomenon on numerous smaller river systems in the interior of the western United States. The downstream integration of the lower Colorado River was particularly effective because it connected a large, well-watered highland source area, progressively lower-lying arid valleys, and the developing Gulf of California.

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Stratigraphy Of Colorado River Deposits In Lower Mohave Valley, Arizona and California

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Abstract

Deposits in lower Mohave Valley and upper Topock Gorge near Topock, Arizona and Park Moabi, California record a succession of depositional and erosional events since late Miocene time that relate to the development of the Colorado River. Upper Miocene alluvial fans were deposited toward a depocenter east of the present valley bottom, indicating there was no valley outlet then through the area of Topock Gorge. The lower Pliocene Bouse Formation subsequently accumulated to thicknesses of at least 330 m in ancestral Mohave Valley before most of it was incised by a Colorado River that carried clasts as large as 0.9 m and drained southward through ancestral Topock Gorge. Pliocene fluvial sandstone and conglomerate exposed in this area record 160 m of subsequent aggradation. One or more sedimentary sequences of varied fluvial, paludal, and alluvial fan environments accumulated on the margin of the floodplain in Pleistocene time, including during deposition of the upper Pleistocene Chemehuevi Formation of Longwell (1936). River boulders as large as 1.2 m derived from Pliocene gravel were deposited in terraces along the river during later incision. Vertebrate fossil finds and historical descriptions of the 19th century enrich the history of the area.

Tectonic Framework

The Colorado River flows south through Mohave Valley and Topock Gorge, delineating the California-Arizona border (Fig. 1). The valley and gorge lie in the 100-km wide Colorado River extensional corridor, where extreme tectonic extension in early and middle Miocene time stretched and thinned Earth's crust by a factor of about two (Howard and John, 1987; Miller and John, 1999). Movement along northeast-dipping low-angle normal faults, often called detachment faults, extended the upper and middle crust in the corridor. Detachment faults are exposed around the Chemehuevi Mountains and other domal metamorphic core complexes in the central part of the corridor (John, 1987). These faults cut down section in the northeast direction of tectonic transport from a breakaway zone 50 km west of Mohave Valley. Three of these faults are exposed in Topock Gorge, each dipping gently east (John, 1987). Cumulative slip on the fault system in the corridor totals an estimated 40-75 km (Howard and John, 1987; Hillhouse and Wells, 1991). Regional geologic relations indicate that the basal detachment faults cut initially to depths of 10 to 15 km. By the time the extensional episode ended in late Miocene time, the ranges and basins produced by the tectonism included ancestral versions of the Chemehuevi and Black Mountains and Mohave Valley.

Stratigraphic and Geomorphic Evolution

Evidence for the later geomorphic evolution of southern Mohave Valley is recorded in a series of mostly undeformed upper Miocene and younger sedimentary deposits that unconformably overlie deformed older rocks of the

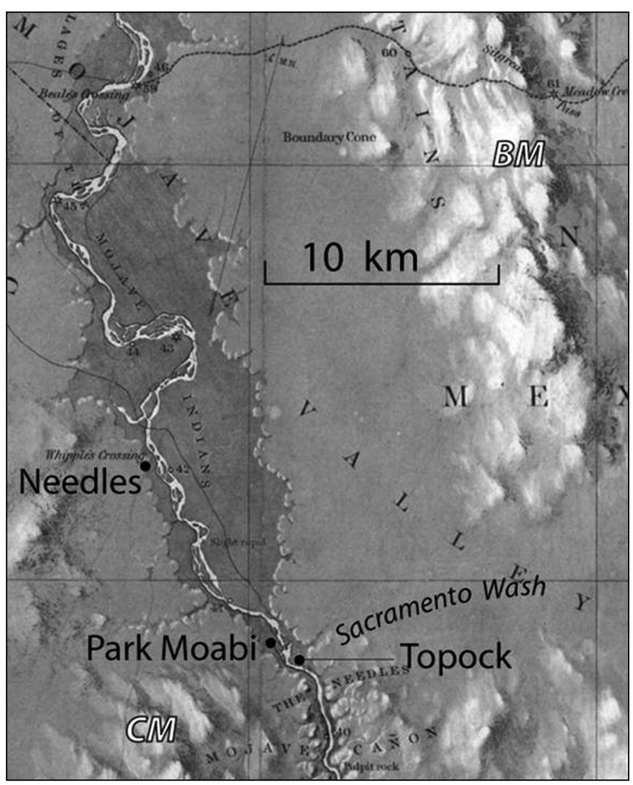

Figure 1. Ives' (1861) map of southern Mohave Valley, with modern localities added. Topock Gorge (south of Topock) was then called Mojave Cañon. CM, Chemehuevi Mountains, BM, Black Mountains.

extensional corridor. These deposits include Miocene fanglomerate, quiet-water limestone and mudstone of the Pliocene Bouse Formation, and Pliocene to modern alluvial deposits of the Colorado River and tributaries. The deposits record late Miocene filling of an internally drained ancestral Mohave Valley, its subsequent early Pliocene flooding associated with arrival of Colorado River water, and younger degradational and aggradational episodes as the river incised and aggraded multiple times. The deposits include a stratigraphic record of interactions between the river and the large tributary Sacramento Wash (Figure 1) during aggradational episodes.

Late Miocene. Locally derived upper Miocene fanglomerate constitutes the oldest undeformed stratigraphic unit. It overlies deformed rocks and predates the Bouse Formation and the arrival of the Colorado River (Metzger and Loeltz, 1973). The fanglomerate represents alluvial fans that filled an ancestral Mohave Valley basin before evidence of the lower Colorado River appeared in the stratigraphic record. The fanglomerate unit typically is moderately consolidated and commonly forms resistant cliffs. Where exposed in a railway cut 1 km northeast of Topock, the fanglomerate consists largely of greenish granitic clasts containing epidote and chlorite, characteristics of rocks that occupy the footwall of the Chemehuevi and Mohave Wash detachment faults in the Chemehuevi Mountains (John, 1987). The presence of these clasts and of northward-directed current imbrication indicates that the fanglomerate was deposited on alluvial fans emanating from the Chemehuevi Mountains area toward a basin depocenter east of the modern valley bottom. This position suggests that no outlet then existed through the route of subsequent Topock Gorge. No other candidate outlets are known. The fanglomerate is also exposed extensively along the north base of the Chemehuevi Mountains (Metzger and Loeltz, 1973).

Early Pliocene. The Pliocene Bouse Formation overlies the fanglomerate and is considered to record deposition in lakes and (or) a brackish estuary draining toward or connected to the proto-Gulf of California; it includes deltaic sand beds considered to be deposited by early lower Colorado River waters (Metzger, 1968; Buising, 1990; Spencer and Patchett, 1997; Lucchitta et al., 2001; Spencer and Pearthree, 2005; House et al., 2005). The Bouse Formation contains the record of the initial entry of Colorado River waters into Mohave Valley. At the north end of the valley, House et al. (2005) presented evidence that a boulder-carrying erosive flood immediately preceded quiet-water deposition of the Bouse Formation, and they argued that this flood occurred following the catastrophic failure of a pre-existing divide between Mohave Valley and Cottonwood Valley upstream to the north. In this view, the Bouse Formation in Mohave Valley was deposited in a lake when the initial Colorado River ponded behind a bedrock divide in the area that later became Topock Gorge (Spencer and Pearthree, 2005). The model postulates that this lake spilled southward to fill subsequent lake basins downstream.

The abrupt Bouse Formation contact on the Miocene fanglomerate is exposed in numerous places between the Chemehuevi Mountains and Park Moabi campground. A good example is about 500 meters south of Park Moabi in a railway cut, where a 1 m-thick basal white marl of the Bouse Formation overlies the fanglomerate. Elsewhere in southern Mohave Valley the Bouse Formation consists of 1-6 m of the basal white marl; tens to hundreds of meters of an overlying interbedded member consisting of green and yellow claystone, siltstone, and pink sandstone (Fig. 2); and local occurrences of tufa deposited at a shoreline. At least 15 m of the interbedded member are exposed near Park Moabi, and subsurface thicknesses of 57 and 77 m are recorded from two drill holes south of Sacramento Wash, where it occurs to elevations as low as 110 m (Metzger and Loeltz, 1973). Sandstone dikes (Metzger and Loeltz, 1973) and an intraformational unconformity over tilted and slumped beds of the Bouse northwest of Park Moabi suggest that there may have been seismic activity during deposition. The interbedded unit occurs up to elevations as high as 440 m in southeastern parts of Mohave Valley (Howard et al., 2000), an indication of the minimum level filled by the deposits. Limestone and tufa of the Bouse Formation exposed in southeastern Mohave Valley reach at least as high as 482 m elevation, where they overlook Arizona highway 95 in the present saddle connecting to Chemehuevi Valley downstream (Howard et al., 2000). This configuration suggests the possibility that the Bouse Formation bridged between the two valleys, a scenario not explained by the current version of the lake-spillover

Figure 2. Bedded mudstone and sandstone of the interbedded member of the Bouse Formation, unconformably overlain (half way up the slope) by younger alluvial fan deposits. Sacramento Wash and the Black Mountains in the background.

model of the origin of the Colorado River (Spencer and Pearthree, 2005).

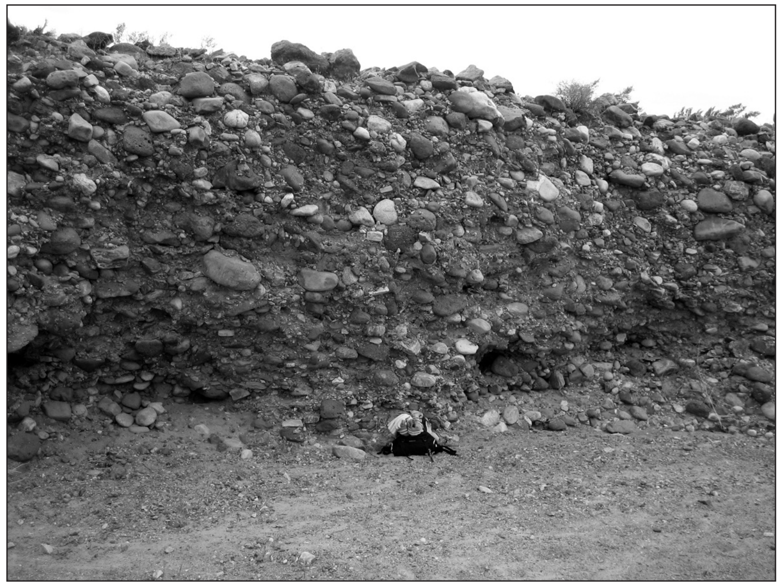

Figure 3a. Imbricated cobble to boulder conglomerate in unit B, 3 km SE of Topock. Rucksack for scale.

Figure 3b. Rib bone in unit B (in Figure 3a outcrop).

Early to late Pliocene. Pliocene fluvial roundstone conglomerate and quartz-rich sandstone containing far-traveled clasts typical of the Colorado River overlie the Bouse Formation unconformably and demonstrate that the Bouse was deeply incised by the Colorado River to an elevation of about 150 m or less above modern sea level prior to a major aggradational phase. The aggradational fluvial deposits were assigned to “unit B” by Metzger and Loeltz (1973) and are equivalent to the alluvium of Bullhead City of House et al. (2005). Fluvial cross bedding, rounded quartz grains, and rounded chert, quartzite, and limestone pebbles derived from hundreds of kilometers upstream in the upper Colorado River drainage are hallmarks of much of this unit. In the Bullhead City area of northern Mohave Valley this unit was shown to be older than 3.3 million years and mostly older than 3.6 or 4.2 million years, on the basis of tephra beds found in the unit (House et al., 2005). The basal part of the deposit near Park Moabi, and also 3 km southeast of Topock, forms a paleovalley fill with a thalweg elevation of 145–150 m (Fig. 3a). There it contains imbricated rounded cobbles of quartzite and subrounded boulders of locally derived volcanic rocks and gneiss as large as 0.9 m, among the coarsest Colorado River deposits downstream of the Grand Canyon (Metzger and Loeltz, 1973). We found a rib in a cobble conglomerate of this unit 3 km east-southeast of Topock (Fig. 3b), which has been identified as equine (R. Reynolds, written commun., 2006)

In lower Sacramento Wash and also 6 km to the north, the lowest parts of fluvial unit B are at 140–143 m elevation and consist of at least 40 m of locally folded, medium to coarse pebbly sandstone and pebble-conglomerate. An arkosic sandstone facies present along Sacramento Wash lacks far-traveled pebbles of chert, quartzite, and limestone, and so is inferred to record tributary deposition in an ancestral Sacramento Wash draining from the distant Hualapai Mountains. Deposition in this and other tributaries would have occurred in response to the fluvial aggradation documented by Metzger and Loeltz (1973) and House et al. 2005) of the valley axis. How the boulder conglomerate near Topock relates to the sandstone and arkosic sandstone is not certain. Gravel drilled to a depth of 94 m (elevation 50 m) below central Mohave Valley was suggested by Metzger and Loeltz (1973) to represent structurally lowered parts of their unit B.

Unit B is exposed on the east side of southern Mohave Valley to elevations as high as 311 m on slopes leading from the Black Mountains. Comparison of lowest and highest levels of exposure suggests that the Colorado River first incised at least 290 m into the Bouse Formation and drained southward through the area of Topock Gorge; then the river aggraded at least 160 m during a major unit B depositional episode (Metzger and Loeltz, 1973; Howard and Bohannon, 2000; House et al., 2005). Unit B is succeeded by another erosional unconformity that records re-incision of the valley during an episode of degradation. Northeast of Topock this unconformity is overlain by fluvial rounded gravel and sand inferred to be deposited by ancestral Sacramento Wash, based on southwestward current indicators and on a dominance of basalt and gneiss clasts and rarity of chert, limestone, or quartzite clasts (Fig. 4). Alluvial fan deposits that prograded into the valley also record this interval (Metzger and Loeltz, 1973). Metzger

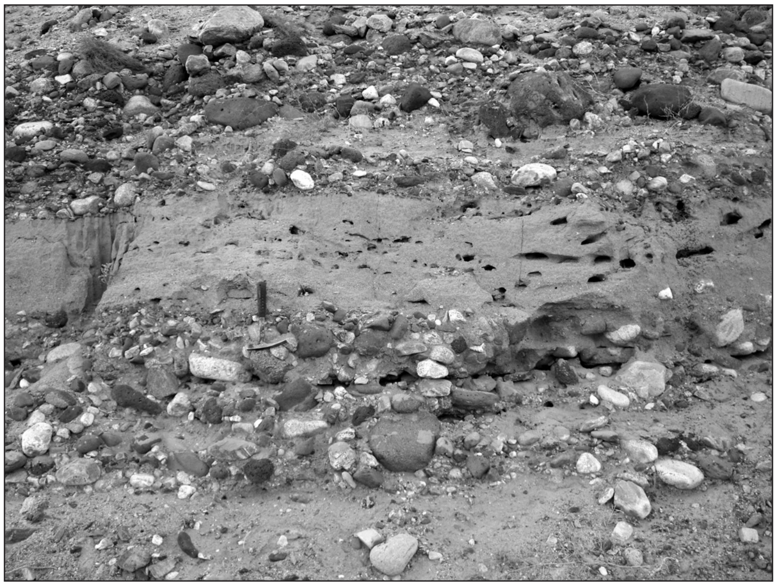

Figure 4. Sand and gravel of Pliocene or Pleistocene age, deposited above the eroded unit B in ancestral Sacramento Wash. Hammer gives scale.

et al. (1973) called these locally derived materials unit C. They may overlap in age with early Pleistocene deposits described below.

Pleistocene. A possible early Pleistocene record of lower Mohave Valley is contained in a locally folded sequence containing the remains of a mammoth that may date to the early Pleistocene. Later Pleistocene events are recorded by the Chemehuevi Formation of Longwell (1936) and younger stepped terraces.

A tentative age of early Pleistocene is assigned to a folded sequence in Sacramento Wash that overlies typical sandstone of the Bullhead City unit and consists of interbedded fluvial gravel, sand, paludal deposits, and alluvial-fan deposits. Metzger and Loeltz (1973, fig. 12) included these in their unit B. Similar correlative(?) deposits 6 km northeast of Topock near Golden Shores hosted a nearly complete mammoth skeleton, which Agenbroad (written comm. 1995 and 2005) concluded from a photograph of the no-longer-available teeth that it was probably a Mammuthus meridionalis, dating to about 1.5 to 1.7 Ma. The mammoth site occurs within a section of interbedded and interfingering Colorado River gravel, sandstone, and mudstone, including paleosols and paludal deposits containing plant roots. The mammoth was encased in fine clays; abundant imprints of reeds and sedges in the clay suggested a marshy environment (Agenbroad et al., 1992). The site reported for another mammoth tooth (M. meridionalis, L. Agenbroad, written commun., 1994) at Topock appears to be in artificial fill for the Arizona abutment of Santa Fe Railway Colorado River bridge, rendering the original sedimentary host unknowable.

A younger unit, the Chemehuevi Formation of Longwell (1936) and comprising units D and E in the Metzger stratigraphy, was deposited during one or more filling episodes in the late Pleistocene. Conspicuous outcrops of the Chemehuevi Formation occur along the eastern side of lower Mohave Valley. These sediments were deposited by the Colorado River and contain the typical two-part stratigraphy described elsewhere (see Malmon and Howard, this volume): a lower, silt- and clay-rich, well bedded subunit (unit D of Metzger and Loeltz, 1973) and an overlying subunit of loose, quartz-rich sand (unit E). The Chemehuevi Formation also occupies an abandoned paleovalley east of River Island in Topock Gorge (Lee, 1908). The Chemehuevi Formation elsewhere has yielded late Pleistocene luminescence ages between 40 and 70 ka (Lundstrom et al., 2004; Malmon et al., 2006).

On the basis of stratigraphic relations, we have also assigned other facies to the Pleistocene Chemehuevi Formation that may be specific to lower Mohave Valley. A member of pinkish fine sand and silt, yellowish gypsiferous mud, and overlying sand that is correlated to the lower subunit of the Chemehuevi Formation is exposed in Sacramento Wash and along Interstate Highway 40 (Fig. 7). The gypsiferous mud member interfingers abruptly with the locally gently folded sequence of interbedded Colorado River fluvial gravel, fanglomerate, clay, paleosols, and paludal deposits containing root casts (Figs. 5, 6). This sequence is exposed near the grade of the abandoned historic A&P Railway on the northwest side of Sacramento Wash and overlies the similar but more intensely folded sequence that was figured by Metzger and Loeltz (1973, fig. 12) in

Figure 5. Interbedded paludal and alluvial deposits on the north side of Sacramento Wash. These deposits interfinger with a gypsiferous facies of the Chemehuevi Formation. The white ledge marks a paleosol.

Figure 6. Vertical iron-stained root casts in paludal deposits near the section shown in Figure 5. Pencil for scale.

Sacramento Wash which we (above) tentatively correlated with the beds containing the Golden Shores mammoth.

We interpret the assemblage near the A&P Railway as having been deposited in an environment near the valley margin of an aggrading Colorado River. The geomorphic setting of these deposits near the mouth of Sacramento Wash suggests the distinctive facies may be the result of sediment mixing at a major tributary junction. The abundant gypsum could be the result of evaporating backwaters near the margins of the main floodplain, possibly isolated from the main stem. They may be an analog of modern ponds formed where bars on the edges of 20th century reservoirs in the river valley have dammed the mouths of tributary washes, forming some 75 backwaters in Topock Gorge. The yellow gypsiferous facies of the Chemehuevi Formation near Sacramento Wash contains gastropod shells assigned to the genus *Planorbella*, which thrives in rich eutrophic environments (C. Powell, pers. commun., 2006).

Stepped terraces along the west side of Topock Gorge 0.2 to 2 km south of Topock record stages of river incision after the Chemehuevi Formation was deposited. The highest is 55 m above the pre-dam river. The terrace deposits consist of fluvial sand and gravel and interbedded locally-derived angular debris. Boulders as large as 1.2 m in the fluvial parts of the gravel are likely derived from upstream deposits of the alluvium of Bullhead City and transported during high flows. The terraces were beveled during temporary periods of lateral erosion within a longer-term trend of Colorado River incision into unit D.

Stratigraphic Summary

Before there was a Colorado River in this area, late Miocene alluvial fans were deposited toward a depocenter east of the present valley bottom, indicating there was no valley outlet then through the area of Topock Gorge. The lower Pliocene Bouse Formation subsequently accumulated to thicknesses of at least 330 m in ancestral Mohave Valley before most of it was incised by a Colorado River that carried clasts as large as 0.9 m and drained southward through ancestral Topock Gorge. Exposed fluvial Pliocene sandstone and conglomerate exposed in this area record 160 m of subsequent aggradation. Sedimentary sequences of varied fluvial, paludal, and alluvial fan environments accumulated on the margin of the floodplain in Pleistocene times, including during deposition of the upper Pleistocene Chemehuevi Formation of Longwell (1936). During subsequent incision, large river boulders were deposited in terraces along the river at the entrance to Topock Gorge. Historical descriptions of the 19th century, described below, enrich the history of the area.

Historical Vignettes

Topock Gorge exposes Miocene to Proterozoic bedrock that is faulted, tilted, and folded in the Colorado River extensional corridor. These rocks make up the canyon walls and spires described below; locally they are overlain by remnants of Pliocene and Pleistocene Colorado River

Figure 7. Gypsiferous facies of the Chemehuevi Formation, exposed 2 km east of Topock in a roadcut on Interstate 40.

deposits. The gorge (then called Mohave Cañon) was explored in 1858 by the U.S. Army's Colorado River expedition, which was undertaken using the small steamboat Explorer for the purpose of determining the navigability of the Colorado River. Expedition leader Lt. Joseph Christmas Ives' (1861) colorful description of the approach to "Mojave Canon" from downstream was made before Lake Havasu flooded the river valley behind Parker Dam:

> To-day has been perfectly serene, and the atmosphere indescribably soft and limpid. For several miles the river assumed a new aspect, being straight and broad, having high banks, and presenting a placid unbroken sheet of water - not a bar being visible above the surface. To one viewing the noble looking stream from the bank, it would have appeared navigable for vessels of the heaviest draught, but the depth of water was scarcely sufficient to enable the Explorer to pass without touching….
>
> After ascending a few yards a harsh grating noise warned us that we were upon a rocky shoal, and Captain Robinson at once backed the Explorer out and went up in a skiff to reconnoitre....There was danger that the after part of the boat in passing might catch upon a rock, and the bow be swung around by the rapid current against another with such violence as to knock a hole in the bottom. An anchor was carried to a point some distance up stream, and a line taken from it to the bow. This line was kept taut, while, with a high pressure of steam, the Explorer was forced up the rapids, once or twice trembling from stem to stern as she grazed upon a rock, but reaching the still water above without sustaining damage.
>
> A low purple gateway and a splendid corridor, with massive red walls, formed the entrance to the cañon. At the head of this avenue frowning mountains, piled one above the other, seemed to block the way. An abrupt run at the base of the apparent barrier revealed a cavern-like approach to the profound chasm beyond. A scene of such imposing grandeur as that which now presented itself I have never before witnessed. On either side majestic cliffs, hundreds of feet in height, rose perpendicularly from the water. As the river wound through the narrow enclosure every turn developed some sublime effect or startling novelty in the view. Brilliant tints of purple, green, brown, red, and white illuminated the stupendous surfaces and relieved their sombre monotony. Far above, clear and distinct upon the narrow strip of sky, turrets, spires, jagged statue-like peaks and grotesque pinnacles overlooked the deep abyss.

From a later era, part of the historic grade of the Atlantic and Pacific Railroad is preserved along the north side of Sacramento Wash, where it flanks and is cut through interbedded Colorado River fluvial gravel, claystone, and paludal deposits. The railroad bed traces toward the Colorado River floodplain 3 km north of Topock, and was constructed across the Colorado River in 1881–1883. According to a sign on the grade:

> In 1890, after several floods, a new bridge was built at Topock and this portion of the railroad was abandoned. The A&P railroad became the Santa Fe Pacific in 1897, and the Acheson–Topeka and Santa Fe Railroad in 1902.

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Overview: the Chemehuevi Formation along the lower Colorado River

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Introduction

A distinctive set of fine-grained deposits occurs throughout the lower Colorado River Valley, extending from just below the mouth of Grand Canyon to well into the river delta below Yuma, AZ (Figure 1), an along-channel distance of over 700 km. Upstream of Parker, Arizona, the deposits consist of scattered erosional remnants up to 150 m above the modern floodplain. Below Parker, they occur in isolated outcrops but also underlie large continuous terraces such as Parker Mesa, Palo Verde Mesa, and Yuma Mesa. These deposits form the Chemehuevi Formation of Longwell (1936).

The Chemehuevi remnants have a distinctive 2-part stratigraphy that is visible from a great distance (Figure 2). The deposits usually consist of a lower, cliff-forming unit composed mostly of fine sand, silt, and clay (Unit D, in the terminology of Metzger et al., 1973), and an overlying subunit of loose, poorly bedded, medium to coarse sand that forms gentle slopes (Unit E). Unit D is well bedded, usually consisting of flat-lying, fining-upward layers tens of centimeters thick. Unit D also contains some sections consisting of several meters of nearly pure clay. Unit E is composed mostly of rounded, well-sorted medium sand with lenses of silt, clay, and angular to rounded gravel. The bedding in Unit E is less prominent than in unit D. Units A, B, and C refer to older units not considered in this paper.

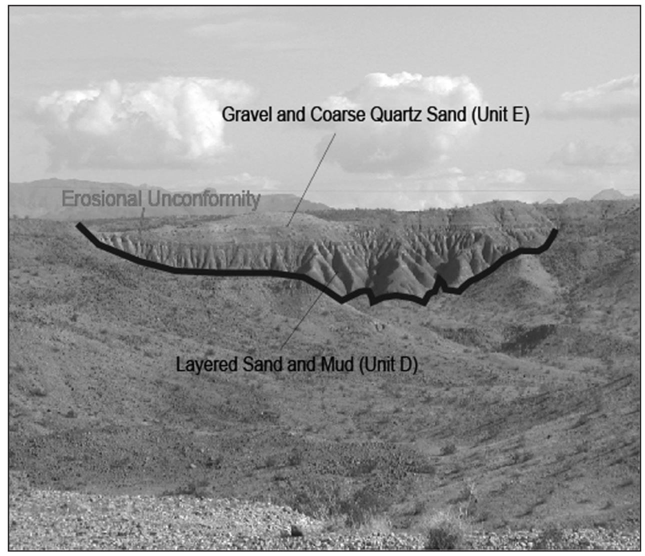

Figure 2. Photograph of Chemehuevi remnant inset into older gravel, showing basic stratigraphic relations. Location is near Golden Shores, Arizona.

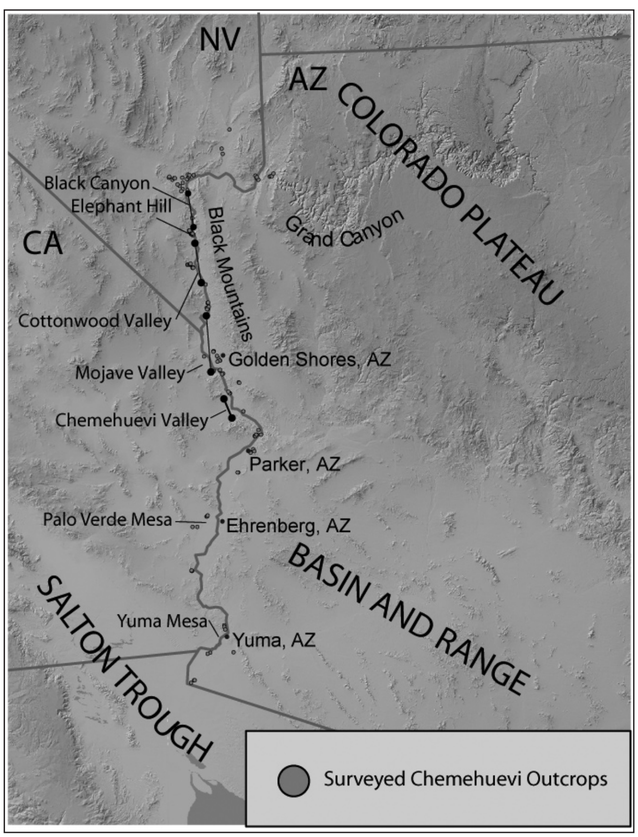

Figure 1. Location map of the lower Colorado River study area, showing the locations of surveyed Chemehuevi remnants and the trace of the subjective valley line used to measure distance on longitudinal profiles.

Interpretation of the Chemehuevi Formation

The first geologist to visit the region, J.S. Newberry (in Ives, 1861) of the 1858 Ives' expedition, described one exposure of these beds just below the mouth of Black Canyon, and found an elephant tooth in a gravel bed beneath them (Figure 3), constraining their age to Pleistocene or younger [this outcrop is now submerged beneath Lake Mojave, so we do not know how this gravel bed fits in the Metzger et al. (1973) alphabetic stratigraphic framework]. Newberry, noting the location of the outcrop just below the mouth of a giant rock-walled gorge , supposed the beds were the remnants of a giant waterfall that formed as the Colorado River eroded headward through the Black Mountains.

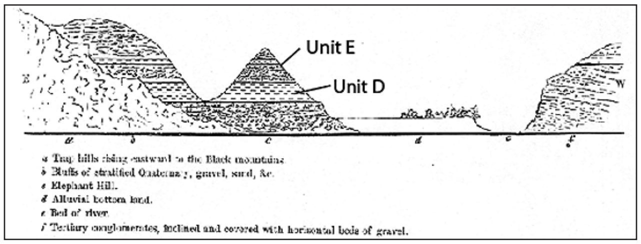

Figure 3. Drawing by J.S. Newberry of the section at Elephant Hill, near the mouth of Black Canyon. Elephant tooth was found in basal gravel of hill above c.

W.T. Lee (1908) examined the deposits throughout their range and named them the "Chemehuevis gravel" after the Chemehuevi Valley (Figure 1). While conducting a rapid characterization of the geology of the land about to be flooded by Lake Mead, C.R. Longwell (1936) renamed these deposits the Chemehuevi Formation. Noting the distinctively fine-grained nature of most of the outcrops, Longwell inferred that they were deposited in a Pleistocene "Lake Chemehuevi", which he presumed to have formed behind a natural barrier such as a lava or landslide dam, or possibly from broad regional upwarp (Longwell, 1946). He further concluded that the finer grained subunit (Unit D) likely was deposited the deeper parts of the lake (bottomset beds), and the overlying coarser sands (Unit E) by a prograding delta.

Longwell's lacustrine interpretation of the Chemehuevi deposits was discounted by Metzger et al. (1973), who noted that, because the deposits extend southward into the modern delta of the Colorado River, "it seems inescapable that the units…were deposited by the Colorado River during a time in which the Colorado was graded to the Gulf of California" (p. G30). Furthermore, Metzger et al. (1973) pointed out that the contact between the two subunits is in fact an unconformity (e.g., Figure 2), disputing Lee's notion that the two units were "perfectly conformable". Metzger et al. (1973) argued that because the subunits are separated by a depositional hiatus, they cannot be considered a formation, and they therefore distinguished them as Units D and E of the older alluvium. Metzger et al. (1973) maintained that Unit D was deposited as the river aggraded, and Unit E while the river subsequently excavated this fill.

For our work, we have generally found it useful refer to the fine-grained Pleistocene deposits collectively as the Chemehuevi Formation, because they comprise a mappable unit; we use the terms "remnants" or "deposits" to refer to individual outcrops of the formation. Also, we adopt the Metzger alphabetic nomenclature, Units D and E, to describe the two distinct subunits.

Since Metzger et al. (1973) defined the general stratigraphy and facies of the Chemehuevi beds, they have received occasional attention from geologists working in the lower Colorado River corridor. In large part this work has aimed to map and interpret the deposits, or to constrain their age (e.g., Bell et al., 1978; Blair, 1996; Faulds, 1995, 1996; Lundstrom et al., 2004; House et al, 2005; Malmon et al., 2006).

Recent mapping in the Cottonwood and Mojave Valleys has brought up the possibility that the stratigraphy of the Chemehuevi beds may be more complex the Metzger model suggests. The presence of erosional unconformities within Unit D (Faulds, 1995, 1996; House et al., 2005; and our unpublished data) may imply that the deposits consist of a nested sequence of inset units. Multiple inset fine-grained units could reflect cyclical climate change in the drainage basin during the Pleistocene, and/or sea level change at the mouth of the river. A recent topographic survey of the collected outcrops (Figure 4) delineates a topographic surface about 50% steeper than that of the historic (pre-dam) valley, which apparently contradicts the theory that aggradation cycle was caused principally by a downstream (i.e., base-level) control such as sea level rise or a natural impoundment.

While a number of possible interpretations have been proposed, the current consensus holds that the Chemehuevi Formation represent the remnants of one or more cycles of alluviation during Pleistocene time, resulting from an upstream change in water and/or sediment input from upstream. The driving mechanism may have been Pleistocene climatic change(s) in the watershed, which includes large portions of the Rocky Mountains and most of the Colorado Plateau. The nature of the climatic shift(s) that caused the Chemehuevi event(s) is not known. Glacial advances or large scale vegetation conversion are both mechanisms that might increase the erosion rate of fine-grained sediment in the Colorado River watershed. Tectonism is not likely to be a cause of the alluviation: the river mouth in the Salton Trough is generally subsiding relative to the Colorado Plateau, a differential movement that would be expected to favor river downcutting, not aggradation. An

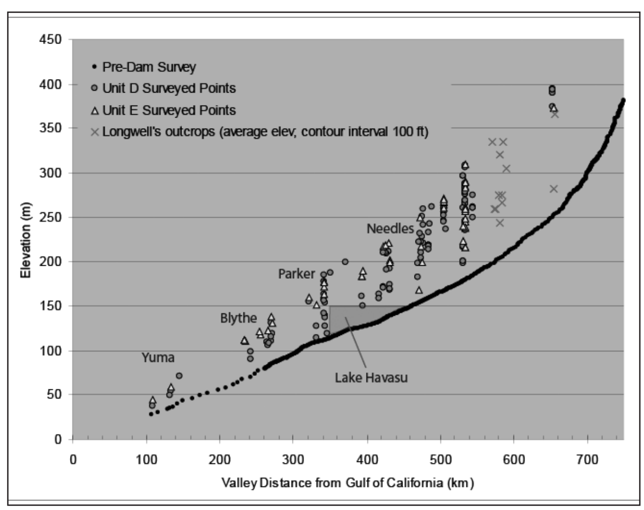

Figure 4. Longitudinal profile of Colorado River valley and the highest and lowest points of surveyed outcrops of the Chemehuevi Formation, delineating an envelope with a steeper gradient than the historic valley. "Valley Distance" is measured along a subjective valley line that is assumed to have been constant over time since the late Pleistocene. Units D and E have a similar distribution, but in some cases one may be found without the other.

alternative hypothesis that has not been ruled out (or mentioned in the literature) is that one or more major stream capture events may have introduced a large quantity of fine grained sediment into the Colorado River drainage, temporarily overwhelming the transport capacity of the river, and leading to rapid alluviation. For example, the stream capture of previously internally drained basins – features common in the Basin and Range geologic province – by Colorado River tributaries could produce a large amount of fine-grained sediment that could instigate aggradation downstream.

Age of the Chemehuevi Formation

Limited age control for the Chemehuevi Formation has hindered the understanding of its origin. In particular, more accurate ages are required to determine (1) if the scattered erosional remnants are products of a single or multiple alluviation episodes; and (2) how these episodes correlate temporally with paleoenvironmental events in the Colorado River watershed.

Vertebrate fossils (Newberry, in Ives, 1861; Metzger et al., 1973; Agenbroad et al., 1992) found within and stratigraphically below subunits of the Chemehuevi Formation show that they are Pleistocene or younger (< 1.8 million years, Ma). Furthermore, paleomagnetic studies show that they are younger than the early Pleistocene (0.78-1.8 Ma), because they yield normal magnetic polarity, consistent with the modern geomagnetic field prevailing for the past 780 thousand years (Bell et al., 1978; Lundstrom et al; in review). More recent attempts to obtain absolute age estimates have produced equivocal results. Bell et al. (1978) reported an age of 102±7 Ka (the average of Th/U and Pa/U ages) on a mammoth tusk found by Metzger et al. (1973) in gravel below Unit D near Ehrenberg, AZ. Blair (1996) obtained a finite 14C age of 35.1 ± 0.4 Ka from a wood fragment found in Unit D, an age that is uncertain because it is near the upper limit for radiocarbon dating at the time the analysis was performed.

New geochronological methods have promise for refining the chronology of the Chemehuevi Formation. Lundstrom et al. (2004) obtained uranium-series ages from carbonate coatings on gravel clasts on the geomorphic upper surface of the Chemehuevi beds, obtaining ages between 30 and 70 Ka. They also reported ages based on optically stimulated luminescence (OSL) in samples collected in Unit D, suggesting their samples were laid down between 30-80 Ka. More recently, preliminary new OSL data (Malmon et al., 2006) indicate ages of 50-60 Ka for Unit D and 30-45 Ka for Unit E, which suggest that at least the sampled outcrops may all be related to the same alluvial episode.

Beds of volcanic ash (tephra) have recently been discovered in Unit D (Figure 5) at three separate localities (House et al., 2005; Lundstrom et al., 2004; Malmon, unpublished data). Volcanic glass from these beds have been geochemically related to one another, suggesting they are probably from the same eruption, but neither the date nor the location of this eruption have been identified with confidence (USGS Tephrochronology Laboratory, unpublished data). We attempted to date the ash directly using the method of potassium-argon dating at the USGS campus in Menlo Park. However, this attempt failed because of the lack of suitable mineral grains in the sample, as well as the extremely fine-grained texture of the glass shards (an indication that the eruptive center was relatively far away).

Ongoing work

Ongoing mapping and topical research aims to further refine the age control of the Chemehuevi Formation, so that the deposits might be related to specific geologic or climatic events. In addition we are conducting geomorphological, geochemical, and paleontological investigations of the Chemehuevi beds to better understand their origin. Specific unanswered questions include:

• Was the Chemehuevi fill event(s) caused by stream capture or climate change?

• How many depositional cycles are represented by the deposits?

• If they are the result of climatic changes, what was the nature of the late Pleistocene climatic change caused the most recent "Chemehuevi cycle"? Is this climate change expressed in other climate proxies or is it a previously unknown climatic event?

• Can geochemical "fingerprinting" of the Chemehuevi deposits and modern Colorado River sediments be used to determine the source areas for the Chemehuevi fill?

• What do the vertical and lateral facies relationships of the Chemehuevi Formation indicate

Figure 5. Photograph of volcanic ash (tephra) bed in Unit D, near Ft. Mojave, AZ. The glass composing this ash has been geochemically correlated with ashes found elsewhere in Unit D-like beds further upstream; however, the ash has not yet been reliably dated.

about the geomorphologic response of a large river to a fluctuating sediment load over geologic time? In particular, what are the genetic implications of the succession of unit E (sand) following unit D (mud), and of the steeper longitudinal gradient of the valley at the peak of aggradation?

Acknowledgements

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Preliminary observations of geochemical properties of Colorado River and related sediments

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We recently conducted a pilot study of the trace-element geochemistry of Colorado-River and related sediments, to investigate whether the various important formations have characteristic geochemical "fingerprints" that could help us better understand their origins. Instrumental neutron activation analysis (INAA) uses gamma-ray spectroscopy on samples irradiated by neutrons to determine their elemental composition. This technique is capable of providing accurate measurements of the abundances of up to about 50 elements in geologic materials. Analysis of rare earth elements using INAA has been useful in identifying the likely source areas of eolian sediments, and deducing paleowind directions from loess deposits. This technique is described in detail elsewhere (e.g., Budahn and Wandless, 2002).

The samples analyzed by INAA are listed in Table 1. The 23 samples represent several broad groups: Deposits that predate the Colorado River (Bouse Formation and Muddy Creek Formations, and the Lost Cabin beds of House et al., 2005), Pliocene-lower Pleistocene Colorado River deposits (such as the alluvium of Bullhead City), Pleistocene Colorado River deposits (upper and lower parts of the Chemehuevi Formation, i.e., Units D and E in the terminology of Metzger et al., 1973), and modern (post-Hoover Dam) Colorado River sediments collected near the head of Lake Mead at the mouth of Grand Canyon. In addition, the data set contains two samples of sediment from the Virgin River near the modern delta in the Overton Arm of Lake Mead: one from the modern river deposits and one mapped as the Chemehuevi Formation by Longwell (1936).

Examples of preliminary results are in Figure 1. In general the data suggest that different stratigraphic units may differ in specific trace element geochemical signatures, such as rare earth elements. The samples that lie stratigraphically below the oldest Colorado River deposits (Bouse, Muddy Creek, and Lost Cabin beds) tend to group separately in Figure 1 from Colorado River-derived sediment (Bullhead, Chemehuevi, and historic). This is particularly evident in the plot of Hf/Ta to La/Yb (Figure 1d), where pre-Colorado River samples all have low Hf/Ta values; Virgin River-derived samples also stand apart in this plot. Another pattern evident in these graphs is that samples of the lower, bedded, mud-rich subunit (unit D) of the Chemehuevi Formation contain higher abundances of most rare earth elements compared to the overlying poorly bedded, sandy subunit (unit E). We plan to compare the chemistry to

Table 1: Description of INAA samples

Sample ID	Description[1]
Modern Colorado River samples	
311-5	Delta cross-bedded fine sand, Pierce Ferry, AZ
311	Modern Colorado River, overbank sand near delta, Pierce Ferry, AZ
Chemehuevi Formation upper subunit (unit E)	
06-501-5	Low elevation unit E, near Elephant Hill, AZ (OSL)
145	Sandy Point fine sand bed, AZ (OSL)
227	Unit E Copper Basin Wash, below Parker Dam, CA
Chemehuevi Formation lower subunit (unit D)	
06-429-1-3a	North of Nelson, NV upper outcrop, unit D sand (OSL)
06-429-1-3b	North of Nelson, NV, upper outcrop, unit D mud
06-501-4	Low elevation unit D, near Elephant Hill, AZ (OSL)
228	Unit D below Parker Dam, CA (OSL)
305	South Cove exposure, AZ, top of unit D-like beds
Bullhead City unit (unit B) and related samples	
LM-SP-2	Sand below 4.4-Ma basalt, Sandy Point, AZ
162e	Tyro Wash unit B, AZ
HD-SL-SD-0	Sand from Hoover Dam paleochannel, AZ
156	Base of Bullhead City unit, "Panda Wash", NV
Bouse Formation samples	
331-8	Bouse clay, Parker Valley, AZ, pmag site (normal polarity)
321	Bouse clay near Park Moabi, CA
H06-TO-21	Bouse sand, Topock quad, CA
Other Samples	
322	Pink beds at Denny's, Needles, CA[2]
HS-LC-132	Lost Cabin beds, Cottonwood Valley, AZ
Tributary Sediment	
21	Las Vegas Wash Chemehuevi, NV[2]
216	Virgin River Chemehuevi - the narrows (OSL), NV
218	Modern Virgin River sand, south of Overton, NV
166	Muddy Creek Fm near Overton, NV

Notes:
1. OSL in parentheses in description indicates sample came from bed also sampled for optically stimulated luminescence
2. Samples not plotted in Figure 1.

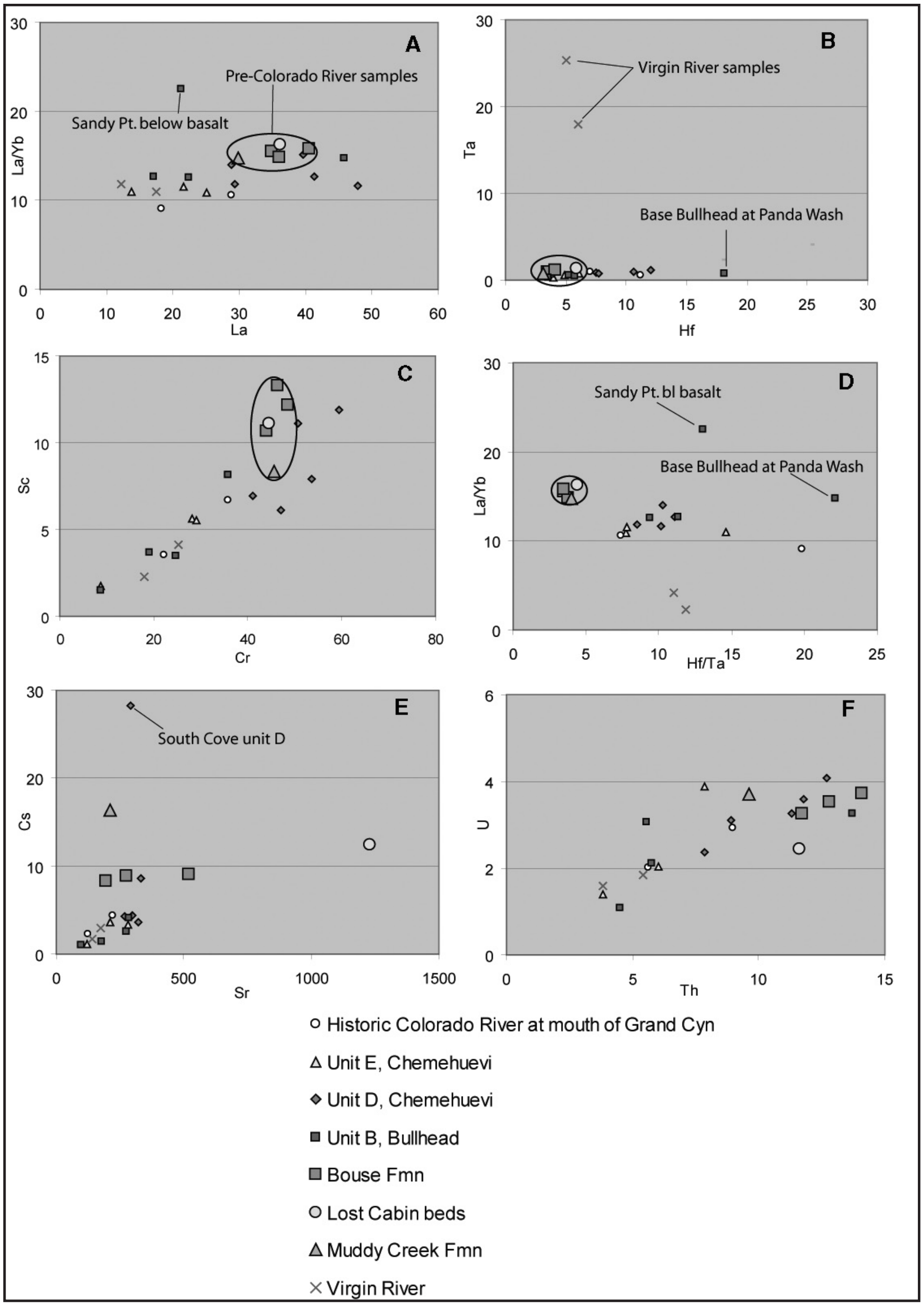

Figure 1. Selected plots of trace elements in Colorado River sediments based on INAA. "Sandy pt bl basalt" is fluvial sand below 4.4-Ma basalt at Sandy Point, sample LM-SP-2; Base Bullhead at Panda Wash, sample 156; South Cove unit D, sample 305.

particle-size data that we have for many of these samples to investigate the role that grain size plays in determining trace element chemistry.

These preliminary results are not yet synthesized or rigorously evaluated. We plan to expand this work to incorporate samples from major tributaries of the Colorado River, and conduct mixing analyses to examine the relative contributions of different potential source areas to the major stratigraphic units of the lower Colorado River.

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Notes on the fringe-toed lizard (Genus *Uma*) habitat in the east Mojave Desert, San Bernardino and Riverside Counties

William Presch Ph.D., *Director, Desert Studies Center, Zzyzx, CA*

Fringe-toed lizards are a moderate-sized, heliothermic phrynosomatid lizard in the genus *Uma* that are restricted to wind-blown sand dunes and ramps in the southwestern United States and south into northern Mexico. The genus contains five species: *Uma inornata, Uma notata, Uma scoparia, Uma exsul,* and *Uma paraphygas* (Fig.1). Within the California Desert sand ecosystem are *Uma inornata*, the Coachella Valley fringed-toed lizard, an endangered species; *Uma scoparia*, the Mojave fringed-toed lizard; and *Uma notata*, the Colorado Desert fringed-toed lizard. The latter two are listed as species of special concern in the State of California (Jennings and Hayes, 1994). *Uma rufopunctata* is restricted to southeastern Arizona. *Uma exsul* and *U. paraphygas*, found in northern Mexico, are endemic and endangered (Trepanier, et al., 2001).

Figure 1. Uma scoparia .

Fringed-toed lizards are found in the peripheral areas of sand dunes and associated with vegetation. They do not appear to inhabit unstable sand dunes that lack vegetation. Pliocene and Pleistocene lake beds and streams are the source of the sand and the historical connections between populations (Murphy et al., 2006). As the lakes and streams dried, the dune populations became isolated. Fringe-toed lizard distributions are related to the pluvial drainage and occupy many of the aeolian sand concentrations in the Mojave Desert: Bristol Dry Lake, Cadiz Dry Lake, Dale Dry Lake in San Bernardino County; Rice Valley, Pinto Basin, Palen Dry Lake and Ford Dry Lake in Riverside County; Bouse Wash southeast of Parker, La Paz County (Hollingsworth, et al., 1998; Murphy et al. 2006). Habitat between populations is unsuitable, resulting in "island" populations preventing exchange of genetic material between populations. The habitat is directly affected by factors that distribute sand distribution such as wind patterns, removal of sand by floods, off-road vehicle misuses and urban development, conversion of land to agriculture and other uses (Turner et al., 1984; Jennings and Hayes, 1994; Barrows, 1996).

The distribution of the Mojave fringed-toed lizard, *Uma scoparia*, extends from southern Inyo County through most of eastern San Bernardino County, south and east through the eastern portion of Riverside County to the area of Blythe (Fig.2). A single record occurs at Parker, Yuma County, Arizona and in Bouse Wash. The species ranges from below sea level to about 900 feet in the Kelso Sand Dunes, San Bernardino County (Jennings and Hayes, 1994).

FTL habitat is characterized as fine, aeolian sand dunes and ramps on the margins of lakebeds and washes. Some populations are restricted to isolated pockets of wind-blown sand on the sides of hills. In all habitats, widely distributed plants provided shade for thermoregulatory behavior and burrowing cover to escape heat and predators.

Fringe-toed lizards exhibit a number of specializations for life in an aeolian environment. Fringes on the hind toes enable the lizard to move on the substrate without sinking. Posterior-oriented external nares help prevent sand from entering the nose when burrowing. Ear coverings and "eyelashes" help keep sand out of the ears and eyes.

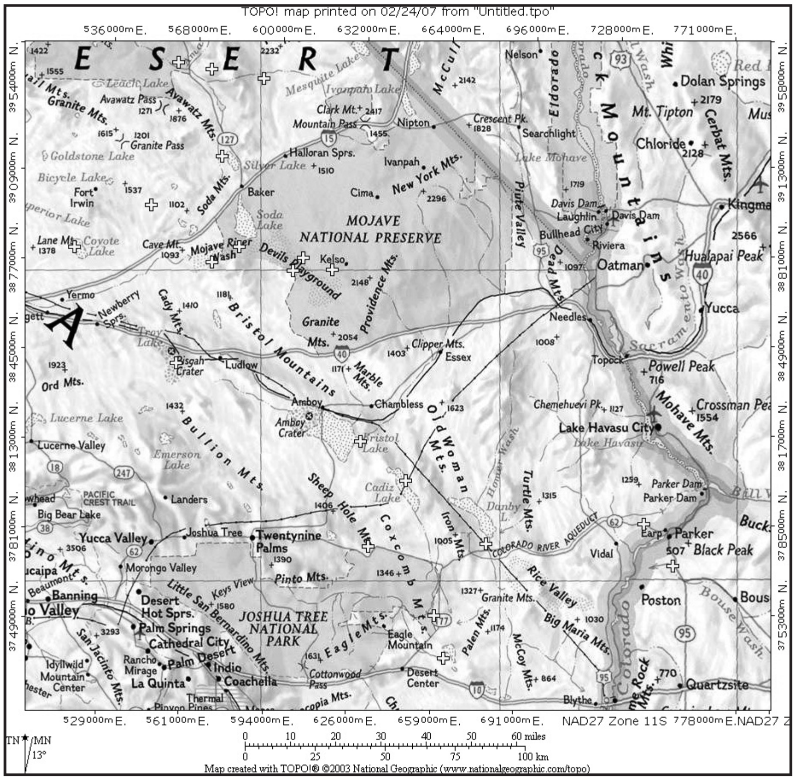

Figure 2. Locations of known populations of Uma scoparia in San Bernardino and Riverside counties.

A countersunk jaw contributes to a "shovel-shaped" snout and a dorsoventrally compressed body contribute to the ability to dive (burrow) into the sand. These lizards are often referred to as "sand swimmers" (Norris, 1958; Pough, 1969, 1970; Stebbins, 1944, 1972; Carothers, 1986).

Fringed-toed lizards are omnivorous, feeding on a range of seeds, flowers, grasses, leaves, and insects. Plant material forms the major food source in the spring but a shift occurs to insects later in the year (Minnich and Shoemaker, 1970).

Fringe-toed lizards reach sexual maturity in the second summer after hatching (Jennings and Hayes, 1994). Adults breed between April and May. Reproduction is dependant on the amount of rainfall the previous winter. One to six eggs are laid in the sand in May to July. No parental care has been observed.

The preferred temperature for activity is between 30°C and 50°C sand temperature. The preferred body temperature is maintained at 37.5°C (range 28.5°C-44.2°C) (Mayhew, 1964).

Predators include badgers, coyotes, roadrunners, burrowing owls, other lizards and various snakes (Norris, 1958; Miller and Stebbins, 1964; Gracie and Murphy, 1986).

The route of the field trip will take us by several of the sand systems with an opportunity to observed fringe-toed lizards. A sand field six miles south of the Desert Studies Center in the BLM Rasor open area provides a good representation of the habitat as described above. Areas of sand ramps can be seen between Afton Road and Basin Road along I-15. Sandy portions of the field trip route include Arizona Highway 95 from Bullhead City north to Golden Shores; along the Colorado River at Parker, and in sand deposits in Chemehuevi Wash and Bouse Wash south of Parker. Sandy habitats occur in the northwest area of the Cactus Plain. The route will end near a sand field and dunes west of Danby Dry Lake.

Acknowledgement

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The native freshwater fish fauna of the lower Colorado River: history and demise

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Over 5 million years ago, the ancestral Colorado River entered the Salton Trough of southern California and started depositing sediments that later formed its large delta. At that time, the trough was inundated by a northern extension of the Gulf of California. Later, sediments eroded from the Colorado River drainage eventually filled the southern end of the basin and blocked the marine waters of the Gulf.

Most of the native freshwater fishes of the lower Colorado River are found no where else in the world. Over the last 120 years a number of man-made changes to the river and its basin brought most of the species to the brink of extinction.

Natural History

In the late nineteenth century, there were 9 species of native freshwater fish in the lower Colorado River. "Fish found in the lower Colorado River were unique, having the highest proportion of endemism (75 %) in the Nation. They occurred nowhere else in the world. They evolved over millions of years in the harshest river environments known. They had learned how to survive floods, prolonged droughts, extreme temperatures, and salinities that few other fish could tolerate [Meuller and Marsh, 2002]."

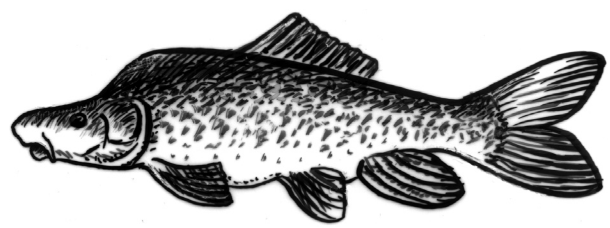
Floods would scatter the fish. Droughts caused large fish kills. Air temperatures along the river seasonally ranged from below freezing to over 120 degrees F. Not only did the Colorado River carry loads of suspended mud and sand, it also contained a number of dissolved salts which affected its salinity. Yet, under these harsh conditions, these fish not only survived but flourished.

Information concerning lower Colorado native freshwater fishes was taken from the following references: McClane, 1978, McGinnus, 1984, Mueller and Marsh, 2002, Page and Burr, 1981 and Swift, el al, 1993.

Catostomidae-suckers

Xyrauchen texanus—razorback sucker

The razorback sucker (formerly called the humpback sucker) has a pronounced hump or keel dorsally just posterior to the head and is the largest catostomid (sucker) in the Colorado River drainage. It has been a popular belief that the "hump" helps with improved stability in swift moving water, but recent studies indicate that it may prevent this fish from being swallowed by predators. Razorback suckers reach lengths up to one meter (39 inches) and weights up to 7.3 kg. (16 lbs.). To the Native Americans living along the lower Colorado River and Lake Cahuilla,

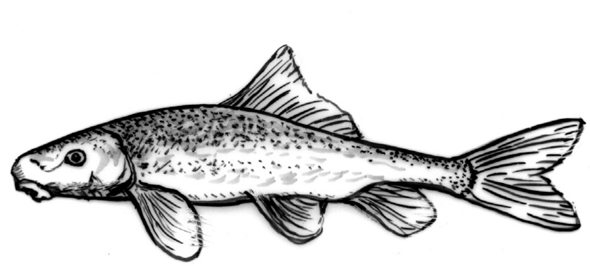


Xyrauchen texanus—razorback sucker

this fish was an important food item (Gobalet 1992, 1994; Gobalet and Wake, 2000, McGinnus 1984). The razorback sucker prefers the slower moving parts of large streams and their backwaters. It is a winter-spring spawner, starting in late January and continuing into April. This sucker feeds on detritus and algae.

Catostomus latipinnus—flannelmouth sucker

This sucker, which is smaller than the razorback, reaches lengths up to 2 feet. The flannelmouth sucker is only found in the Colorado River drainage from southwest Wyoming to southern Arizona. It inhabits creeks and rivers and feeds on vegetation. A spring spawner, the flannelmouth sucker is considered a food fish.

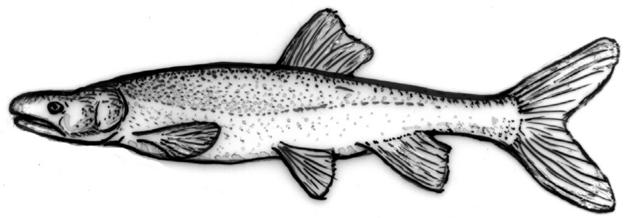


Catostomus latipinnus—flannelmouth sucker

Cyprinidae-minnows

Ptychocheilus lucius—Colorado pikeminnow

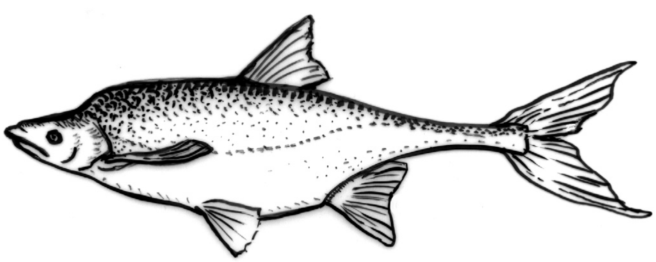
Formerly called squawfish, the Colorado pikeminnow is the largest cyprinid (minnow) in North America, attaining lengths up to 1.8 meters (nearly 6 feet) and weights of 45.5 kg. (nearly 100 pounds). The top predator in the Colorado River, this fish feeds on other fish and a variety of terrestrial animals. Pikeminnows are usually found in deeper stream channel or backwaters where they can move into shallower water to ambush prey. In the 1880's, there was a fishery for this fish at Yuma, Arizona (Dill 1944). Called "salmon" by the early settlers, pikeminnows migrate hundreds of miles to spawn in upstream nursery areas and confluences of major tributaries. Spawning happens after the peak of the spring runoff and when water temperatures reach 70 degrees F.



Ptychocheilus lucius—Colorado pikeminnow

Gila elegans—bonytail

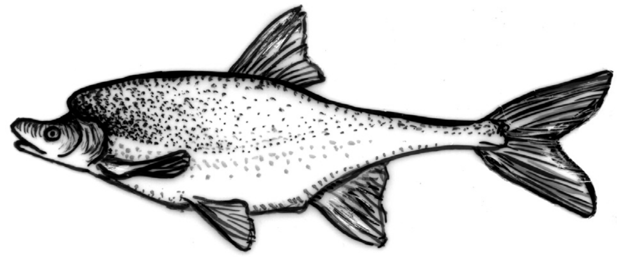
Called a "chub", the bonytail is the largest of several chubs (genus Gila sp.) found further upstream. It is the most common chub with a streamlined body with an extremely thin or "bonytail" and a smaller hump between the head and the dorsal fin. Bonytails attain lengths up to 62 cm (24.5 inches). This fish feeds on a wide variety of aquatic and terrestrial insects, worms, algae, plankton, and plant debris. Bonytails are usually found in eddies and pools along streams, and spawn in deeper habitats in April or May when temperatures reach 60 to 65 degrees F. Bonytail bones are the most adundant fish species recovered from archaeological sites around the old shorelines of Lake Cahuilla in the Salton Trough of southern California (Gobalet, 1992, 1994; Gobalet and Wake, 2000).



Gila elegans—bonytail

Gila cypha—humpback chub

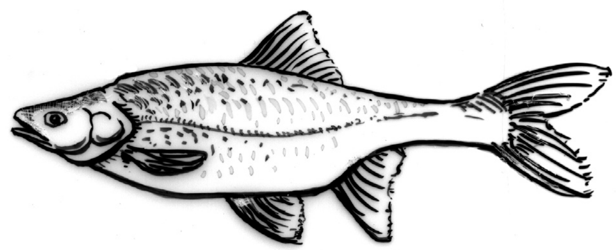
Like the bonytail chub, humpback chubs are streamlined and have an extremely pronounced dorsal hump. A little smaller than the bonytail, the humpback chubs reach lengths up to 50 cm (20 inches) and were not recognized scientifically until 1946. These fish live exclusively in deepwater canyon stretches along the main Colorado River and like other Colorado River fishes, spawning took place during the high spring runoff. Humpback chubs feed on wide variety of terrestrial and aquatic invertebrates, algae, and plant material.



Gila cypha—humpback chub

Gila robusta—roundtail chub

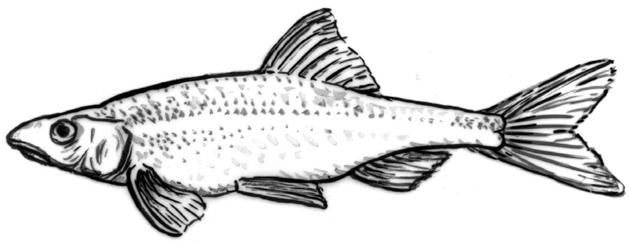
Smaller than the other two chubs, roundtail chubs attained lengths up to 43 cm (17 inches). From the early 1900's, there are a few records of these fish in the lower Colorado River, but today roundtail chubs still occur farther upstream in larger tributaries. Their elongated and streamlined fins make them well adapted for swift moving water. Spawning usually occurs in May when water temperatures reach 65 degrees F.



Gila robusta—roundtail chub

Plagopterus argentissimus—woundfin

Collected before 1900 in the lower Colorado River, the woundfin undoubtedly was in the main river in California, although collected specimens do not exist. Unlike the larger chubs, these minnows only attain lengths up to 9 cm (3.5 inches). As their body shape suggests, woundfin prefer living in waters that are silty and swift moving. These fish are usually found in streams less than 3 feet deep and is rarely found in pools. Woundfin feed on a variety of aquatic and terrestrial invertebrates, plant material, and detritus.

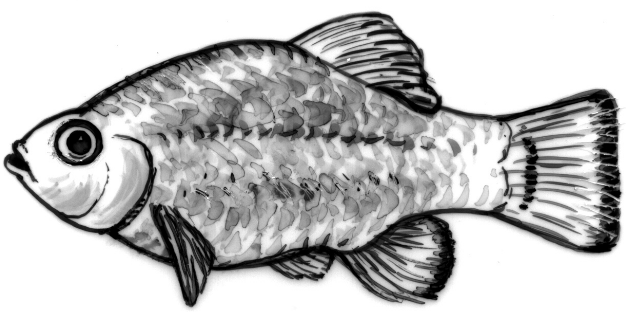


Plagopterus argentissimus—woundfin

Cyprinodontidae—Pupfishes

Cyprinodon macularius-desert pupfish

Found in the lower Colorado River and the Gila River, the desert pupfish is less than 2 1/2 inches in length. Truly a desert fish, pupfish can tolerate wide range of temperatures



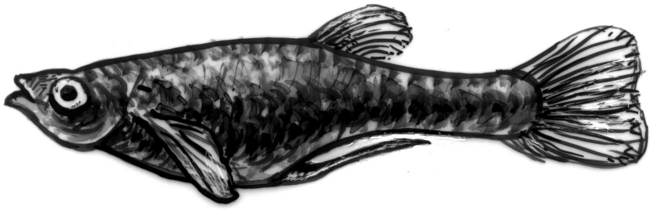
Cyprinodon macularius-desert pupfish

(from O to 112 degrees F.) and salinities more than twice that of sea water (90 ppt). Because of these tolerances to extreme environmental conditions, they thrive in waters where few other species can survive. Along the lower Colorado, the desert pupfish is found in springs, marshes, lakes, and pools of creeks over mud or sand bottoms. They feed on a wide variety of invertebrates and algae.

Poeciliidae—livebearers

Poeciliopsis occidentalis-Sonoran topminnow

Also known as the Gila topminnow, Sonoran topminnows are found in the Gila River drainage, and streams south into western Mexico. The only records for California were before 1900. The Sonoran topminnows inhabit shallow, marginal spring and stream habitats. They prefer moderate depths and areas near aquatic vegetation. Males only reach lengths approaching one inch and females



Poeciliopsis occidentalis-Sonoran topminnow

nearly 2 inches. Topminnows are live bearers, with the female carrying the young internally until birth. They feed on a wide variety of small invertebrates.

The Fossil Record

During the early Pliocene, a large freshwater lake or system of lakes covered most of northeastern Arizona and extended into New Mexico. This extensive lacustrine environment supported a diverse community of fossil cyprinid fishes (Uyeno and Miller 1965). Fossil remains of an extinct genus (Evomus) and moderated-sized chubs related to the genus Gila were collected and described from the Bidahochi Formation. Also remains of a extinct large carnivorous minnow, Ptychocheilus prelucius, was recovered and described from this fauna. Its elongate jaws and pharyngeal bones were adapted for piscovy (fish eating). No doubt this is the ancestor of the modern pikeminnow (Ptychocheilus lucius).

Another fossil fish fauna from Gila River drainage, Arizona, the Snowflake fauna has yielded Pleistocene aged Cyprinidae and Catostomidae fossils (Uyeno and Miller, 1963).

Over the last 60 years, fossil collecting by volunteers and staff in the Anza-Borrego Desert State Park and the Natural History Museum of Los Angeles County has yielded fossil records of native freshwater fishes of the lower Colorado River. At least three species occur in the Pliocene and Pleistocene deposits of the Palm Spring Group. This rock unit represents sediments of the Colorado River delta and locally derived basin fluvial and lacustrine deposits. Based on ash dates and paleomagnetic samples, the age of the Palm Spring Group ranges from 0.9 to 4 million years. From the Arroyo Diablo Formation rocks came one of the more interesting fossil freshwater fish records. A nearly complete fossil razorback sucker skeleton in a sandstone concretion was collected by Ruth Coyle in the mid 1970's on the north side of the San Felipe Hills (Stewart and Roeder, 1993; Hoetker and Gobalet, 1999). In the early 1980's, the specimen was donated by Ruth to the San Bernardino County Museum and was brought to my attention by Bob Reynolds. The Arroyo Diablo Formation is roughly 2-4 million years old and represents old Colorado River delta sands, silts, and clays. According to Hoetker and Gobalet (1999), the San Felipe Hills Xyrauchen is indistinguishable from the modern X. texanus (razorback sucker). Other younger occurrences of this species in the Palm Spring Group consist mainly of isolated bones and vertebrae. They are the most common fossil fish remains in the Anza-Borrego sediments and are found throughout the Hueso Formation and Ocotillo Conglomerate (Roeder, 2005;Gensler, Jefferson and Roeder, 2006).

Another fossil lower Colorado River fish recorded from Anza-Borrego is the modern pikeminnow (Ptychocheilus lucius). Unlike Xyrauchen, fossils of Ptychocheilus, which consist of isolated vertebrae, are rare. Based on their stratigraphic occurrence, the age of the earliest fossil pikeminnow remains at Anza-Borrego is about one million years (Roeder, 2005; Gensler, Jefferson, and Roeder, 2006).

Based on isolated pharyngeal (throat) teeth, a second cyprinid fish was identified as Gila sp. (chub) from Anza-Borrego sediments. Today, there are three species of Gila (G. cypha, G. elegans, G. robusta) that occur in the Colorado River drainage system. Also two species, Gila orcutti and G. bicolor (=Siphateles of some researchers) that occur to the west and north of Anza-Borrego respectively. Until more diagnostic elements of these fossil fish are found, identification will remain only at the generic level. The earliest occurrence of Gila remains in Anza-Borrego sediments is about one million years.

Recently, additional Colorado River freshwater fossils were recovered by volunteers, students, and staff of Arizona Western College and the George C. Page Museum in Irvingtonian-aged sediments of the El Golfo region of northwestern Sonora, Mexico (Roeder, this volume). Isolated bones of Xyrauchen texanus (razonback sucker) and vertebrae of Ptychocheilus lucius (Colorado pikeminnow) were collected. These fossils may be 1-2 million years old.

The fossil fish species identified from ancient Colorado River delta sediments at El Golfo and Anza-Borrego Colorado River delta derived sediments indicate that all species identified may represent extant or living forms.

The Demise

Today in the California and Arizona portion of the lower Colorado River, most of the native freshwater fish fauna is either locally extinct or extirpated, or endangered. Prior to 1900, all of these fish were abundant in the lower Colorado River, but since the 1950's have become extirpated or very rare. Two of the main causes of the decline of the native freshwater fishes of the lower Colorado River are dramatic changes of the physical and hydraulic changes to the basin by water development agencies and the introduction of nonnative freshwater fish stocks by fish and game agencies and sport-fishery enthusiasts.

The 1890 to 1935 Decline

As areas developed agriculturally along the Colorado River basin in California and Arizona, a number of changes affected the native freshwater fishes. Probably the most severe changes were stabilization of the flow of the lower Colorado River and its tributary streams by dams and other water diversion structures. Starting in 1909 with the construction of the Laguna Dam at Yuma, Arizona, the first physical barrier across the Colorado River, this and later water diversion projects greatly affected the native fishes. Although Laguna Dam only diverted summer flows into irrigation ditches, it reduced flows downstream to the Colorado River delta, an area used by bonytails, razorback suckers, and pike minnows as a spawning and nursery area. The purpose of this dam and later dams were flood control and water diversion for agricultural irrigating the desert lands and later sending water to growing urban centers. The following dams were built on the lower Colorado River; Laguna Dam-1909, Hoover Dam-1936. On the Salt and Verde Rivers, Arizona, the following dams were built; Granite Reef Diversion-1908, Theodore Roosevelt-1911, Mormon Flat-1926, Horse Mesa-1927, Stewart Mountain-1930. And on tributary rivers, the following dams were built; Coolidge-1928, Carl Pleasant (New Waddell)-1928. These dams had a profound effect on the populations of native fishes. These projects changed the physical character of the river which originally was warm, muddy, and swift. The dam made the river water cooler, clearer, and slow. These greatly affected the ecology and habits of native fishes. For example, irrigation in March and April diverted water into agricultural lands and affected spawning areas. Newly hatched fish would drift downstream to historic nurseries only to be swept into fields where they died. Also adult fishes would end up in the same areas and were stranded and died by the thousands. Another effect of the water diversions was dewatering of habitats along the Colorado River. As the river was controlled by occupying a restricted course, areas such as marshes, backwaters, shallow lagoons and lakes were reduced and in many areas eliminated which had a dramatic effect on native fish populations. These were areas where native fishes fed, bred, lived and died.

Probably as destructive as the water diversion projects, were commercial fishing and the introduction of nonnative species. Fish made up 15% to 20% of the diet of Native Americans, but the effect on native freshwater fishes was minimal. But later settlers were more effective ways in harvesting fish, and reduced the numbers of native fishes. Until 1910, fish stranded in canals or concentrated behind dams were easily harvested for meat, fertilizer, and hog feed. Pikeminnows were taken to supply food for construction crews and for a time were commercially canned near Yuma. Fish added protein to the diets of locals and fishing became a popular pass-time.

The growth of the sportfishing industry started with the introduction of nonnative game and bait fishes and the creation of reservoirs and canals that were used for fishing.. Stocking of nonnative fish started as early as 1881 and by 1910, common carp, bullhead, and channel catfish were common throughout the lower river (Mueller and Marsh, 2002). The following fish were stocked in the lower Colorado River; common carp-1881, American shad-1884, channel catfish-1892, yellow bullhead-1899, cutthroat trout, rainbow trout-pre-1900, black bullhead-1904, brown bullhead-1910, brook trout-1920, mosquitofish-1922, brown trout-1924, white crappie-1934, black crappie-largemouth bass-1935. All of these fish had a dramatic effect on the native fish populations. These nonnative fishes competed with the native fishes for food and space. An example of the decline of one species in a short period of time was the introduction of the smaller minnow-like fish, the mosquitofish in 1922. Within a year or two it had effectively replaced the Gila or Sonora topminnows by eating their eggs and young. This probably happened to the young of larger native fishes. The native fishes could not compete with the nonnative game and bait fishes for several reasons. The native fishes had adapted to the cycle of drought and flood sequences of the free flowing lower Colorado River. The populations of exotic fishes were small because of environmental hazards, such as drought and floods. Construction of dams and other diversionary structures that controlled and smoothed out the highs and lows of the river flow, greatly aided the survival and greatly increased reproduction rates along the nonnatives game fishes. And what the exotics did was to feed on the eggs and young of native fishes.

The 1936 to 2007 Disappearance

With the construction of Hoover Dam in 1936 and other later large water development projects for flood control and irrigation projects, the demise of the native fishes continued. Water for vast farming areas and faraway metropolitan population centers provided the impetus for larger water reclamation projects: Since 1936, the following dams have been built on the lower Colorado River: Hoover-1936, Imperial-1938, Parker-1938, Headgate Rock Diversion-1944, Morales-1950, Davis-1953, Palo Verde Diversion-1957. On the Verde River, Arizona, the following dams were built: Barlett-1939, Horseshoe-1946. And on tributary rivers, the following dams were built: Painted Rock-1960, Senator Wash-1966, Alamo-1968, Quail Creek-1985, Henderson (Lake of Las Vegas)-1992. The lower Colorado River was segmented. The river went from a free flowing river that was subjected to periods of low water and times of flood, to a tamed river with large portions of its course underwater because of reservoirs and lakes impounded behind dams and other diversionary structures. All of these structures greatly affected the physical nature of the lower Colorado River basin. And these changes greatly benefited the nonnative game and bait fishes.

The following nonnative fishes were introducted into the lower Colorado River basin: bluegill, sunfish-1937, smallmouth bass-1940's, goldfish-1944, fathead minnow-1950's, banded cichlid-1950's, guppy, Mexican tetra, sailfin molly, shortfin molly, green swordtail, southern platyfish-1950, redside shiner, leatherside chub, Utah sucker, Rio Grande chub, Rio Grande sucker-1950, dusky mountain sucker-1950, Sacramento hitch- 1950, Rio Grande killifish, longjaw mudsucker-1950, mottled sculpin-1950, Utah chub-1951, redear sunfish- 1951, White River spinedace, yellow perch-1951, California killifish-1951, golden shiner, red shiner-1953, threadfin shad-1953, warmouth-1958, stripped bass-1959, blue catfish, Mozambique mouth-brooder-1960's, zilli tilapia-1960's, flathead catfish-1962, sockeye salmon-1962, white catfish-1963, white sturgeon-1967, coho salmon-1966, rainbow trout-1969, walking catfish-1970, walleye-1971, freshwater eel-1972, cutthroat trout-1972, cuttbow trout-1975. Many of these fish, especially the exotic bait fishes, did not take and were not seen again. But others had a disastrous effect on the native fishes. As exotics like bluegill, common carp, largemouth bass, striped bass, channel catfish, sunfish, black crappie, and other fish became abundant and were taken by anglers, the native fish population plummeted. Fish and game agencies considered them "trash fish" and in some cases they were poisoned to make room for more desirable game fish. Or native fishes were caught by anglers and thrown up on the bank to feed the coyotes. Although rare, native fishes were taken from the lower Colorado River basin until the 1960's and early 1970's. Because of their peculiar looks, anglers often brought them to the attention of fish and game wardens and other wildlife officials. In the early 1980's, a few very old native fishes were present in reservoirs, but most of the native fishes were gone.

The current state of the native fish community

"In the early 1960's, the people of the United States began to recognize that many of the native fish, wildlife, and plant species had been so depleted in numbers that they were in danger of or were threatened with extinction [Johnson and Rinne, 1982]". Out of the awareness came a number of legislative attempts to protect these vanishing species. At the forefront was the Endangered Species Act of 1973 which was later amended in 1978, 1979, and 1982. The Act was implemented in three areas: listing, protection, and recovery.

Listing defined an endangered species as any species that was in danger of extinction throughout its range or a significant portion of its range, and listing as a threatened species as any species that is likely to become an endangered species within the foreseeable future thoroughout all or a significant portion of its range. Of the 9 native fishes of the lower Colorado River, 6 species (Sonoran topminnow-1967, woundfin-1967, humpback chub-1967, pikeminnow-1976, bonytail-1980, razorback sucket-1991) have been listed as endangered species in the lower Colorado. The roundtail chub is being considered for federal listing in the lower basin. The desert pupfish and flannelmouth sucker have not been listed.

Protection of native lower Colorado River fishes consists of two parts. One involves environmental reviews of federal agency actions that might affect list fish species and their critical habitats. The other is the enforcement of the prohibition of taking of listed or protected species by individual citizens.

As for environmental reviews, federal projects, especially construction projects that might impact native fishes, are studied in the early phases of planning by consultants "to make reasonable and prudent alternatives that minimize or eliminate the potential adverse impacts to the species and its habitat [Johnson and Rinne, 1982]".

Recovery, unlike listing and protection which are activities to slow the decline or maintain the existing populations of listed species, are actions that are design to actively reverse declining trends. These actions are the most expensive and controversial of the three principal areas of the endangered species act. The following section will detail the present state, recovery effects and future of each of the native fishes of the lower Colorado River. Unless noted, this information was taken from the following references: Johnson and Rinne, 1982, Swift, el al 1993, and Mueller and Marsh, 2002.

Razorback sucker

Razorback suckers, which were once one of the most abundant and widely distributed species in the larger streams of the lower Colorado River basin, today number less than 4000 individuals in the wild. Most of these are old fish found in Lake Mojave and Lake Mead. Of the four "large river" fish species of the lower Colorado (razorback sucker, humpback chub, Colorado pikeminnow, and bonytail chub), only the razorback sucker appears to be able to survive in abundance in reservoir habitats, but only as adults. The young and juvenile survival is extremely poor because of predation by large exotic game fishes such as catfishes, and basses. Reintroduction of this species back into its former habitat has for the most part has been a failure. Over 12 million suckers were reintroduced into central Arizona streams, but recapture censuses could only find less than 150 fish, suggesting that larger exotic fishes had eaten them. In one example, entire truckloads of suckers were eaten by catfish within a few days of release (Marsh and Brooks, 1989). A new strategy of stocking larger juvenile of razorback suckers has been implemented at Lake Mohave and Lake Havasu. Over 85,000 juvenile suckers have been placed in these two lakes, but factors such as predators and loss of spawning habitat, prevent young from surviving and the prospect for recovery continues to be bleak.

Flannelmouth sucker

Although abundant in higher gradient streams, the flannelmouth sucker is rare in the lower Colorado River. Although not a federal or state listed species for protection, specific populations are being monitored. In 1976, Arizona biologists collected 611 flannelmouth suckers from the Paria River, and released them in the lower Colorado River near Bullhead City. Recent studies have shown that this fish has successfully colonized a 20 mile stretch of the river downstream from Davis Dam. It is the only successful reintroduction of a native fish into the lower Colorado River.

Bonytail

Once, they were one of the most adundant medium-sized fish in the lower Colorado River, but today bonytails are extirpated or locally extinct in the lower Colorado River. These fish, as well as other native species have declined because of water development projects and habitat degradation. Although a few old individuals may still exist in Lake Havasu, recovery of this species is grim. In 1981, U.S. Fish and Wildlife biologists captured wild bonytails from Lake Mohave and transferred them to the Willow Beach National Fish Hatchery, Arizona for propagation. When hatchery raised young are stocked in refugia areas such as small ponds by themselves, bonytails did well, young grow to adults that in turn, produce many young, but when young were introduced into habitats with channel catfish, sunfish, and bass, they were quickly eaten. So recovery efforts to successfully reintroduce bonytails into the mainstream Colorado River are doubtful due to the predation issue.

Humpback chub

Although never very common in the lower Colorado River system, today humpback chubs are totally absent below the Grand Canyon. The largest population is present in the Little Colorado River drainage. Although federally listed as endangered since 1967, at this time there is no effort to reintroduce this species into the lower Colorado.

Roundtail chub

Although not found in the mainstream Colorado River, the roundtail chubs are present in small populations in portions of the Gila, Salt, and Verde Rivers in central Arizona. This fish is the most common of the three chubs in the lower Colorado River drainage. Roundtail chubs are being investigated for federal listing and protection.

Colorado pikeminnow

Although Colorado pikeminnows were abundant in the lower Colorado, Gila, and Salt River, today this species is only found in portions of the upper Colorado, Green and Yampa Rivers. Listed as an endanger species in 1976, attempts are being made to reintroduce pike minnows from fish raised at the Willow Beach National Fish Hatchery, Arizona into the San Juan and Salt Rivers, which are major tributaries to the lower Colorado. Because of federal protection issues, there are no plans to stock this fish into the main river system.

Desert pupfish

Because of loss of habitat and competition and predation from nonnative fishes, the desert pupfish has disappeared from most of its historic range. Populations survive in the wild in several areas in Salton Trough, in La Cienega de Santa Clara, Baja California, Mexico and in several agency administered refugia.

Sonoran topminnow

Although listed as an endangered species in 1967, less than a dozen natural populations of the Sonoran topminnows exist in the United States, all in Arizona. A number of refugia populations have been established in small, isolated to maintain the current numbers of fish. The good news is wild population still are found in northern Mexico.

Woundfin

Although federally listed as an endangered species, woundfins are only found in small areas along the Virgin River in Utah. These areas are threatened by water development, urbanization, and introduced species. At this time, there are no plans to re-introduce woundfin into its former range.

The future of Native Fish

"The future is grim for native fish in the lower Colorado River [Mueller and Marsh, 2002]". Native fish populations continue to decline. Habitat destruction, water reclamation projects, and introduced nonnative fishes are the main causes for their demise. Probably most damaging is the perception of native fish by the public. Outside of the environmental and conservation agencies, the public view of these fish is that they are worthless compared to the value put to the nonnative recreational fish. Four decades of research on reestablishing native fishes into their original ranges and habitats, and failed stocking programs have shown that the young of most of these species cannot survive in the current conditions of the lower Colorado River. Probably the best answer to the survival of these fishes is, where possible, setting up refugia that are free of nonnative predator fish species. Because of the use of the Colorado River water by over 30 million people in the western United States and Mexico, the river will never be returned to its original state. The fate of these native fishes is sealed.

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Ghost trail of the Carrizo Corridor

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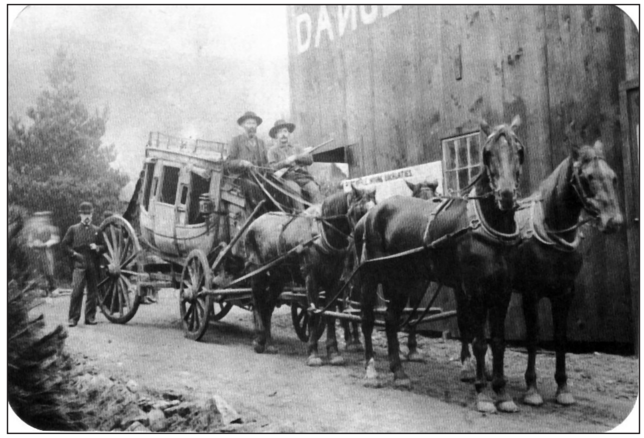
Waterman Lily Ormsby's carriage pulled to a stop at the Carrizo Station in the early morning hours of October 6, 1858. In nearly one hundred miles, this was the first significant watering hole reached after crossing the Colorado River at Fort Yuma. There was almost nothing to eat at the crude abode outpost that served as a Butterfield station house, and the little water that was available tasted of sulfur and salt, lacking essential thirst-quenching properties. Ormsby had just traveled across the worst stretch of the entire trip, some of the most rugged and inaccessible land in the scorching southern California Desert, known as the *Jornada del Muerta* [Journey of Death]. The trail was aptly described by Ormsby as "lined on both sides with carcasses of animals which had perished on the way." He was riding in a four-mule Celerity mud wagon driven by Warren Hall, which replaced the classic Concord stagecoach that had rolled out of Tipton, Missouri more than 2000 miles, nineteen days, and fourteen hours earlier. Hall served as road agent and superintendent of the Butterfield Overland Mail route between Tucson and Los Angeles.

For the last two days, since leaving Arizona, the raw-boned mules had dragged the coach through the dry, heavy, bottomless sands of the Colorado Desert. The trail led twelve miles westward into California, dipped into Mexico to skirt the great Algodones sand dunes, visited the Alamo Mocho Station, and recrossed the border near modern-day Calexico to Indian Wells. "Twenty-four miles of heavy sand riding brought us to Indian Wells," noted Ormsby, "where we found the station men recently had some difficulty with Indians who refused to let them have water for their animals." From there it veered northwest 32 miles across a barren unforeseen plain, following the old Pedro Fages path through the portal in the badlands of Carrizo Creek.

As the stage neared Carrizo, it passed several unmarked gravesites and a large group of cattle that had been abandoned by a recent wagon train since the animals were too weak to travel. "There they stood," Ormsby wrote later,

> almost living skeletons, gradually dying of thirst with water within a few miles of them. Some were standing, others lying, and others just gasping in the agonies of death – a sight almost enough to sicken the stoutest heart.

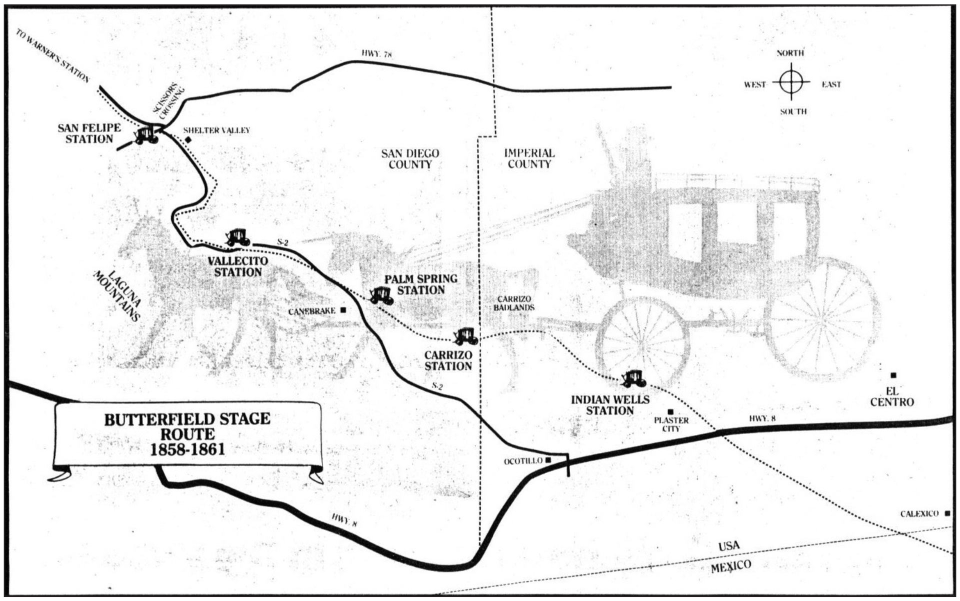
Ormsby still had more than 700 miles to go. He was traveling on the maiden westward-bound voyage of the Butterfield Overland Mail, a stage which carried mail and passengers from Missouri to San Francisco 149 years ago. Ormsby, a special correspondent for the New York Herald, booked passage on the full route, and periodically sent first-hand accounts on his journey back to New York for



all the world to read. He realized how significant the stage route would be to the future of the United States. "If the overland mail succeeds, the [Pacific] railroad and telegraph will soon follow its course; the settlements along the lone will be built up with rapidity, and our vast possessions [in the West] will be opened up to us and the world…." (Smith, 1984).

John Butterfield, a wealthy New Yorker, founder of the Butterfield and Wasson Express Company, and close friend of President Buchanan, was awarded the lucrative contract for an efficient long-distance overland mail route that stretched two-thirds of the way across the nation by the United States Postal Service in 1857. For the sum of $600,000 a year he agreed to establish the first transcontinental stage line that would carry mail and passengers on a semi-weekly basis from Missouri to San Francisco in twenty-five days or less. The gold rush of 1849 had increased California's population dramatically, and the desire for the establishment of a reliable transportation system to facilitate the movement of mail, supplies, and people was one of several badly needed government services in the new state. There were political reasons for setting up a scheduled, workable enterprise as well, not the least of which was strengthening American presence in the vast Southwest, a region formerly dominated by hostile Indians and natives of unfriendly Mexico.

The Postal Service at first considered a route from St. Louis to Salt Lake City and on to Sacramento and San Francisco, but rejected it because they were certain the route where it crossed the Rocky and Sierra Nevada mountains would be closed by snow during the winter months. And so a southern, or Oxbow, route open year 'round was chosen that led from Tipton, Missouri through Arkansas, to El Paso, Texas, then west following the 32nd parallel into the Chihuahuan Desert of western Texas and New Mexico,



the Sonoran Desert of southern Arizona, and northward through eastern San Diego County [Carrizo Corridor] across the Colorado Desert to Temecula, Los Angeles, and San Francisco. Explorers, Indians, U.S. Army troops, emigrants, and adventurous settlers had been using this historic route of travel for decades.

Butterfield invested nearly one million dollars to develop station facilities and figure out logistics along the route: 139 stations were built and supplied, 1200 horses and mules were obtained, one hundred stage coaches suitable for the conveyance of passengers as well as to the safety and security of the mails were ordered, and hundreds of trail crews, blacksmiths, cooks, station-keepers, and men of widely diverse talents were hired. The 2800-mile, 540-hour trip between Tipton and San Francisco became the longest stage ride in the world. From September 15, 1858 to March 1, 1861, Butterfield sent two eastbound coaches each Monday and Thursday from San Francisco and two westbound coaches from Tipton, Missouri traveling relentlessly, day and night, with no more than brief moments at relay stations for often poor food and no rest. He warned his passengers at the outset of a trip that "you will be traveling through Indian country and the safety of your person cannot be vouchsafed by anyone but God" (Conkling and Conkling, 1947). William Tallack, who traveled on the route in 1860 worried about "

> how far he might be able to endure a continuous ride…with no other intermission than a stoppage of about forty minutes twice a day, and a walk, from time to time, over the more difficult ground…with only such repose at night as could be obtained whilst in a sitting posture and closely wedged in by fellow-travellers and tightly-filled mail-bags.

Other notable worries were of

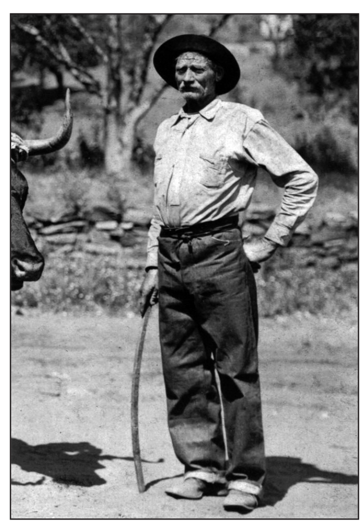
> no absolute security against Indian attack, whilst murders and robberies were known to be of constant occurrence along the line,"and "the extreme liability of vehicles to overset [overturn] during a journey through regions possessing no macadamized roads, and often only a route the most rugged and steep. (Tallack, 1865).

Carrizo Station, a handful of miles east of County Road S-2 in Anza-Borrego Desert State Park (San Diego County, township 15, range 8, section 12) (Conkling and Conkling, 1947), has long since turned to dust. All that exists of this famous saltgrass cienega landmark today is a layer of stone and charcoal, and melted adobe, the remnants of the old station foundation and walls.

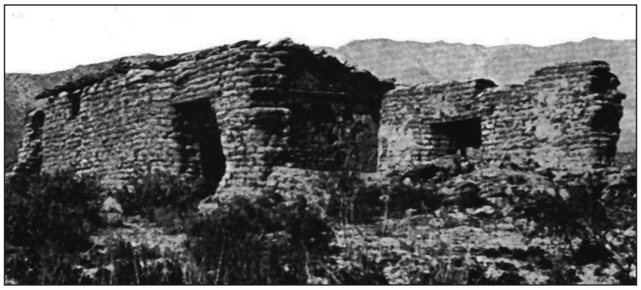
From Carrizo, the overland trail led across washboard creek beds and soft sand dunes, angling through some of the most barren untouched land in the county. The trail followed the most practical, natural pathway of Vallecito Creek nine miles north to the solitary oasis of Palm Spring. Of his progress, Ormsby could well have revisited an earlier complaint he had on his inaugural ride that the

> thumping and bumping at a rate which threatened not to leave a whole bone in my body. What with the dust and the sun…I found the day's ride quite unpleasant.

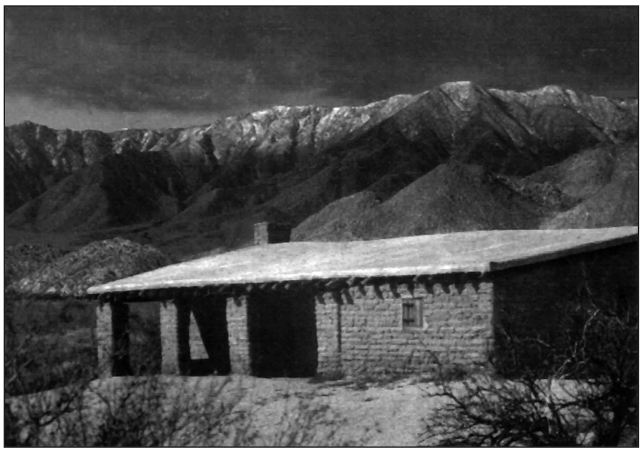
Almost destitute of vegetation, historic Palm Spring is located at the base of undulating ridges of sandstone midway between Carrizo and Vallecito (San Diego County, township 14, range 7, section 25) (Conkling and Conkling, 1947). Built by Warren Hall in 1858, the relay station of Palm Spring marks the location of the first palm tree oasis in California to have been discovered, and described in writing, by a white man [the Spanish soldier and explorer Pedro Fages mentioned it in a diary entry in 1782] (Rensch, 1955). Standing serene and alone, Palm Spring consisted of a few seepage springs hidden in a wild thicket of mes-



John McCain.



After 1861, the Vallecito Stage Station was abandoned to degrade and disintegrate travelers.



Between 1934-1936, the old Vallecito Stage Station was reconstructed as a government relief project, and donated to San Diego County. In 1939, it was set aside as a state monument. Today, it is a county park, located on County Road S-2 between Shelter Valley and Canebrake, furnished with picnic and campground facilities. It remains as a famous historic landmark of the Carrizo Corridor, dedicated to the sacrifices, struggles, and triumphs of early American overland travel.

quite and arrowweed with no palm trees. Historians argue about whether a palm grove grew at nearby Mesquite Oasis instead of at Palm Spring. Regardless, the palms were cut down for firewood by early travelers passing through. J.M. Farwell1, riding west to east in 1858 accounts that "this place takes its name from a species of palm trees which formerly grew here, and which within a few years were standing, as I saw the trunks as they lay upon the ground, and the stumps from which they were cut." Today, only a dim memory of the Palm Spring Station exists, with mounds of melted adobe hidden in the mesquite thicket to mark its location.

Beyond Palm Spring, the pathway twists in and out of Vallecito Creek through cactus, ocotillo, and creosotebush, heading to a place called by many as the most historic spot in San Diego County. "Vallecito, or Little Valley, is a beautiful green spot—a perfect oasis in the desert," wrote Ormsby.

> It is almost five miles square, surrounded by rugged timberless hills, and the green bushes and grass and hard road are a most refreshing relief from the sandy sameness of the desert.

Discovered by Fages in 1772, the cienega of Vallecito served as a welcome oasis to early California pioneers who stopped here on their way to destiny, including Kit Carson and the Army of the West led by General Stephen Watts Kearny, and the Mormon Battalion led by Lieutenant-Colonel Philip St. George Cooke. Here travelers found plenty of water, pasturage, and shade, good indications that the long arduous desert was at last behind them. By the early 1850's, Vallecito was considered the last outpost of civilization and a small sod-house was constructed by the military who occupied the site on a part-time basis following the Garra Revolt of 1851. With the establishment of the Jackass Mail and Butterfield Overland Mail, the house was rehabilitated into a first-class stage station, sturdily constructed of thick earthen walls of sod-bricks. It became one of the most important trading and supply stations on the route. Any traveler passing through would certainly stop, rest, and resupply.

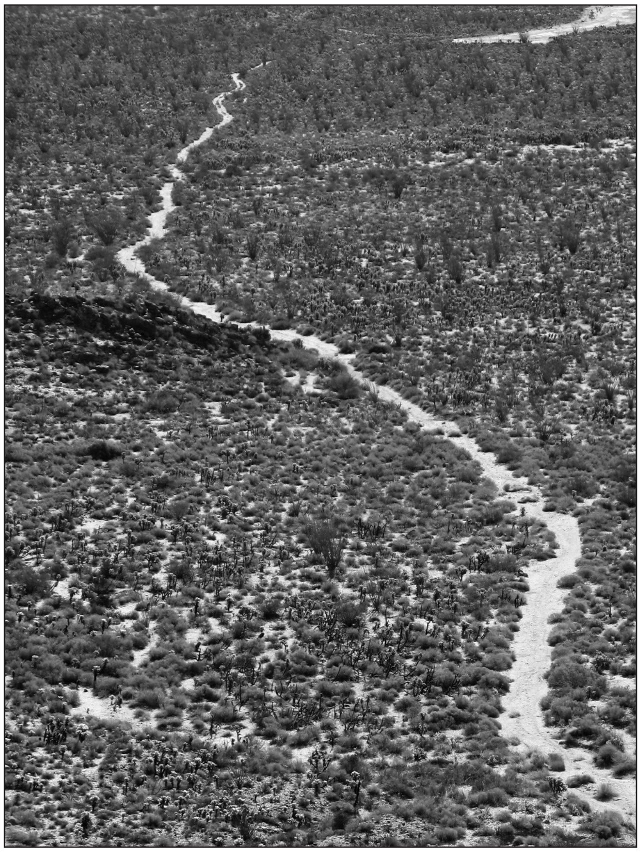
By 1860, traffic on the trail was intense. The overland stage traveled day and night on a tight, pre-arranged schedule, stopping only briefly at Vallecito to change horses, and let passengers eat a hurried meal. Beef, butter, and eggs were luxuries at desert stations, which were provisioned by local station-keepers employed by Hall. Vallecito was the only place between Warner's Ranch and Fort Yuma where a good hot meal was served. Today, it remains a major historic landmark along the Carrizo Corridor, faithfully restored according to historical information, as a memorial on one of America's most historic trails. Three rooms of the adobe-brick building are original. It is located on the roadside of County Road S-2 (San Diego County, township 14, range 6, sections 3-10) (Conkling and Conkling, 1947).

A few miles north of Vallecito, the overland trail climbed Vallecito Hill [Campbell Grade] and entered Box Canyon [Devil's Canyon], an infamous narrow gorge that knifes through a steep mountain ridge between Mason Valley and Blair Valley. To get around this barrier, Warren Hall and his trail crew built a passageway that detours precariously around a 20-foot dry fall. Ormsby described Box Canyon as

> a very narrow pass—the most wonderful on the route. It appears to have been the bed of a fierce torrent, but it was now dry. Our progress through this portion of the road was quite slow, necessarily, and it required all Mr. Hall's skill to guide our team and wagon safely through the pass; for in some places there was hardly an inch to spare. It is the most wonderful natural road I ever saw or heard of."

Box Canyon marks the end of the Carrizo Corridor and the least spoiled section of the Butterfield route in California.

At the outbreak of hostilities during the Civil War in 1861, the federal government canceled the mail contract on the southern route. Thus, one of America's boldest achievements was suspended as the Butterfield Overland Mail was transferred north to the 34th parallel to avoid going through Texas, New Mexico, and Arizona, which were contested territories and under Confederate control.



The historic pathway of the Butterfield Overland Mail is preserved cutting through the desert floor a few miles north of Vallecito. This portal through the Carrizo Corridor used to be known as the Sonoran Trail, Southern Emigrant Trail, the Mormon Battalion Trail, and Jackass Mail Route prior to 1858.

The Carrizo Corridor holds a romantic place in southwestern history. During its heyday, it served as the main pioneer route of overland travel into California, trekked by Indians, Spaniards, Mexicans, ambitious frontiersmen, emigrants, gold-hungry forty-niners, settlers, and military men. Today, the old Butterfield mail and stage route is abandoned, a silent ghost trail of forgotten gravesites, wagon wheels, bones and skeletons of livestock, old posts and markers, lost trails and oases, plus stage station ruins that litter the desert floor as the only reminders of the hardships, sacrifices, courage, and perseverance of travelers in the epoch-making early settlement and development of the American West.

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Notes:

1. Account of Mr. J. M. Farwell printed in the Daily Alta California of San Francisco, November 6, 1858.

Pliocene angiosperm hardwoods and recycled Cretaceous palynoflora of the ancestral Colorado River, Anza-Borrego Desert State Park, California: a review

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Abstract

The Vallecito-Fish Creek Basin of Anza-Borrego Desert State Park includes a continuous time-stratigraphic sequence of lower Pliocene pro-delta/delta-front (Coyote Mountain Clays and Yuha Formation) and delta-plain (Palm Spring Formation) sediments sourced from the Colorado Plateau and deposited by the ancestral Colorado River. Exposed along Fish Creek Wash, these fine-grained, extralimital sediments yield locally-derived silicified angiosperm fossil wood (dicotyledons and monocotyledons) of the Carrizo Local Flora within the Palm Spring Formation, and reworked Cretaceous pollen zoned throughout the stratigraphic section. Based on tracheid cell structure, seven families of fossil wood are recognized in the paleoflora, representing eleven taxa. Eight are new for the area. Families include the Lauraceae, Salicaceae, Oleaceae, Hippocastanaceae, Arecaeae, Juglandaceae, and the Cupressaceae. Climate data inferred from this riparian association and on tree-ring growth analyses suggest that the paleoclimate of the northern Gulf of California was more temperate and wetter than now with winter rainfall dominant. The presence of Cretaceous pollen and microscopic foraminifers in time- constrained Pliocene sediments of Anza-Borrego is significant and contributes to the understanding of the biogeographic history of Anza-Borrego. The stratigraphic occurrence and distribution of the microfossils *Proteacidites, Mancicorpus*, and *Aquilapollenites* along Fish Creek Wash, for example, indicates that erosion of their source rock, the Mancos Shale, in the southern part of the Colorado Plateau began about 4.5 Ma, and in the northern part of the plateau at about 3.9 Ma. Thus, the presence of these microfossils testifies that the erosional events on the Colorado Plateau, including the extensive down-cutting of the Grand Staircase and Grand Canyon and the achievement of a through-flowing drainage by the perennial Colorado River to the northern Gulf of California, is a relatively young geomorphologic phenomenon.

Introduction

Historically, relatively little is known of the silicified woods from the Colorado Desert. Paleobotanic evidence is scanty and discussed matter-of-factly by early investigators with less attention than to invertebrate and vertebrate fossils. William Phipps Blake is perhaps the first to note the occurrence of silicified wood in the region. "Some small specimens of silicified wood were found on the surface" near the location of Salt Creek south of the San Felipe Hills (Blake in Williamson, 1853). Farther south, between Big Lagoon and Carrizo Creek, he

> secured many fine specimens of silicified wood. It occurs in great abundance lying loose on the surface, and in places, tons of it could be collected from an area of one or two acres.

Later, in the sandstone hills on each side of the dry arroyo of Carrizo Creek, he

> found several masses of silicified wood, one of them so large that I was unable to bring it away; smaller specimens were, however, obtained. In these the grain of the wood was well preserved; and all the small knots, and the rings of annual growth, were remarkably distinct. One of the specimens, in which the rings were perfectly shown on the ends, appeared to have been flattened by pressure, so as to produce ellipticity in all the rings. The specimens picked up on the desert slope were of similar formation and color. Wherever these specimens have lain out upon the surface exposed to the continued action of the loose sand, they have become worn and polished, so that the grain of the former wood is more distinctly displayed by which their beauty is greatly increased" (Blake in Williamson, 1853).

Although silicified woods have been reported in the literature since 1853, the woods remained erroneously identified until 1988. Blake (in Williamson, 1853) first observed

> Fine specimens of silicified wood were found to be abundant and of various sizes, from one or two inches in length to as many feet. They are generally of a brown color, and retain all the appearance of wood; the grain and knots show distinctly, and resemble the wood of the mezquit [mesquite].

Dibblee (1954) reported that "Fragments of silicified wood, chiefly ironwood [desert ironwood], are common throughout the formation [Palm Spring Formation of Woodring, 1931]", a hitherto unrecognized assumption at the time, later shared by Woodard (1963) and many other investigators. In 1984 and 1996, Dibblee revised his earlier assessment, siding with Blake that "In many places this formation contains fragments of dark gray silicified iron wood (Mesquite) with grain well preserved". Pinault (1984) attributed the wood found in sediments exposed south of the Volcanic Hills and west of the Coyote Mountains to be generally oak (*Quercus* spp.), without supplying any supportive evidence.

Because silicified wood was a neglected resource of study, Remeika et al. (1988) were the first researchers to provide a definitive identification of the wood, describing three dicotyledon tree genera as part of the lower Pliocene paleoflora recovered from the Palm Spring Formation. The most common fossil dicots include ancestral forms of diffuse-porous California bay-laurel, *Umbellularia salicifolia* (Lauraceae), and semi-ring-porous California walnut, *Juglans pseudomorpha* (Juglandaceae), both showing strong affinities to the modern species *Umbellularia californica* and *Juglans californica*, respectively. In addition, the willow family Salicaceae was also introduced, with specimens of diffuse-porous *Populus* and *Salix*. Remeika and Fleming (1995) recognized additional taxa with affinities to extant *Persea podadenia* (avocado), *Fraxinus oregona* (ash), and *Aesculus californica* (buckeye). No samples of fossilized desert ironwood, mesquite, or oak have been found or identified from the Palm Spring Formation, although many lapidary rock shops in Arizona and the desert southwest sell silicified wood collected on terrace gravels south of Yuma (Nations and Gauna, 1998) under the pseudoname of "petrified desert ironwood".

Figure 1. The Palm Spring Formation is characterized in outcrop by a multistory sequence of alternating reddish-brown claystones and siltstones and massive, fine-grained marker beds of quartz arenitic sandstones. The sandstones form conspicuous strike ridges and yield an abundance of concretions and fossil woods. View looking north from Fish Creek Wash towards the Vallecito Mountains.

Stratigraphy and age

The Vallecito-Fish Creek Basin is a 373 km² syn-rift half-graben, structurally depressed as part of the geodynamic Colorado Desert Breakaway Margin of southern Anza-Borrego Desert State Park, California. In response to the kinematics of lithospheric extension, this basin contains one of the most complete stratigraphic sequences associated with the lower ancestral Colorado River. The sedimentation record is represented by a vast delta system built upon rhythmitic delta-front deposits of the Coyote Mountain Clays and Yuha Formation, and meandering river delta-plain deposits of the Palm Spring Formation that prograded and matured along the extensional northern Gulf of California embayment during the Pliocene. In the field, sediments are vertically stacked, and generally dip west in response to younger dip-slip activity along the basin-bounding Tierra Blanca Mountains Frontal Fault (Figure 1).

The Coyote Mountain Clays (Hanna, 1926) represent a thick accumulation of distal bar deposits sourced from distributary mouth bars on the delta-front subenvironment. The lower half consists of sand- and clay-dominated aggradational to progradational deposits of a low to moderate energy outer- to inner-shelf marine environment, keeping pace with relative sea-level rise. The depth of the offshore is constrained by the ubiquitous presence of an in situ upper bathyal benthic foraminifer fauna (e.g., *Bolivina interjuncta, Planulina ornata, Cassidulina subglobosa*), indicative of paleodepths of at least 100 m (Ingle, 1974). The upper half is composed on a monotonous silty claystone facies of mixed affinity, interpreted to be deltaic foreset deposits made up of pro-delta silty clays and coarse sands. The abundance of the benthic foraminifer *Hanzawaia nitidula* suggests relatively warm, marginal-marine waters (Crouch and Poag, 1979). Up section, there is a thick stratigraphic sequence of unfossiliferous, rhythmically alternating couplets of fine-grained gypsiferous claystone and siltstone. The progradational stacking of distal bar rhythmites is laterally persistent and a distinctive feature on the delta-front environment of the Coyote Mountain Clays. With

Figure 2. The untouched, remote, arid, and uninhabited terrain of the Palm Spring Formation along Fish Creek Wash preserves and protects a high content and quality of surface silicified woods and other fossil materials. View looking west from Loop Wash across Blackwood Basin.

time lines crossing lithofacies boundaries, the strata record a forced regression into the fluviodeltaic Yuha Formation.

The Yuha Formation (Remeika, 1998) consists of terrigenous sediments supplied by the river into a tidally-dominated, delta-front environment of the northernmost Gulf of California. In profile, it is distinguished by the infrequent presence of siliciclastic-carbonate, epifaunal oyster-anomiid-dominated coquina beds that occur in upward-coarsening cycles capped by arenites and/or coquina, separated by thick stratigraphic intervals of rhythmically-alternating wafer-thin couplets (1-10 cm thick) of gypsiferous claystone and siltstone laminations typical of the Coyote Mountain Clays. The sharp-based, channelized arenites and associated shell coquinas represent lateral accretionary deposits on a shallow-marine, mixed wave- and river-dominated subenvironment (tidal flat) across the entire head of the northern gulf, and today form resistant depositional landmarks (ledges, strike ridges, cuestas) along Fish Creek Wash. The shell concentrations are the coarsest sediment fraction on the delta. They are composed of death-assemblages preserved in a mixed siliciclastic-carbonate cemented quartz sand matrix with subordinate lithics. They are 30 cm to 5 m thick, discontinuous, and dominated by disarticulated valves and fragments of the small plicate oyster *Dendostrea vespertina*, and lesser *Anomia subcostata* (clam), and *Argopecten deserti* (scallop). The Yuha Formation interfingers with nonmarine arenites of the Palm Spring Formation.

The conspicuous Palm Spring Formation (Woodring, 1931) is an extensively-exposed delta-plain deposit, renowned for its abundant ichnofossils (Remeika, 2001), mammalian fossils (Downs and White, 1968), silicified woods (Remeika, et al., 1988), and reworked Cretaceous pollen (Fleming, 1994). Sediments include extraregional pale pink to cream-colored, fine to very fine-grained quartzitic arenites and gravelly sandstones with a common vertical pattern of zonation, punctuated stratigraphically by subordinate brown-colored, ripple-laminated, gypsiferous overbank claystones and siltstones. The high percentage of arenites resulted from current-deposited bedload that contain cross-bedding structures and textural features indicative of an outgrowth from a highly meandering fluvial river system. The result is relatively persistent, uniform, multilateral arenites that are distinctive in outcrop, oftentimes forming prominent strike ridges (Figure 2).

This distinctive suite of lithostratigraphic Colorado River deposits crosses several magnetic zones of both normal and reversed polarity. A normal remnant magnetic signature is documented along Fish Creek Wash between Fish Creek Canyon and Arroyo Seco del Diablo (Opdyke et al., 1977; Johnson et al., 1983). This signature is interpreted to represent the Gauss normal polarity magnetochron (C2An) (3.550–2.581 Ma). Down section, the Gilbert reversed magnetochron (C3r-C2Ar) has been determined blanketing the Yuha-Palm Spring transition, placing the base of the Palm Spring Formation at between 4.08–3.95 Ma. Two short normal intervals matching the Cochiti and Nunivak magnetosubchrons (J.C. Liddicoat, written comm., 2001) occur lower in the section in the Coyote Mountain Clays. This determination is consistent with a lower Pliocene age assignment for the deltaic package along Fish Creek Wash. The top of the Palm Spring Formation roughly correlates with the Gauss/Matuyama boundary at 2.58 Ma. To help constrain this age further, two key time-stratigraphic markers are provided by interbedded volcanic ash beds outcropping above the Gauss/Matuyama boundary west of Fish Creek Wash. The lower ash yields a fission-track date of 2.3 Ma (Johnson et al., 1983). The upper ash, fortuitously discovered, is considered to be equivalent to the Huckleberry Ridge Ash (2.01 Ma) given its petrographic features and glass composition.

Carrizo Local Fauna

Today, the Carrizo Local Flora (Remeika, 1994, 2006) contains well-preserved silicified and calcified fossil dicotyledon and monocotyledon wetland woods and parallochthonous macrodetritus recovered from massive Colorado River-derived deltaic deposits of the Palm Spring Formation within the Vallecito-Fish Creek Basin (Remeika, 1991, 1997). Based on tracheid cell structure, seven families of fossil wood are recognized. These include the Lauraceae, Salicaceae, Oleaceae, Hippocastanaceae, Arecaeae, Juglandaceae, and the Cupressaceae, members of which have been anatomically described, identified and analyzed (in prior references), and summarized below.

The Palm Spring Formation is named for redbed sediments outcropping in and around the Carrizo Stage Sta-

Figure 3. Fragments of silicified fossil wood litter the desert floor along portions of Fish Creek Wash, eroding out of the Palm Spring Formation.

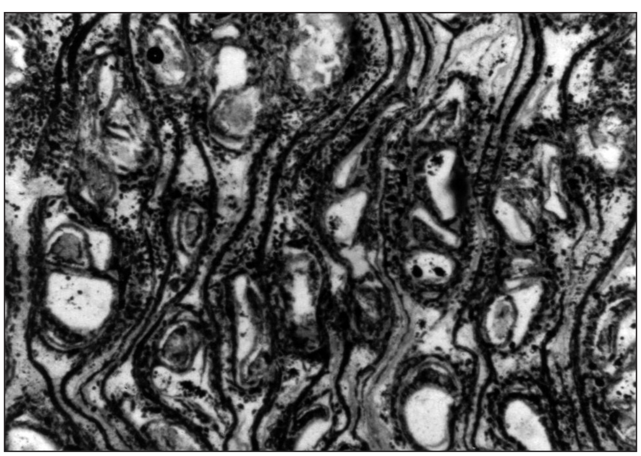
tion on Carrizo Creek (Woodring, 1931). Along Fish Creek Wash, these redbeds are exposed as multistoried, high-energy pale pink to cream-colored, fine to very fine-grained channel sandstones (arenites), point bars and natural levees that are dominated by woody debris. In turn, the sandstones are capped by low-energy reddish-brown overbank clays and silts that yield reworked Cretaceous pollen. Based on field evidence, the majority of woods grew on a wide deltaic apron, characterized by high ground-water levels that existed on the mouth of the perennial Colorado River across the northern embayment of the Gulf of California during the lower Pliocene. Seasonally, the small- to medium-diameter woody debris was uprooted from growth position on natural levee deposits by recurrent inundation from the river and transported bayward as driftwood by suspension-load/bed-load meandering floodwaters. The wood eventually became water-logged, sank, and was differentially buried, petrified and preserved, oriented parallel to flow direction, in a subaqueous environment of the deltaic-plain sediment-water interface. Over the course of millennia, silica and calcium carbonate minerals replaced the original woody material, retaining the internal structure (Figure 3).

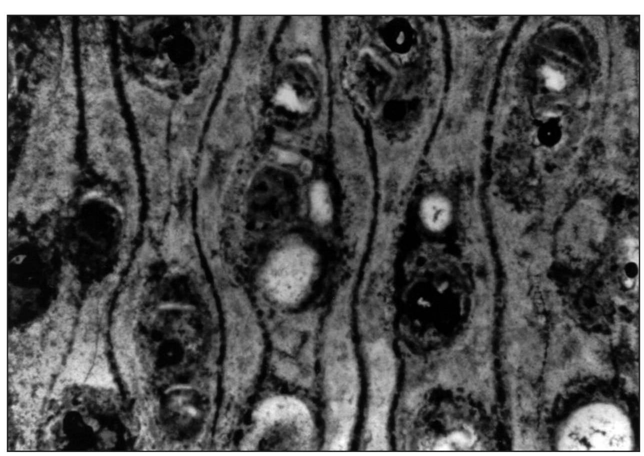
The anatomical structure of the majority of the fossil woods provides a unique paleoenvironmental archive indicating a species rich, mixed dicotyledonous angiosperm flora typical of the Madro-Tertiary Geoflora of Axelrod (1950, 1958), with a complex standard tree regime that has affinities to three distinct tree associations. The three associations, referable to extant genera based on the presence of morphologically and anatomically identical leaf impressions, seeds, nuts, and woods, are:

(1) mixed-evergreen fossil hardwoods of the California Woodland Element (Axelrod, 1950) locally featuring California bay-laurel *Umbellularia salicifolia*, California walnut *Juglans pseudomorpha*, Oregon ash *Fraxinus caudata*, black cottonwood *Populus alexanderi*, willow *Salix gooddingii*, and California buckeye *Aesculus sp*. Descendants make up the hydrophytic mixed-evergreen understory element of the coastal redwood community along the central California coastline at Big Sur (Remeika, 2006).

(2) members of the Sierra Madrean Woodland Element (Axelrod, 1950) with the fossil fan palm (*Washingtonia* sp.), avocado (*Persea coalingensis*), and black cottonwood (*Populus alexanderi*), descendants of which today survive in the semi-arid and arid lands of southwestern U.S. and subtropical and tropical Mexico.

(3) the Conifer Woodland Element (Axelrod, 1950) with one softwood representative, *Pineoxylon* sp.; whether it is a





Figures 4 and 5. Transverse sections. Umbellularia salicifolia *(Lesquereux) Axelrod 80X IVCM 800-04. Photomicrographs show diffuse porosity with numerous multiple pores. Associated with the pores are deformed structures and zones of partially collapsed parenchyma cells and pores. Note that the parenchyma forms a thickened sheath around the pores, typical of bay-laurel.*

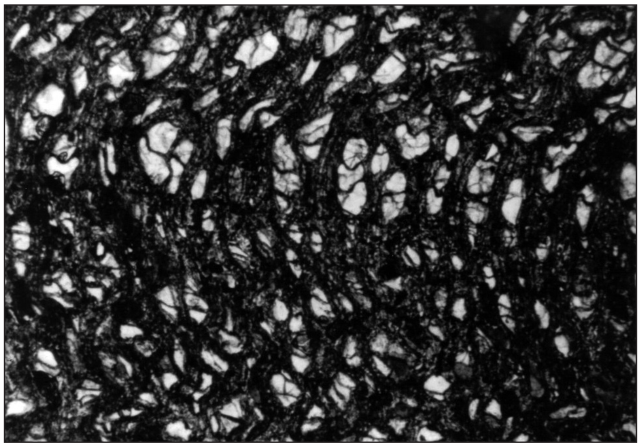


Figure 6. Transverse section. Salicaceae (may represent Populus *sp.) 80X IVCM 803-06. Typical example of fossil wood structure identified as willow family showing semi-diffuse porosity between springwood and summerwood. This sample shows considerable deformation of the tissue structure.*

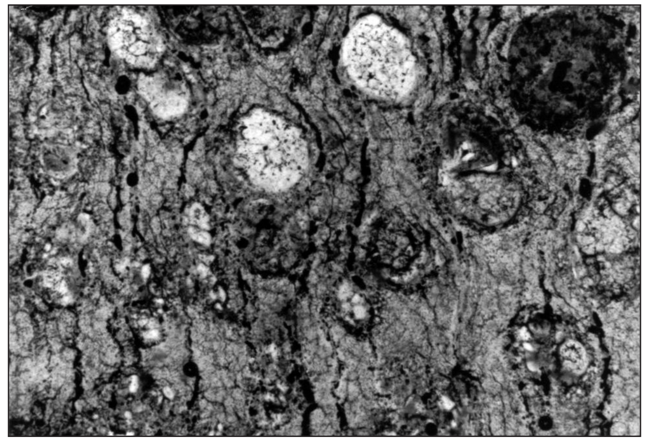


Figure 7. Transverse section. Juglans pseudomorpha *Axelrod 80X IVCM 804-01. Photomicrograph shows semi-ring porosity. The dramatic difference in pore size between early springwood and late summerwood is readily apparent.*

cedar or juniper remains indeterminate.

The accurate identification of fossil wood includes the recognition of gross and minute anatomical features that are specific to a given kind of wood and which will identify its botanical source. Systematic examination of the wood includes thin-section preparation and photomicrography (selected Figures 4 - 7) of the well-preserved details of the wood. Based on the macroscopic appearance of the wood, cell dimensions, and frequency of cell distribution in the growth ring on end-grain surfaces, seven families are currently recognized, representing six species. Among the taxa described in Remeika (1994), five are new for the area. Also included are well-preserved anatomic leaf impressions of *Populus trichocarpa* (black cottonwood) and *Salix gooddingii* (Dudley willow), and herbaceous monocot palm fronds that have a suite of diagnostic features, including attached petiole fragments, most similar in appearance to the California fan palm *Washingtonia filifera*. Also, associated fossil wood material is indistinguishable from the genus *Washingtonia* sp., based on anatomical pore structure morphology, and does not resemble *Sabal* palm which this taxon has been assigned erroneously to in the past (see Table 1).

Recycled time travelers

Reworked microfossils from the Colorado Plateau include ten taxa of Cretaceous foraminifera (Merriam and Bandy, 1965), plant pollen, spores, dinoflagellates (Fleming, 1994; Fleming and Remeika, 1994; Remeika and Fleming, 1995), and Eocene pollen *Pistillipollenites* reworked from the Green River Formation of Utah-Wyoming (F. Fleming, pers. comm., 1993). Reworked palynological assemblages from the Palm Spring Formation are sparse, with most samples barren of palynomorphs. Marginally adequate assemblages were recovered from the upper half of the Yuha Formation (Remeika, 1988) and the lower half of the Palm Spring Formation. Two palynomorph categories are potentially useful as diagnostic stratigraphic markers and

Table 1

Distribution of Anza-Borrego fossil taxa comparable to allied living species and occurrence (■) in the Madro-Tertiary Geoflora. 1= California Woodland Element; 2= Sierra Madrean Woodland Element; 3= Conifer Woodland Element. ■= new to Anza-Borrego.

FOSSIL WOODS	ALLIED LIVING WOODS	MADRO-TERTIARY GEOFLORA		
		1	2	3
Arecaceae				
Gen. et. sp. indet.	monocot (palm)		••	
Washingtonia sp.	*Washingtonia filifera* (Linden) Wendland (California fan palm)		••	
Hippocastanaceae				
Aesculus sp.	*Aesculus californica* (Spach) Nuttall (California buckeye)	••		
Juglandaceae				
Juglans pseudomorpha Condit	*Juglans californica* S. Wats. (California walnut)	•		
Lauraceae				
Persea coalingensis Axelrod	*Persea podadenia* Black (avocado)		••	
Umbellularia salicifolia Axelrod	*Umbellularia californica* (Hook and Arn.) Nuttall (California bay-laurel)	•		
Oleaceae				
Fraxinus caudata Dorf	*Fraxinus oregona* Nuttall (Oregon ash)	••		
Salicaceae				
Populus sp. indet.	*Populus* sp. (cottonwood)	•	•	
cf. *Populus alexanderi* Dorf	*Populus trichocarpa* Torr. And Gray (black cottonwood)	••	••	
Salix sp. indet.	*Salix* sp. (willow)	•		
Salix gooddingii Ball	*Salix gooddingii* Ball (Dudley willow)	••		
Cupressaceae				
Pineoxylon sp.	softwood (cedar or juniper) ?			••

for making paleoclimatic interpretations. The first category is algal. Two samples from the lowermost Palm Spring Formation (samples D7697-83TC38 and D7697-83TC39 in Remeika and Fleming, 1995) produced assemblages overwhelmingly dominated by coenobia of *Pediastrum* spp., with *Scenedesmus* spp. as a subordinate. The second category is reworked Cretaceous palynomorphs which compose a significant part of the assemblage. The presence of reworked Cretaceous triprojectate pollen, spores, and dinoflagellates in these sediments indicates that they were transported from the Colorado Plateau, via the ancestral Colorado River, and redeposited in the lower Pliocene deltaic sediments outcropping throughout the Fish Creek and Carrizo Badlands. These fossils include *Proteacidites* spp., *Aquilapollenites* spp., *Mancicorpus* spp., *Tricolpites interangulus, Pistillipollenites* sp. cf. *P. mcgregorii, Corollina* sp., *Appendicisporites* sp., *Cicastricosisporites* sp., *Camarazonosporites* sp., *Dinogymnium* sp., and *Palaeohystrichophora infusoriodides*.

The distinctive Cretaceous genera *Proteacidites* spp. (Coniacian to Maastrichtian), and *Aquilapollenites* spp. (Campanian to Maastrichtian), including *Mancicorpus* spp., which is closely related to *Aquilapollenites* spp., have known restricted stratigraphic and paleobiogeographic distributions from the upper Cretaceous Mancos Shale of the Western Interior of North America. The Mancos Shale and its stratigraphic equivalents outcrop in and around Capital Reef National Park, Grand Staircase-Escalante National Monument, and near the town of Grand Junction, Colorado, are known to contain abundant Cretaceous palynomorphs (Thompson, 1972; Franczyk et al., 1990; Cushman and Nichols, 1992). Other interesting reworked fossils discovered in the Palm Spring Formation include silicified freshwater gastropod specimens of *Goniobasis tenera* transported from *Turritella*-agate beds in the Laney Member of the Green River Formation (S. Good, written comm., 1996).

Paleoclimate implications

Microscopic study of the fossil wood, including its internal structure, and ring counts, yields important clues to the climate of the Colorado Desert during the Pliocene. Many bay- laurel ring counts, for example, show growth rates ranging from 1 to 5 rings per cm (3/8 inch), with an average range of 2–5 rings. This reinforces the hypothesis that there was no rain shadow, and that the region was subject to the maritime influence (coastal fog and overcast skies) of both the ancestral northern Gulf of California Imperial sea and the Pacific Ocean. In addition, the Plio-Pleistocene paleoclimate inferred from the riparian wetland association suggests that the Colorado Desert was also more temperate than now with winter rainfall dominant, clearly predating the dramatic uplift and breakaway of the Peninsular Ranges, and evolution of a dynamic rain shadow desert setting for Anza-Borrego.

The chronologic distribution of reworked fossils incorporated within the deposits along Fish Creek Wash is compelling evidence that erosion of Cretaceous rocks in the southern part of the Colorado Plateau that contain *Proteacidites*, but lack *Mancicorpus* and *Aquilapollenites*, began about 4.5 Ma. *Proteacidites* first appears in the marine-deltaic Yuha Formation during this time. The presence of *Mancicorpus* and *Aquilapollenites*, from the northern part of the Colorado Plateau, in the Palm Spring Formation, occur about 3.9 Ma. This palynostratigraphic zonation indicates that the rapid pace of Colorado River drainage integration and erosion of Cretaceous rocks from the northern section of the Colorado Plateau commenced during the Pliocene, rather than earlier in the Tertiary. Erosion and transportation of a large volume of sedimentary rock from the plateau to the northernmost Gulf of California implies that there were correspondingly significantly higher precipitation levels throughout the southwestern United States than today (Thompson, 1991), suggesting a much wetter climate supported by the ancient woodland, and an adequate flow of runoff necessary for the systematic overflow of lake-filled interior basins by spillover from the mouth of the Grand Canyon (House et al., 2005). The presence of small foraminifers in the stratigraphic package of the Vallecito-Fish Creek Basin testifies that extensive downcutting of the Grand Staircase and Grand Canyon, and the achievement of a through-flowing drainage by the perennial Colorado River to the northern Gulf of California, is a relatively young geomorphologic phenomenon (Remeika and Fleming, 1995; Bell and House, 2005).

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A Late Miocene record of a fossil gulper shark (Family Centrophoridae) from the Mud Hills Member of the Deguyos Formation of the Imperial Group, Lycium Wash, Fish Creek Basin, Anza Borrego Desert State Park, San Diego County, California

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As part of the paleontological resource management program at Anza Borrego Desert State Park, volunteers, seasonal workers, and staff under the supervision of park paleontologist George Jefferson have been surveying large areas of the park for fossils. Recent field work in the Lycium Creek in the Fish Creek Basin of the park in rocks assigned to the Mud Hills Member of the Deguyos Formation of the Imperial Group has turned up two types of fossil shark teeth.

In this area the Mud Hill Member of the Deguyos Formation, which dates from 4.8 to 5.1 million years, consists of gray marine clays that chronicle the Colorado River entering the Salton Trough. Recent geologic work on this rock indicates that these sediments were deposited in prodelta environment.

The first tooth (Fig. 1), found by volunteer Jimmy Smith, appears to be from a gulper shark (Family Centrophoridae). Gulper sharks, related to the modern dogfish (genus *Squalus*), are mainly a group of deepwater marine bottom-dwelling species. Today, these smallish sharks are found almost worldwide from cold temperate to tropical seas, except in the northeast Pacific (this area) and high latitudes. To date, about 16 species have been placed in the genera *Centrophorus* and *Deania*. Although one species has been recorded in depths as shallow as 50 m, most are found in between 1000–1500 m.

The second tooth (Fig. 2), collected by volunteer Judy Smith, is not identifiable at this time. The two shark teeth represent the first records of fossil sharks from the Mud Hills Member of the Deguyos Formation of the Imperial Group.

Figure 1. Gulper shark tooth in matrix.

Figure 2. Unidentified shark tooth in matrix.

Pleistocene geology and paleontology of the Colorado River Delta at Golfo de Santa Clara, Sonora, Mexico

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The El Golfo Badlands – geographic and tectonic setting

Exposures of ancient Colorado River sediments, known as the El Golfo Badlands, are located near the small fishing community of Golfo de Santa Clara (El Golfo), in northwestern Sonora, Mexico. The northwest-southeast oriented El Golfo Badlands are exposed along a narrow strip of coast on the Sonoran-side of the upper Gulf of California (Figure 1). The Badlands are bounded to the northeast by the expansive Gran Desierto de Altar; to the southwest by coastal dunes and the gulf shoreline; to the northwest by the silty, salt flats of the modern Colorado River delta; and by faulting on the southeast flank.

The southeast faults juxtapose late Pleistocene (~125 ka; Colletta and Ortlieb, 1984), near-shore, marine sediments with the older Colorado River delta sediments. Numerous faults dissect the exposures with a pronounced northwest-southeast strike and a subset of north-south oriented tension gashes. The Cerro Prieto fault forms the principal active fault paralleling the coastal exposures and lies just off shore (Merriam, 1965; Colletta and Ortlieb, 1984).

The region occurs within an active trans-rift tectonic boundary (Kovach et al., 1962; Biehler et al., 1964; Elders et al., 1972, Pacheco et al, 2006). The El Golfo beds are caught between two giant plates, the Pacific Plate on the western side and North American Plate on the eastern side - grinding and spreading about 5 cm/yr (Winker and Kidwell, 1986).

Recent work by Pacheco et al. (2006) suggests that the crustal block containing the El Golfo area and much of the Altar Desert has rotated basinward (northeast to southwest) along a detachment fault as widening of the upper gulf proceeds. The sediments have been domed into a broad antiform that is broken by numerous north-south strike slip faults, small flexures, and slight northeasterly-rotated blocks. The rotation and uplift have been occurring at least since the late Pleistocene because near-shore marine sediments and the 125 ka sea-level high stand are deformed (Colletta and Ortlieb, 1984). It is this constant movement, uplift, and erosion that have resulted in the development of the Badlands.

History of fossil collecting

Sedimentary rocks that have formed Badlands to the north, east, and south of El Golfo contain a vast resource of interest to vertebrate paleontologists. The Badlands cover

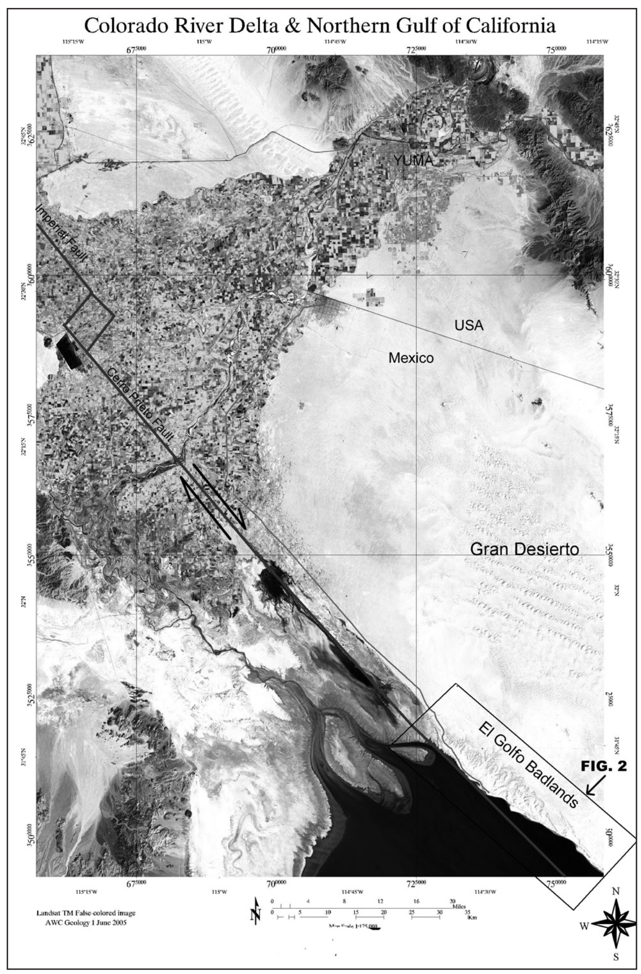

Figure 1. Vicinity map.

approximately 255 sq km (99 sq mi). Even though only a small portion of these Badlands has been prospected, vertebrate fossils with similar preservation have been recovered from a wide area.

The first collection of fossils from the vicinity of El Golfo by an institution took place in 1938. Chester Stock of the California Institute of Technology sent two collectors (J. Dougherty and A. Drescher). They recovered two-dozen vertebrate specimens representing giant tortoise, sloth, horse, camel, deer, and mammoth, and some petrified wood. Since that time, eight other institutions have acquired small and large collections. The University of Arizona, Tucson, collected over 50 fossil vertebrate specimens in the late 1960s and 1970s from various canyons north, east, and south of the village of El Golfo, including the most southerly exposures near El Tornillal. Another 63 specimens were recovered by the University of California, Berkeley in 1970 from one of the first canyons encountered as you approach the northwestern edge of the Badlands. In the same year, eleven vertebrate fossils were collected from "Shaw Canyon" (northeast of El Golfo) by the University of California, Riverside. In 1983, the San Diego Society of Natural History received a donation of about 30 specimens collected from "Trash Canyon" (east of El Golfo) in the 1960s and 1970s by amateur collectors, Foss and Emma Corley. Among these items was the only North American record of giant anteater. Between 1976 and 1979, a large collection was accumulated by students under the direction of professors Tom Lubin and Keith Green of Cypress College, California. It was this collection, along with specimens from the Natural History Museum of Los Angeles County (LACM), which formed the core of material (over 200 identifiable specimens) leading to the completion of a Masters Thesis on the El Golfo Local Fauna by Christopher A. Shaw in 1981. The LACM collections were made initially by Theodore Downs and Harley Garbani in 1963, and later by Chris, Harley, Antonia Tejada-Flores and others between 1979 and 1981.

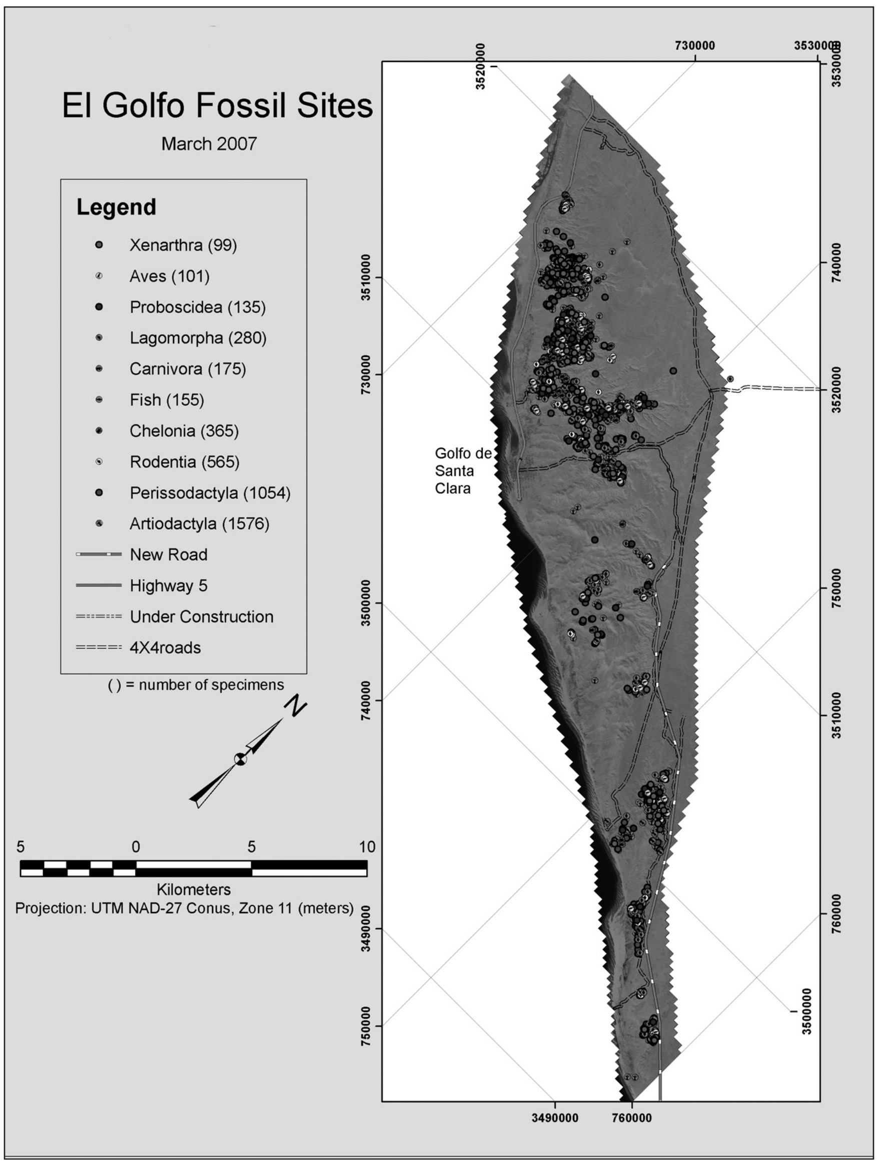

Figure 2. El Golfo fossil sites.

By far the largest and most comprehensive data and fossil collections in the vicinity have been made by students and staff of Arizona Western College, in Yuma under the supervision of Fred Croxen, David Sussman, and Robert Predmore. For the past 12 years there has been a cooperative effort between the Biosphere Reserve in Sonora, the George C. Page Museum in Los Angeles, and Arizona Western College (AWC) in Yuma, to systematically prospect, recover, and record fossil localities within these vast badlands and canyons. This has resulted in the recovery of over 5,200 fossil vertebrate specimens. Even so, less than 20 percent of these deposits have been explored and documented to date (Figure 2).

The contribution of amateur collectors has been immense. Several taxa of mammals are only known from fossils collected by a visitor camping on the beach and hiking through the canyons. Examples include a single metacarpal of a giant anteater, a lower jaw of the only capybara found, and the best dental material of a wolf and a ground sloth. Most of the fossils recovered by amateur collectors, especially those of rare taxa, have been transferred into institutional repositories.

Geology

The sediments comprising the El

Golfo Badlands are undifferentiated and have not been formerly assigned a stratigraphic name. There have been few previous geologic investigations in the area. The exposed sediments have a Colorado River delta source and range in thickness from about 500 to 2500 meters (Pacheco et al., 2006). It is estimated that a maximum of 200 meters are exposed along the coast. We informally refer to them as the El Golfo beds in this paper. However R. Dorsey (pers. comm., 2006) has suggested that based upon sedimentation patterns and structures the sediments could be a younger facies of the Pliocene age Arroyo Diablo Formation of the Vallecito Creek–Fish Creek Basin (VCFCB) and the Borrego–San Felipe Basin in Anza-Borrego Desert State Park and Ocotillo Wells State Vehicular Recreation Area of southeastern California. This would suggest that following the tectonic isolation of these paleo-basins exposed along the western Salton Trough from the Colorado River delta, deposition of deltaic sediments persisted into the early and middle Pleistocene as the Colorado River source on the east side of the Trough shifted to the southeast along the San Andreas, Imperial, and Cerro Prieto faults.

The El Golfo beds are fluvio-deltaic, massive sandy siltstones and claystones interbedded with conglomerates and indurated sandstones (Merriam, 1965; Colletta and Ortlieb, 1984, Shaw and McDonald, 1987). The siltstones are tan to yellow and interbedded with fine to medium sand. The claystones are tan to brown, up to 4 meters thick, and lens-like, sometimes extending more than a kilometer in lateral exposure. Thin beds of tan to light gray, pebble conglomerates and lenses of tan, indurated sandstones are found in varying frequency within these massive siltstones.

We interpret the massive siltstones and claystones as overbank flood deposits and the lenses of sandstone and conglomerates as channel-fill and crevasse-splay deposits. The basal portions of the channel lithofacies commonly contain rip-up clay balls and convolute lamination that grades upward into trough cross-stratification. These lithofacies are similar in many ways to the exposures seen in the VCFCB and the Borrego–San Felipe Basin. Downs (1963) found these sediments to be lithologically similar to the lower Palm Spring Formation recently revised to the Arroyo Diablo Formation of the Palm Spring Group in southeastern California by Cassiliano (2002; = Palm Spring Formation of Woodring, 1931).

The clast composition of the conglomerates indicates a dominant Colorado Plateau and lower Colorado River corridor source. Pebbles consisting of black chert, brown to tan quartzite, and gray cherty limestone are common lithologies. Some of these clasts contain reworked late Paleozoic invertebrate marine fossils (i.e. silicic sponges, bryozoans, and brachiopods). Clasts of locally derived granites, volcanics, and gneiss or schist can comprise as much as 50% of the pebble lithologies.

Exposures at most localities show no extreme dip, with a range of between zero and eight degrees. Dips can locally increase to as much as forty five degrees near fault zones.

The sediments are weakly indurated with occasional well-cemented sandstone lenses. Diagenesis of the sediments suggests cementation by low-temperature fluids associated with fractures or differential permeabilities (Jennings and Thompson, 1986). Secondary gypsum fills tension gashes or is concentrated in units lying above siltstone or claystone lithology. Davies (2000) found that ferruginous-cemented zones were aligned with faults and fractures often associated with pipe-like vent structures. Selective preservation of petrified wood is often found near these vents.

Large fossil vertebrates commonly occur as float associated with outwashed pebble conglomerates. However, vertebrate fossils of small animals most commonly have been found in sandy siltstones below conglomerates and generally without any associated large vertebrates. Fossil specimens occur as isolated elements, mixed, and many showing evidence of transport and very few being found in place. There are no articulated specimens and only a few instances where elements appear to be associated. There seems to be no stratigraphic variation in the number and faunal type of specimens. Fossil wood occurs as transported logs, mostly deposited in sandy, channel-fill facies.

Age and paleoecology

Over the past 70 years, fossils representing a rich, early to middle Pleistocene (Irvingtonian Land Mammal Age) terrestrial and freshwater assemblage have been recovered from fluvio-deltaic deposits of the Colorado River in the El Golfo Badlands. Taxa represented include approximately 80 vertebrates, four plants, and one freshwater mollusk. Additional taxa, most notably fish and birds, are undoubtedly present in as yet undiagnosed fossil material. A complete list of flora and fauna (modified from Lindsay, 1984 and Davis et al, 1990) comprising the El Golfo local paleobiota is given in Table 1.

Determining the age of the El Golfo beds is problematic. They cannot be dated radiometrically. Previous and ongoing work on magnetostratigraphic chronology, while showing a reversed period (Matuyama Reversed Chron) that verifies a late Pliocene to early to middle Pleistocene time, may not be able to further delineate the age of the deposits. At present, their age is based upon biochronologic correlation of the mammalian part of the assemblage. An Irvingtonian Age is implied by the joint association of *Mammuthus meridionalis, Megalonyx wheatleyi, Nothrotheriops texanus,* and *Sigmodon curtisi* (Shaw, 1981). In addition, the fossil mammals correlate well with those from the Irvingtonian Age portion of the Anza-Borrego Desert State Park stratigraphic section (Downs and White, 1968; Cassiliano, 1999; Jefferson and Lindsay, 2006), and the Curtis Ranch assemblage in Arizona (Lindsay and Tessman, 1974; Lindsay, 1978). [The Curtis Ranch fauna is now considered by some to be late Blancan (Bell et al., 2004).]

In a sample of over 4300 individual identified fossil specimens from El Golfo, the relative numbers of vertebrate taxa shows a preponderance of herbivores (81%), of which 23% are horse, 19% deer, and 14% camel. Rodents

Table I
El Golfo local paleobiota
Irvingtonian Land Mammal Age (Early to Middle Pleistocene)

Explanation: † = extinct taxon; cf. = compares favorably with; aff. = close affinity to; ? = tentative assignment. Most taxonomic names conform to current usage. Higher taxonomic categories are included where known, and common names appear in parentheses where applicable.

Division Magnoliophyta
Class Liliopsida
Order Arecales
Family Palmeae
Washingtonia sp. (fan palm)

Class Magnoliopsida
Order Ranales
Family Lauraceae
Umbellularia salicifolia † (bay laurel)
Order Juglandales
Family Juglandaceae
Juglans pseudomorpha † (walnut)
Order Salicales
Family Salicaceae
Populus sp. (cottonwood)

Division Animalia
Class Bivalvia
Order Unionoida
Family Unionidae
Anodonta sp. (freshwater mussel)

Class Osteichthyes
Order Cypriniformes
Family Catostomidae
Xyrauchen sp. (razorback sucker)
Family Cyprinidae
cf. *Ptychocheilus* sp. (pike-minnow)
Order Siluriformes
Family Ictaluridae (catfish)
Order Perciformes
Family Istiophoridae (billfish)
Other Unidentified Teleost fish

Class Amphibia
Order Anura
Family Bufonidae
Bufo alvarius (Colorado River toad)
Family Ranidae
Rana sp. (frog)

Class Reptilia
Order Chelonia
Family Kinosternidae
cf. *Kinosternon* sp. (mud turtle)
Family Emydidae
Trachemys scripta (pond slider turtle)
Family Testudinidae
Hesperotestudo sp. † (giant land tortoise)
Gopherus sp. ? (tortoise)
Order Squamata
Family Iguanidae (iguanid lizard)
Family Boidae
Constrictor constrictor (boa constrictor)
Family Colubridae
Masticophis sp. (whip snake)
Rhinocheilus lecontei (longnose snake)
Family Crotalidae
Crotalus sp. (rattlesnake)

Class Aves
Order Ciconiiformes
Family Phoenicopteridae
Phoenicopterus sp. (flamingo)
Order Anseriformes
Family Anatidae
cf. *Anas* sp. (duck/teal)
Subfamily Anserinae
Chen sp. ? (goose)
Order Accipitriformes
Family Vulturidae
Coragyps occidentalis † (western black vulture)
Gymnogyps sp. (condor)
Family Accipitridae
Buteo sp. (hawk)
Order Galliformes
Family Tetraonidae ? (grouse)
Order Gruiformes
Family Gruidae
Grus sp. (crane)
Order Strigiformes
Family Strigidae
Bubo sp. (great horned owl)
Other unidentified birds

Class Mammalia
Order Insectivora
Family Soricidae
Sorex sp. (shrew)
Order Xenarthra
Family Megalonychidae
Megalonyx wheatleyi † (Wheatley's ground sloth)
Family Megatheriidae
Nothrotheriops texanus † (small ground sloth)
Family Mylodontidae
Paramylodon harlani † (Harlan's ground sloth)
Family Myrmecophagidae
Myrmecophaga tridactyla (giant anteater)
Order Carnivora
Family Canidae
Canis lepophagus † (wolf)
Canis edwardii † (wolf-coyote)
cf. *Urocyon* sp. (fox)
Family Ursidae
Tremarctos cf. *T. floridanus* † (Florida cave bear)
Family Procyonidae
Procyon sp. (raccoon)
Family Mustelidae
Taxidea sp. (badger)
genus of unidentified skunk
Family Felidae
cf. *Lynx* sp. (bobcat)
Felis rexroadensis † (Rexroad cat)
Panthera sp. cf. *P. onca* (jaguar)
Homotherium sp. † (dirk-toothed cat)
Miracinonyx inexpectatus † (American cheetah)
Family Hyaenidae
Chasmaporthetes ossifragus † (American hyena)

Order Lagomorpha
  Family Leporidae
    Sylvilagus hibbardi † (Hibbard's rabbit)
    Lepus sp. (jack rabbit)
Order Rodentia
Family Sciuridae
Spermophilus sp. (squirrel)
  Family Geomyidae
    Geomys sp. (gopher)
  Family Heteromyidae
    Dipodomys sp. (kangaroo rat)
  Family Castoridae
    Castor sp. cf. *C. californicus* (Kellogg's beaver)
  Family Cricetidae
    Sigmodon curtisi † (Curtis cotton rat)
    Sigmodon sp. (cotton rat)
  Neotoma (Neotoma) sp. (woodrat)
    Neotoma (*Hodomys*) sp. (woodrat)
    Ondatra sp. ? (muskrat)
    Unidentified sp. of mouse
  Family Erethizontidae
    Coendou sp. (porcupine)
  Family Hydrochoeridae
    Neochoerus dichroplax † (capybara)
Order Proboscidea
  Family Gomphotheriidae
    Cuvieronius sp. † (Cuvier's gomphothere)
  Family Elephantidae
    Mammuthus sp. cf. *M. meridionalis* † (southern mammoth)
Order Perissodactyla
  Family Equidae
    Equus sp. (large) † (giant horse)
    Equus (*Plesippus*) sp. † (zebra-like horse)
    Equus sp. aff. *E. scotti* † (Scott's horse)
    Equus sp. cf. *E. conversidens* † (Mexican horse)
  Family Tapiridae
    Tapirus sp. cf. *T. haysii* † (Hay's tapir)
    Tapirus sp. cf. *T. californicus* † (California tapir)
Order Artiodactyla
  Family Tayassuidae
    Platygonus sp. cf. *P. vetus* † (Leidy's peccary)
  Family Camelidae
    Gigantocamelus sp. or *Titanotylopus* sp. † (giant camel)
    Camelops sp. † (western camel)
    Hemiauchenia sp. cf. *H. blancoensis* † (Blanco llama)
    Hemiauchenia sp. (small) † (long legged small llama)
    Hemiauchenia gracilis † (gracile llama)
    Palaeolama sp. † (stout-legged camel)
  Family Cervidae
    Odocoileus virginianus (white-tailed deer)
    Odocoileus hemionus ? (mule deer)
    cf. *Navahoceros* sp. † (mountain deer)
  Family Antilocapridae
    Capromeryx furcifer † (Matthew's pronghorn)
    Tetrameryx (*Stockoceras*) sp. † (4-horned pronghorn)
    gen. et sp. nov. † (4-horned pronghorn)
Family Bovidae
    Bovidae ? gen. et sp. indet. (unknown bovid)
    Ovis sp. nov. (mountain sheep)

comprise 11% of the fossils, rabbits and hares 6%, pachyderms 3%, and ground sloths 2%. Carnivores (mostly cats, followed by canids and others) comprise 4% of the total. Turtles and tortoises are 8%. Fish and birds have a smaller representation but remain to be fully studied.

The assemblage is assumed to represent a single paleobiota. Based on the ecological requirements of the closest living taxa to those represented by the fossils, the existence of at least four ecological communities is suggested. They are freshwater aquatic, riparian galleria forest, shrub and brush woodland, and savannah-like grassland.

The Colorado River provided a stable supply of water that sustained a permanent aquatic community. This is shown by the presence of a freshwater mussel and fishes, amphibians, turtles, aquatic birds, tapirs, capybara, beaver, and the tentatively identified muskrat indicating running streams with local ponding. Riparian woodlands would have included fan palm and cottonwood.

Given the type and distribution of habitats on the historic Colorado River delta, localized patches of scrub and woodlands probably occupied edaphic areas at slightly higher elevations proximal to depositional sites. They were perhaps separated by expanses of grassland on the wider and highest interfluves. Petrified wood is abundant throughout the region and indicates mixed woodlands; identified hard woods include laurel, walnut, and cottonwood. This woodland community is represented most commonly by deer; other animals include two species of ground sloth, a porcupine, possibly some of the smaller camels, and a browsing gomphothere.

The existence of savannah-like grasslands is shown by a wide variety of grazing and grazing-browsing herbivores. These include several species of horses (the only fully grazing herbivores), larger camelids, and antilocaprids, along with giant tortoise, a mammoth, and a third species of ground sloth. The relative tooth crown height is in general a poor judge of dietary preference, and some of these taxa probably preferred woodland/grassland ecotonal boundary conditions. Habitat and dietary preferences will be better defined for many of the mammalian species as future studies using stable isotopes progress. A further representative of the savannah-like grassland community is the giant anteater, whose El Golfo record is the northern-most known example of this animal (Shaw and McDonald, 1987).

The presence of fan palm, giant tortoise, boa constrictor, two species of tapir, capybara, and giant anteater in the fauna suggests that tropical to subtropical climates may have predominated at that time (Brattstrom, 1961; Shaw and McDonald, 1987).

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A preliminary report on fossil bony fish remains recovered from early Pleistocene Colorado River delta deposits exposed in northwestern Sonora, Mexico

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For more than a decade, staff and volunteers from Reserva de la Biosfera–Alto Golfo de California y Delta del Rio Colorado (SMERNAP), Arizona Western College, and George C. Page Museum have been surveying exposures of Irvingtonian age Colorado River delta sediments for paleontological resources in northwestern Sonora, Mexico. More than 3500 mapped vertebrate localities have yielded over 66 genera. Many fossil wood sites have also been located. In addition to the larger vertebrate fossils, several new microvertebrate localities have been identified. In conjunction with this work, a small number of bony fish remains have been recovered.

The bony fish remains consist primarily of isolated bones and vertebrae. At least three species are represented. Two of the taxons, the razorback sucker (*Xyrauchen texanus*) and pike minnow (*Ptychocheilus lucius*), are found in the present-day Colorado River, but are rare or locally extinct in the lower portions of the basin. Over the last one hundred and twenty years, construction of reclamation dams and the introduction of exotic game fishes have caused dramatic declines in the populations of native fishes in the Colorado River and its major tributary rivers.

A third species appears to be from the family Sciaenidae (croakers), a marine fish family that up until recently was very abundant in shallow marine waters of the northern Gulf of California. The fossil partial sciaenid vertebra probably represents a four to five foot individual. Today, at least eleven kinds of croakers inhabit the northern Gulf, but only two species—the white seabass (*Atractoscion nobilis*) and totoaba (*Totoaba macdonaldi*)—reach lengths of four or more feet. Until more comparative material becomes available, identification of this vertebra can only be taken to familial level.

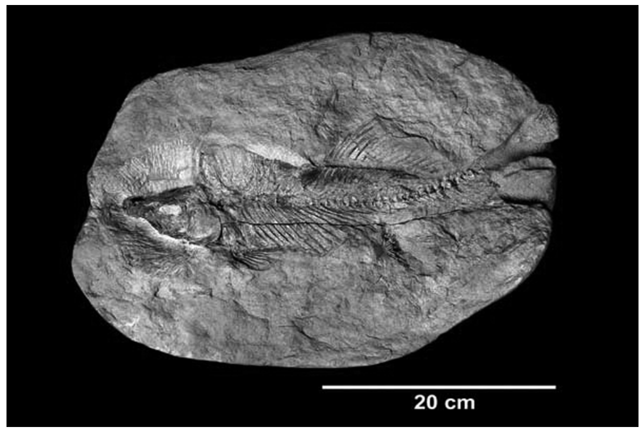

*Fig.1. The razor-back sucker (*Xyrachen texanus*) skeleton from the Arroyo Diable Formation of the Palm Spring group (Stewart and Roeder, 1993) is indistinguishable from the modern* X. texanus *found historically in the Colorado River. Photograph by B. Marrs, from Gensler et al. in Jefferson and Lindsay (2006).*

Playas, dune fields, and global aeolian processes—the paradox of the "dust-free" Mojave Desert

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Abstract

On a global scale, deserts are the dominant source of atmospheric soil dust. As in other desert regions, aeolian landforms are common in California deserts, however, undisturbed areas of the Mojave Desert have unusually low atmospheric dust loading compared to many other desert and semi-desert regions. Moreover, the dominant source of fine soil dust in the air over the Mojave Desert appears to be Asia. Contributing factors for this seeming paradox possibly include the local climate, the structural geology of the Mojave region, its vegetation and land use history, and its position relative to global atmospheric circulation patterns.

Introduction

Earth's great deserts, such as the Sahara, are known for their dust storms. In these deserts, dust is generated almost daily, and it is often visible in satellite images (Figures 1 and 2). Similarly, some semi-desert regions, such as west Texas are also known for their dust storms (Figures 3 and 4).

In California deserts, such events are relatively rare. For example, the US EPA has declared the Owens Lake playa the largest single dust source in the U.S. (USEPA, 1999) (Figure 5), but its dust storms only escape the immediate area a few days per year, and significant dust deposition is confined to areas within a few tens of kilometers from the lake (Reheis, 1997). In undisturbed desert areas, such as Death Valley National Park, aerosol mass loadings are

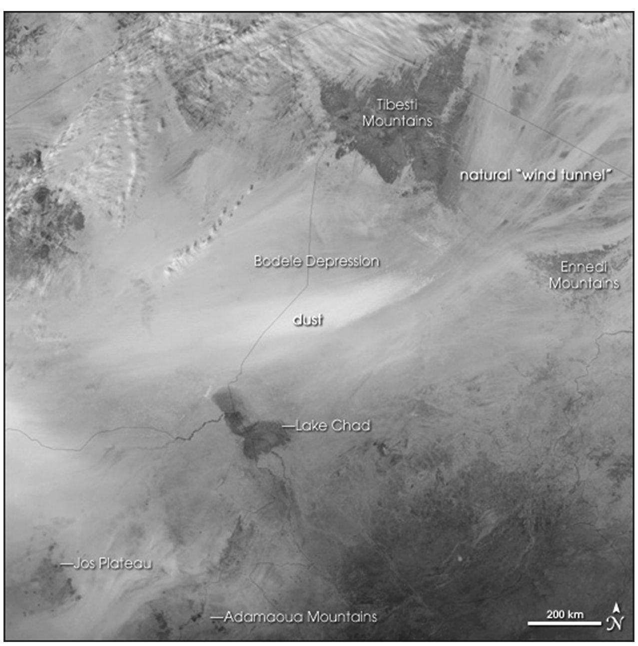

Figure 1. Dust generation in the Bodele Depression, Chad (NASA; Washington et al.*, 2006).*

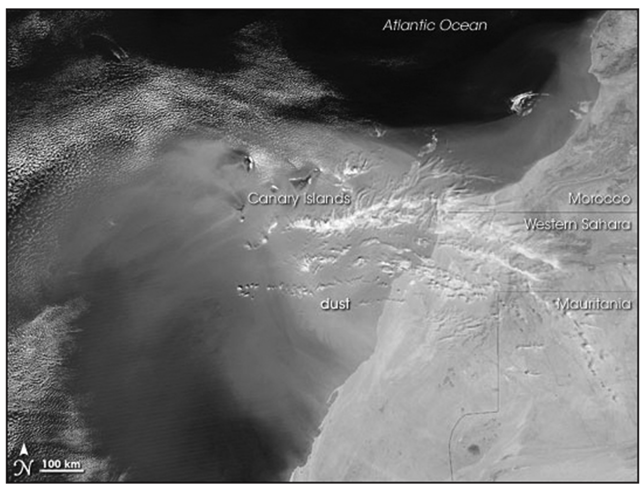

Figure 2. Dust plumes from Africa impacting the Canary Islands (NASA).

Figure 3. "Haboob" dust front near Lubbock, Texas, 2006.

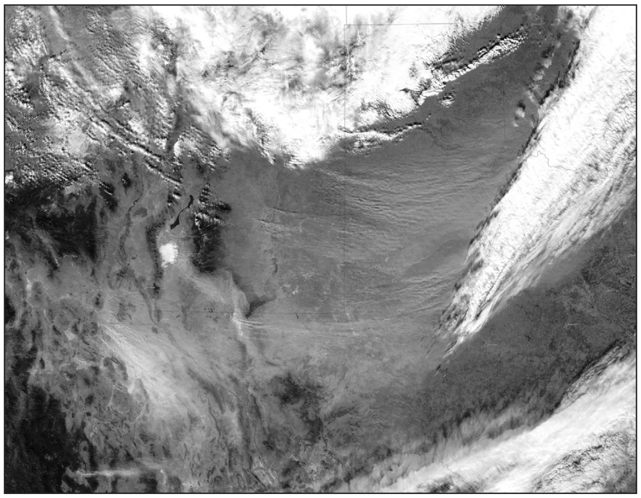

Figure 4. Dust plumes from alluvial plains and playas in Rio Grande and Pecos valleys (NASA).

Figure 5. Owens Lake basin and Sierra Nevada crest with (right) and without (left) dust (GBUAPCD).

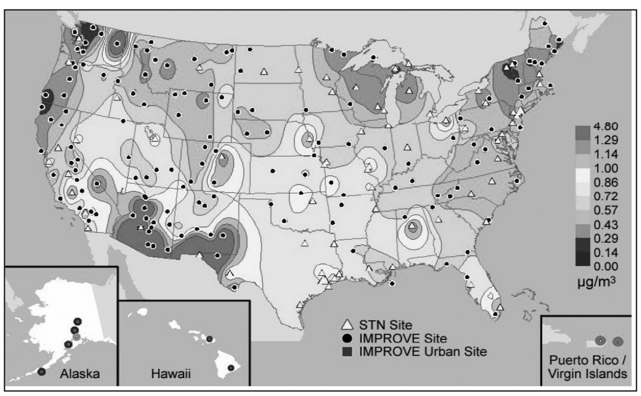

Figure 6. Annual average fine (<2.5µm) soil particle concentrations in the U.S. (DeBell, 2006).

generally low (Figure 6; DeBell, 2006).

The factors that control wind-blown dust generation in an instantaneous, micro-scale context (e.g. a wind tunnel) are wind shear at the surface, soil grain size, soil moisture, soil organic content, plant cover, and surface roughness. Total dust emission can be computed as the integral of the local dust production over time and fetch. Similarly, dust deposition is a function of dust concentration, grain size, wind speed, turbulence, surface roughness, and surface "stickiness," and it, too, can be similarly computed.

In theory, then, it is possible to fairly fully determine the sources and rates of dust generation and deposition for any point or region, given sufficient spatially and temporally resolved data are available (Shao, 2001; Gillette, 1999).

Unfortunately, all of these factors exhibit very high spatial variability (especially instantaneous wind shear and surface roughness), so it is not yet possible to model regional or global wind-blown dust systems from "first principles." Rather, the analysis of regional dust systems must rely on converging inferences from dust and source-soil chemical and physical analysis and somewhat crude dust generation and transport models.

In addition, the study of dust is fraught with opportunities for observer bias, ranging from misapplying what we know from our long subjective experience with dust, to the universal scientific risks of unjustified assumptions and under-conceived experiments.

The following discussion review some recent research on global and regional dust dynamics in the context of the deserts of eastern California and southern Nevada, highlights inconsistencies across disciplines, and concludes with some summary observations and new research questions.

Deserts as Dust Sources

The association of deserts and dust is obvious, and leads to some simple hypotheses as to the causes of dust. Most significant is the lack of plant cover, followed by the general lack of soil moisture. A second, but less obvious factor is the hydrology of desert drainage systems, which leaves fresh alluvium exposed to the air when water is at a minimum. Third is the effect of human activity in deserts – grazing, mining, or other disturbances are more likely to produce long-term disruption of vegetation and loosening of soil materials than in rainy climates.

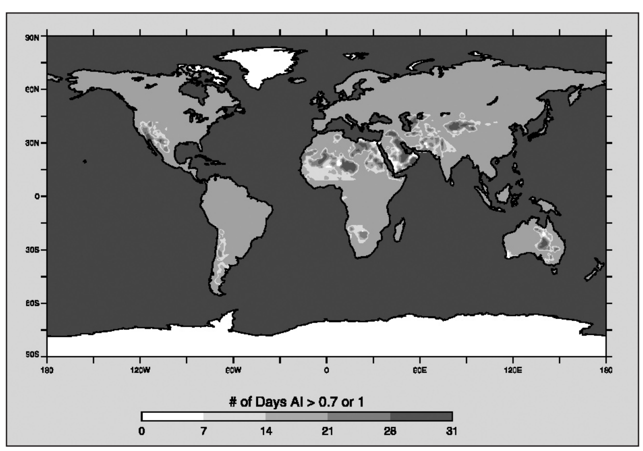

Figure 7. Global dust generation represented by number of days per month with TOMS Aerosol Index greater than 0.7 (Prospero, et al., 2002).

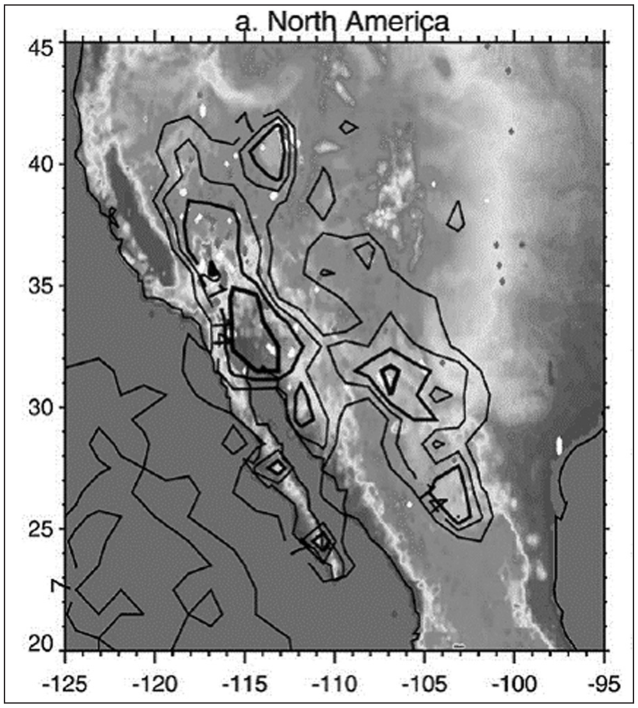

Figure 8. Contours of frequency of Aerosol Index > .7 in southwestern North America, contour interval 7 days/mo. Note "hotspots" of Great Salt Lake, Salton Trough, and the playas of the lower Rio Grande at Samalayuca, near Ciudad Juarez, Mexico and farther south in Coahuila (Prospero, et al., 2002).

Nonetheless, dust is not confined to deserts, and not all deserts produce large amounts of dust. None of these factors are unique to deserts, and their mixes in different desert areas are highly variable.

California deserts in the context of global dust

Observations

With the advent of satellites, it has become possible to map the global distribution of dust. Prospero *et al.* (2002) compiled a picture of global dust from satellite data (Figure 6). Focusing more closely on the Southwestern U.S. (Figure 8), which shows "hotspots" around the Great Salt Lake, the Salton Trough, and the lower Rio Grande playas and dunefields near Ciudad Juarez, Mexico and farther south in Coahuila (Prospero *et al.*, 2002). Note that this map is not congruent with the historical aerosol data in Figure 6.

Washington *et al.* (2006) also developed maps of global dust from the TOMS AI data, but found only weak AI enhancement in the Great Basin (Table 1) compared to other regions, and noted the lack of congruence with historical dust patterns (Figure 9).

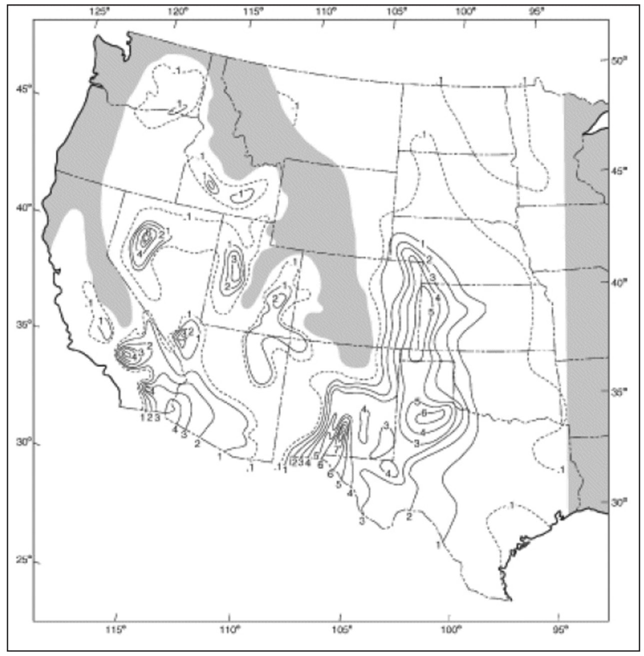

Figure 9. Frequency of surface dust observations in the United States. Shaded areas represent no observations of dust storms. (Washington et al., 2005, after Orgill and Sehmel (1976)).

Table 1. Maximum Mean Aerosol Index (AI) Values for Major Global Dust Sources Determined from TOMS (Washington et al., 2005).

Location	Mean AI Value
Bode 'le' Depression of south central Sahara	>3.0
West Sahara in Mali and Mauritania	>2.0
Arabia (southern Oman/Saudi border)	>2.1
Eastern Sahara (Libya)	>1.5
Southwest Asia (Makran coast)	>1.2
Taklaamakkan/Tarim Basin	>1.1
Etosha Pan (Namibia)	>1.1
Lake Eyre Basin (Australia)	>1.1
Mkgadikgadi Basin (Botswana)	> 0.8

Models

Models of regional to global dust generation employ surrogates, such as land use/land cover maps, satellite observations of plant cover, maps of soil texture, and areally-averaged wind speed. Some also include topography to identify basins where surficial sediments can accumulate (Engelstaedter *et al.*, 2003). These variables, processed through non-physical "scaling factors," yield dust emission estimates.

Well chosen generalizations seem to work pretty well at the global scale (spatial resolution of many tens to hundreds of kilometers) and capture major dust sources such as the Sahara pretty well, but the efficiency of their

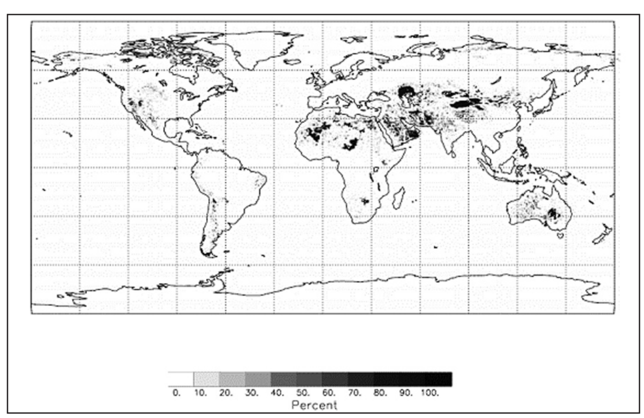

Figure 10. Tegen et al., 2002) percent of 0.5x0.5 degree grind cells with dust potential.

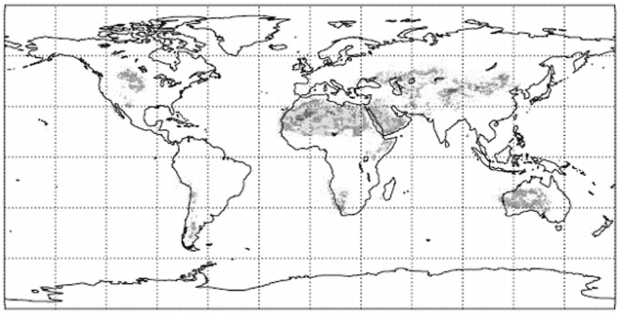

Figure 11. Modeled dust emissions (Tegen et al., 2002).

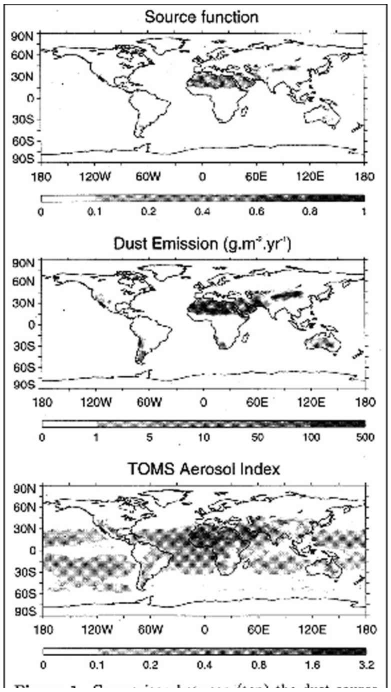

Figure 12 Dust sources and emissions as modeled by Ginoux et al. (2001).

algorithms can be tested by examining how well they do for smaller dust source areas, where local knowledge can be used to test the spatial specificity of the results.

Recent efforts at modeling global dust production locate dust sources in the U.S. somewhat inconsistently. For example, Tegen *et al.* (2001) (Figure 10) identify similar sources in the Southwest to those identified by Prospero *et al.* (2002) from the satellite data, but their dust emission estimate focuses North American dust emissions in the high plains (Figure 11). The Great Plains dust source is not reported by Prospero *et al.* (2002).

The Tegen *et al.* (2001) results contrast sharply with the formulation of Ginoux *et al.* (2001) (Figure 12), which confines most North American dust to the Salton Trough, and misses the dust sources along the U.S.–Mexico border as reported by Prospero *et al.* (2002) and evident in the IMPROVE (Figure 6) and weather observer (Figure 9) data.

Surficial dust chemistry

A somewhat different picture of dust in this region comes from chemical analysis—tracing atmospheric dust to its source(s) by its chemical similarity to source region surface materials.

The driving assumption in interpreting chemical analyses is that the list of potential sources is known, usually assumed to be unconsolidated alluvium or disturbed soil areas upwind of the sampling point, and thus the analysis is inherently spatially constrained to a pre-identified study area. A good example of this approach is the work of Reheis (1997) and Reheis and Kihl (1995) (Figures 13, 14 and 15). Their *a priori* conceptual model is that salt, carbonate, gypsum, and clays would be strongly associated with playas and, possibly, calcareous bedrock, and fractions of these components would decrease with distance downwind from such sources. Conversely, coarser dust and lower soluble content was expected downwind of alluvial flats and granitic or volcanic bedrock. Their data, however, show only salt (NaCl) to be reasonably consistent with this conceptual model, and the data only track the conceptual model in areas very close to strong dust sources such as Owens Lake (Figure 13) and disturbed areas such as farmlands and construction sites.

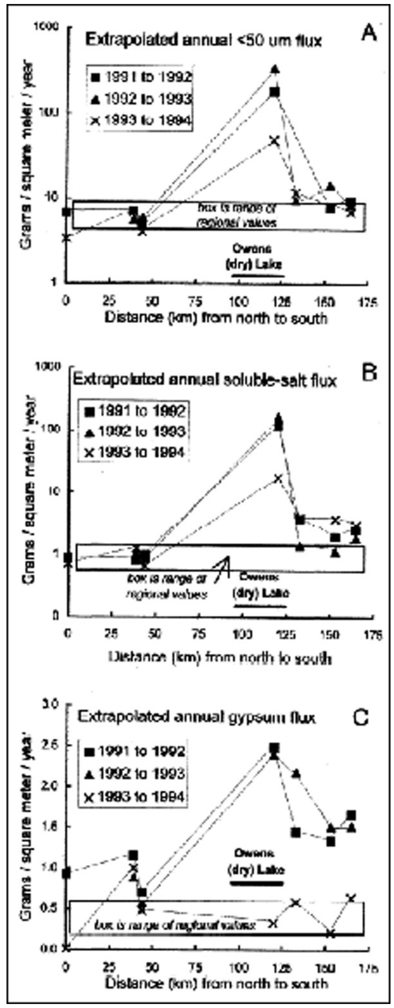

Figure 13. Dust, salt, and gypsum deposition along a transect from Bishop to Little Lake in the Owens Valley (Reheis, 1997).

Reheis and Kihl report several salient findings:

1 - In general, undisturbed surfaces are weak dust sources;

2 - the basin and range topography of the Mojave region runs cross-wise to prevailing winds, thus there are limited opportunities for dust generation over long fetches of topographic lows rich in loose sediment;

3 - there is not a strong link between playas and regional dust loading for undisturbed playas, playas tend to

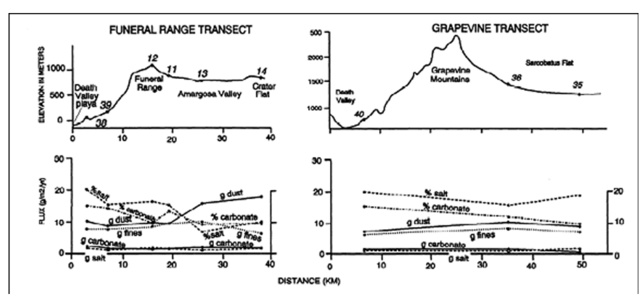

Figure 14. Chemical composition of dust samples in transects across mountain ranges east of Death Valley (Reheis & Kihl, 1995).

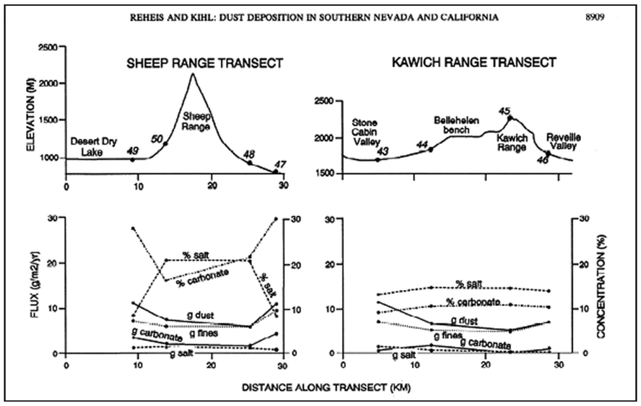

Figure 15. Chemical composition of dust samples in transects across two mountain ranges north of Las Vegas (Reheis & Kihl, 1995).

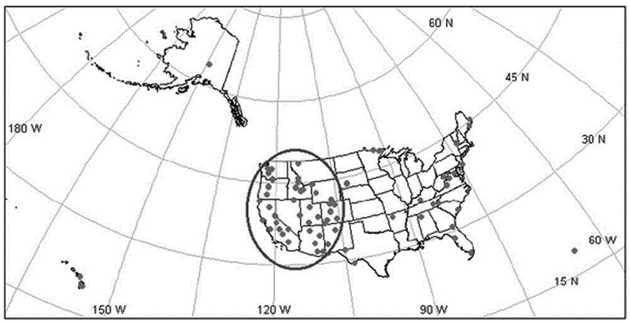

Figure 16. IMPROVE network in the 1990s.

stabilize unless mechanically or hydrologically disturbed (e.g. Owens Lake);

4 - alluvial surfaces are relatively weak dust sources unless disturbed by agriculture, vehicles, or human development;

5 - stream channels are the dominant source of loose alluvium available for aeolian transport,

6 - dust in the Salton Trough is more abundant and somewhat different chemically from that in the rest of the region,

7 - carbonate is unexpectedly high in many areas, and generally, associated with gypsum,

8 - gypsum is associated with areas of least summer rain, and very broadly distributed.

The work of Reheis and others suggests three components of the answer to the question why the dust load in the Mojave is less than in neighboring deserts—less rainfall means less sediment movement, less disturbance (farming and grazing) reduces upland surface dust generation, and topography limits dust storm dynamics.

However, in regard to the chemistry, Reheis and Kihl cannot explain the gypsum and carbonate very well, but they offer one suggested explanation: some gypsum may come from atmospheric conversion of sulfuric acid (originating as SO_2 from fossil fuel combustion) to gypsum on the surface of carbonate particles.

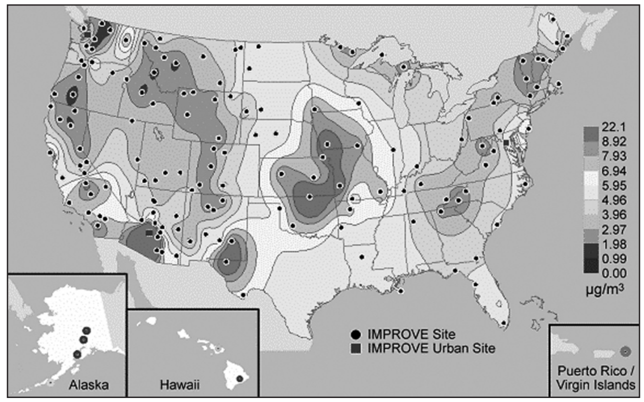

Figure 17. Average coarse (>2.5 and <10µm diameter) particle concentrations from IMPROVE data.

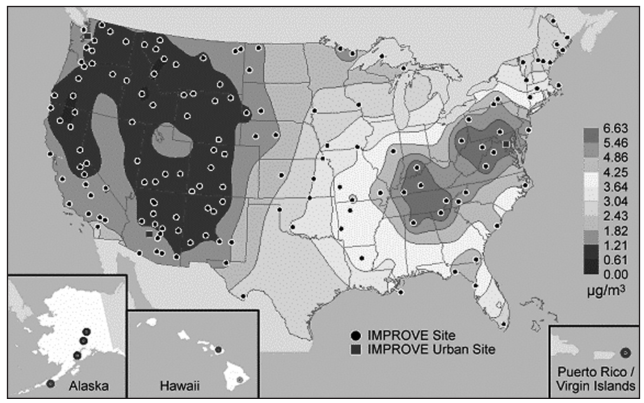

Figure 18. Annual average sulfate concentrations in the IMPROVE network.

Spatial patterns of airborne dust chemistry

The problem of gypsum leads naturally to two questions: what's the source of the carbonate to drive such a transformation, if this is the correct interpretation, and where does the sulfur come from?

Since the IMPROVE network (Figure 16) is spread across the entire U.S. and collects measurements twice-weekly, it is possible to use IMPROVE data not only to evaluate dust for the few monitoring sites in the American deserts, but to examine regional patterns to see if there is a distinct regional dust signature.

Returning to Figure 6, it is clear that there is a regional fine dust maximum in the low elevation and Mexican border areas of Arizona and New Mexico. Inference that this is local dust is supported by the similar spatial distribution of coarse particles (from 2.5 to 10 µm diameter) (Figure 17), which have high settling velocities and thus weak long range transport potential. The coarse particle data show a pattern reminiscent of that reported from weather observation records (Figure 9) and the global dust model of Tegen *et al.* (2001) (Figure 11). Both the coarse and fine maps show a weaker dust signature going west into the Mojave region, consistent with the process generalizations from Reheis and Kihls. Also note that the IMPROVE data does not show the Great Salt Lake source, although this may be a consequence of only having elevated monitoring sites

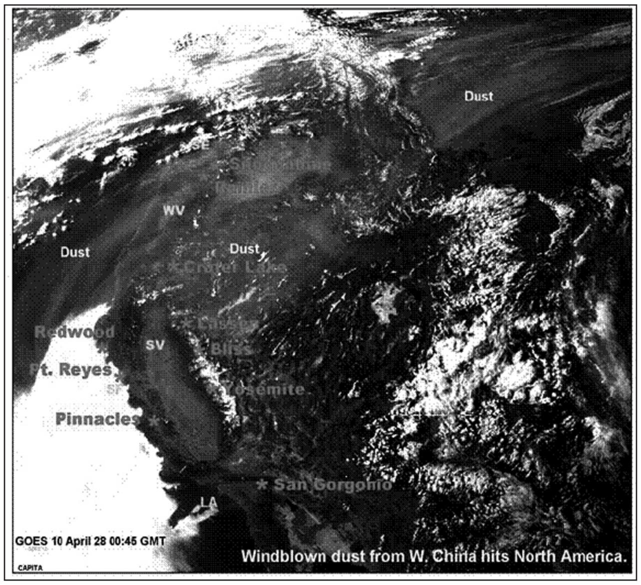

Figure 19. An Asian dust cloud arriving over North America, April 28, 1998.

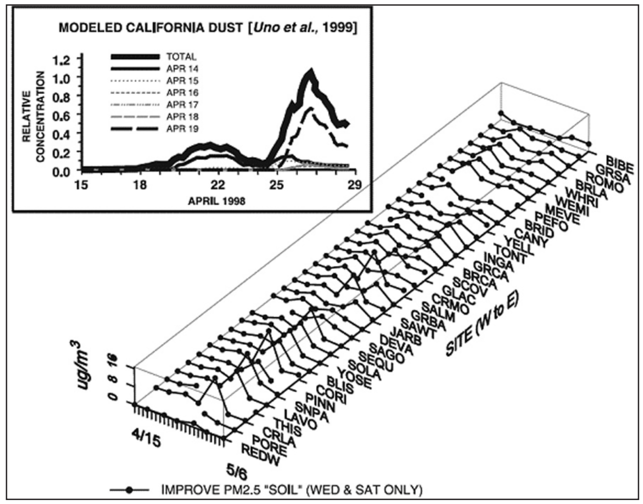

Figure 20. Asian dust pulses moving across the IMPROVE network, April 15-May 6, 1998 (VanCuren and Cahill, 2002).

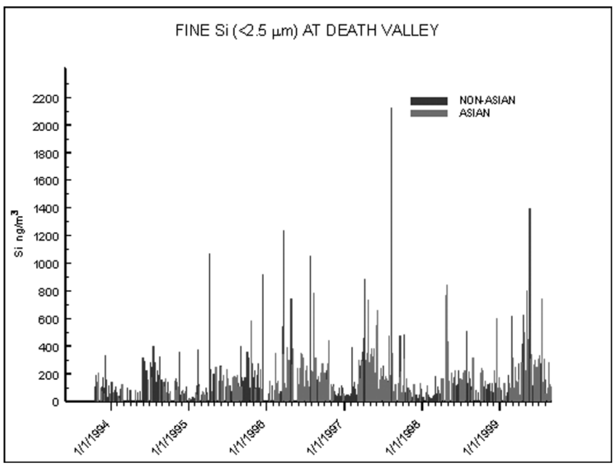

Figure 21. Six years of twice weekly fine aerosol Si concentrations at Death Valley.

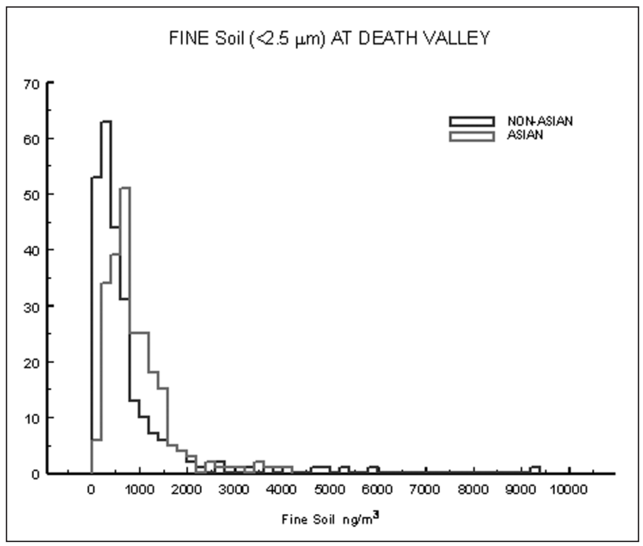

Figure 22. Frequency distributions of Asian and Non-Asian soil chemistry in Death Valley IMPROVE data.

(~2km) near it.

Sulfate (reported as gypsum by Reheis and Kihl, 1995; sulfate ion reported as ammonium sulfate by IMPROVE) is more problematic – IMPROVE sulfate data (Figure 18) do not suggest any significant regional source.

Asian soil

Although it seems counter-intuitive, Asian dust is a regular visitor to North America based on analysis of IMPROVE data from monitoring sites in the Sierra–Cascade and Rocky Mountains regions (VanCuren and Cahill, 2002). Excess soil was observed in Death Valley during the great Asian dust storm of 1998 (Figures 19 and 20, site name DEVA).

Applying the analytical technique from VanCuren and Cahill (2002) to test for the presence of Asian soil in the IMPROVE record for Death Valley results in a division of samples into two dust populations: days with overwhelmingly Asian chemistry, and all other days. The temporal

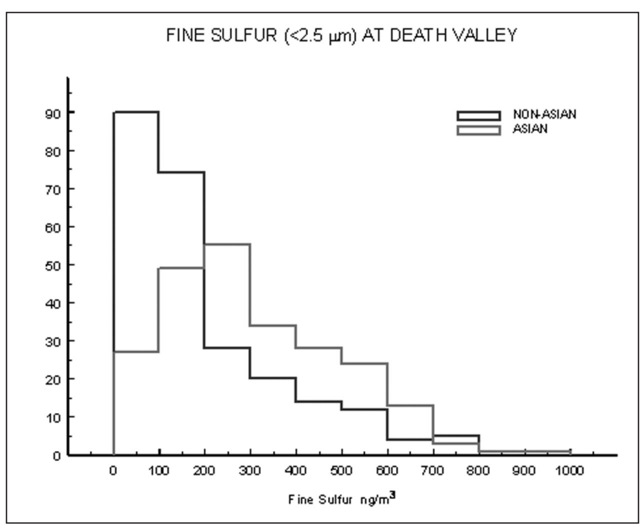

Figure 23. Fine sulfur concentration distributions at Death Valley in the 1990s.

pattern for this division is shown in Figure 21, and the concentration frequency distributions are shown in Figure 22.

The data in Figures 21 and 22 show that there is a significant Asian dust impact at Death Valley. Although the highest dust concentrations are due to local sources, the day-to-day dust load during the 1990s was about evenly split between the local and Asian sources (233 Asian days, 247 non-Asian ones; mean Asian fine soil concentration 0.9 $\mu g/m^3$, mean non-Asian 0.7 $\mu g/m^3$).

The chemistry of the Asian plume may also explain the origin of some of the sulfate observed in Reheis and Kihl's dust samples – analysis of the combustion-related materials traveling with the Asian dust shows that sulfate, especially calcium sulfate formed from acid reaction with carbonates, is a common constituent, with a mean concentration of about 0.2 $\mu g/m^3$, similar to the modal concentration in the Death Valley data (Figure 23) (Cao *et al.*, 2005; Song *et al.*, 2001).

Conclusions

Reviewing the multidisciplinary work on dust in the Mojave region leads to some general conclusions:

1 - Although we are all familiar with dust, this does not manifest itself in common definitions or compatible methods across geology, soil science, and atmospheric research. Even within these groups there is no consensus definition and limited agreement on methods.

2 - Normative expectations of large natural dust yields from deserts are potentially misleading. Models based on geomorphic and/or biotic factors alone are inadequate.

3 - Size matters I. The very long fetch (many hundreds of km) available in the Sahara and Takla Makan deserts is not replicated in North America.

4 - Size matters II: Coarse dust is limited in transport range (sand size material is often trapped in enclosed basins), but fine material (less than ~ 4µm) can reach high altitudes and persist in the atmosphere for many days.

5 - Sediment supply is a paramount factor. Over time, many surfaces can become deflated or crusted. Fresh sediment or surface disturbance are essential to creating regional dust loads.

Finally, since African dust has been shown to be a significant fraction of fine dust at Big Bend National Park in Texas; Chinese dust has been found to form varve-like deposits in Japanese alpine lakes; Asian dust has been retrieved from glaciers in Alaska and British Columbia, and atmospheric measurements suggest persistent Asian dust flux to much of North America, and because regional and global dust flux varies with climate, it would be interesting to conduct a few experiments:

1 - Sample aeolian sediments in remote, preferably alpine traps in the Sierra Nevada, Cascade, or Rocky Mountains, especially glacial lakes and deposits overlying glacial till, to compare to atmospheric measurements.

2 - Perform stable isotope analysis on airborne or surficial dust, especially Sr and Nd, which have successfully been used to trace Asian dust to Greenland.

3 - Perform O-isotope analysis on sulfates to compare oxidative pathways in source or depositional environments to expectations from upwind samples or local materials.

4 - Use X-ray crystallography on clays in aeolian deposits and compare to samples collected by high-volume air sampling.

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A cultural–geological analysis of quartz shatters from the Glamis Dunes region, Imperial County, California

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Abstract

From 2004-2007, LSA Associates, Inc. (LSA) carried out data recovery for ten prehistoric cultural resources and a historic 19th to 20th century wagon road with associated artifact scatters in the Glamis Dunes Region of Imperial County, California. LSA relocated and more precisely mapped previously recorded trail segments and associated features. Locations that had previously been extrapolated from UTMs were remapped with a Trimble GIS. Previously recorded jasper chipping stations and prehistoric pot-drops were found intact and mapped, photographed, and collected.

The field crew found unusual zones with large quantities of shattered quartz boulders protruding from the surface. Three testable hypotheses were developed: (1) they were prehistoric quarry sources that had been split to evaluate the quality of the stone and to provide sharp edge fragments for expedient use; (2) gold-mining prospectors had randomly but systematically broken quartz nodules as they examined the landscape, looking for inclusions of gold ore that might indicate the presence of nearby veins, or (3) they were natural in origin. At this point we would like to add a fourth hypothesis: that the shatters are natural in origin, but acquired cultural significance and are part of a "cultural landscape." Following the line of thought that "if the quartz shatters gave us pause, then maybe someone else will eventually share our uncertainty", further study of the quartz shatters at the project is the subject of this paper. Field research indicated that Pleistocene alluvium covering most of the project consisted of medium to large cobbles of quartz, schist, marble and limestone from metamorphic sources. Miocene volcanics were also present. After deposition, alluvial fans were subject to development of soil profiles which worked their way downward as erosion developed terraces. During this deflation, all cobbles passed through a horizon of expansive clays in the soil profile. Metamorphic rocks with foliation, including quartz, were fractured by forces in these expansive clays. Volcanic rocks also suffered during this process. All cobbles exposed by erosion on the surface were fractured, with fractures showing patination by metallic oxides (pyrolusite and goethite). Recent fractures by Native Americans, prospecxtors and vehicles can be distinguished by fresh, non-patinated surfaces.

Project location

The project location is located in Imperial County, California (Figure 1). The cities of Brawley and Palo Verde are approximately 35 miles to the west and northeast. The Glamis Beach store is three miles southwest.

Previous archaeological research

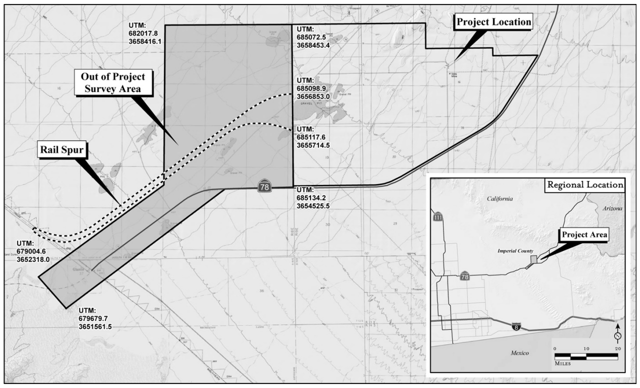
The presence of the naval gunnery range in the Chocolate Mountains, the modern development of the gold mining operations, and the related relocation of State Route 78 led to previous archaeological studies in the MRL area (von Werlof and von Werlof 1977; Bull 1981; von Werlof 1984; Shackley 1988a, 1988b; Schaefer and Pallette 1993). The research questions that were addressed by these reports include hunter-gatherer settlement patterns, lithic procurement strategies, historic use of the area, and World War II activities in a sector of General Patton's Desert Training Center (Bischoff 2000). Of these earlier studies, only von Werlof (1984) mentioned the presence of quartz shatters.

Analysis

As one focus of LSA's research, it was imperative to determine whether or not the quartz shatters were cultural or non-cultural, especially given the cultural connotations attached to other quartz occurrences in the archaeological literature for the California-Arizona deserts. For example,

Cleland and Apple have written (2003:36) with regard to cultural resources identified during the North Baja Pipeline project in the Lower Colorado Desert:

Associated with spiritually significant trails, geoglyphs,



and other places are scatters of shattered milky quartz. White is a symbol of power, and the milky quartz was broken to release power and purify an individual as he or she approached a spiritually powerful area (Lory Cachora, personal communication, 1997; Johnson 1985:37; Weldon Johnson, personal communication, 1997).

Altschul and Ezzo (1994:55) describe the process by which quartz may be deposited at ceremonial sites:

> …as an individual approached an intaglio or sacred area, he would prepare himself by breaking apart milky quartz. The whiteness of the quartz was extremely important, signifying a means of communicating with the supernatural. Embodied in the quartz were supernatural forces that would be freed upon shattering. Once the individual felt sufficiently purified by the experience, he could then proceed to the sacred site itself. Quartz shatter is also evident near one of the other anthropomorph intaglios at Blythe, as well as the horse at Pilot Knob.

At a more general level, Saunders (2003:16) has written:

> From the Amazon to the Andes and from Lower Central America through Mesoamerica and the Caribbean to North America, different philosophies, symbolic associations, technological choices, and materials bolstered or reflected Amerindians' desire for the aesthetic of brilliance. Polished wood, iridescent featherwork burnished pottery, greenstones, obsidian, crystals, greenstones, and a variety of metals were all favored.

Evaluation

(H-1) They were prehistoric quarry sources that had been split to evaluate the quality of the stone and to provide sharp edge fragments for expedient use.

(H-2) Early prospectors had randomly but thoroughly bashed quartz nodules as they examined the landscape, looking for inclusions of gold ore that might indicate the presence of nearby veins.

H-1 was rejected because there no quartz flakes were found in chipping stations at a distance from the shattered nodules and there is no evidence of secondary or tertiary chipping at the site of the shatters. There also was no evidence of expedient use of quartz fragments or flakes.

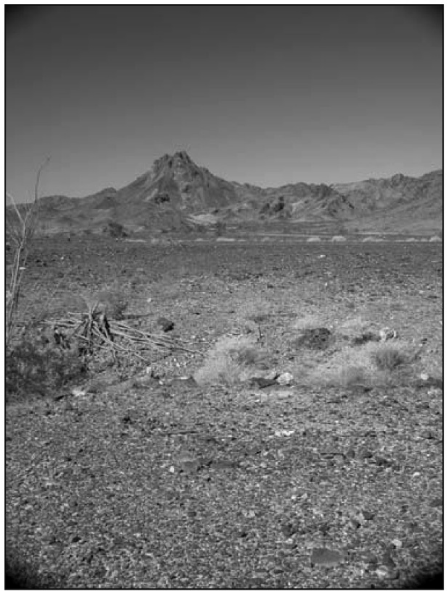
H-2 was rejected because the available surface nodules were entirely shattered. While this thoroughness could theoretically result from repeated visits, practicality suggests otherwise.

(H-3) The third hypothesis posited that the quartz shatters were natural in origin. We consider H-3 confirmed based on an understanding of the desert geology. Specifically, the surface geology of the MRL project area is characterized by late Pleistocene and Holocene alluvial terraces of the Chemehuevi Formation derived from the Chocolate Mountains. This study examined desert pavement development at the northeast margin of the project and along its western margin, with a focus on the upper terraces where the development of soil profiles was greatest and the desert pavement the most mature.

Examination of sediments exposed in gullies noted that the conglomerates consist of mixed lithology clasts or boulders up to three feet in diameter in a groundmass of coarse sand and gravel. The lithologies of the large clasts include metamorphic rocks—marble, foliated micaceous schist, dense hornfels, and milky quartz with schist inclusions—and volcanic rocks including flow-banded olivine basalt. These boulders project above the flat pavement surface and the slopes at the margins of the terraces. Examination of the small (4-inch) clasts and large (36-inch) boulders at terrace margins and projecting above the terrace surface indicated that all were metamorphic or volcanic rock types that had (greater or lesser) pre-existing structural weakness in the form of foliations or bedding planes.

Previous archaeological investigations in and around the project area (Altschul and Ezzo 1994; von Werlof 1984; Cleland and Apple 2003) reported that shattered white quartz boulders had ritual significance in conjunction with the trail networks and associated features. Other observations by geologists (Rob Wailand, BLM geologist, personal communication 2006) suggested that weathering processes involving development of Pleistocene soil profiles fractured the quartz boulders. Lory Cachorro, Quechan cultural expert, also stated (personal communication) that the quartz shatters were natural.

How do these shatters occur naturally? The development of the soil profile underlying the desert pavement involved chemical reduction of ferromagnesian minerals, development of vesicles in silt, and migration of calcium carbonate downward as erosion deflated and lowered the pavement surface on the top of the terrace. As the terrace surface deflated, the large clasts passed through an expanding/contracting zone of silts and clays in the vesicular A-horizon. This expansion fractured the pre-existing weak structural foliations in the cobbles. Further terrace deflation left the large cobbles in elevated positions on the surface of the desert pavement. If the fracturing has not



Photograph 1:

View north northwest toward the southeastern Chocolate Mountains. Terraces in the foreground are capped by imbricated desert pavement that is coated with dark desert varnish.

Photograph 2: *View of fractured block of metamorphic hornfels.*

Photograph 3: *View of fractured block of marble.*

Photograph 4: *View of fractured block of schist.*

been extensive, the broken boulder remains cohesive and sits in an elevated position above the surrounding imbricated clasts.

As clasts and cobbles are exposed on the surface and they are coated (patinated) by desert varnish, a dark brown manganese oxide, on their upper surface. Orange goethite (or limonite), hydrous iron oxide, often coats the buried lower surface. Desert varnish appears to be more pronounced on ferromagnesian rocks (e.g., basalt, hornfels, and schist) than on dense siliceous rocks and quartz. The quartz cobbles remain a lighter color due to differential patination and, therefore, are more visible than their dark, varnish-coated ferromagnesian counterparts. The quartz cobbles do, however, receive desert varnish that coats the exterior surface and precipitates along the walls of interior fractures. Upper, lower, and interior color stains are important for determining the relative age and history of disturbance of the quartz cobbles. If the individual components of the fractured cobble remain in relative positions, show dark stains on outer and interior surfaces, and if no orange stains are showing from the surface, the cobble and its components are in their original position. If white quartz exhibits no exterior or interior dark stains or if orange coloration is upward and dark brown is down, then part or all of the clast has been disturbed. The latter situation is readily apparent on cobbles located in the paths of tracked vehicles.

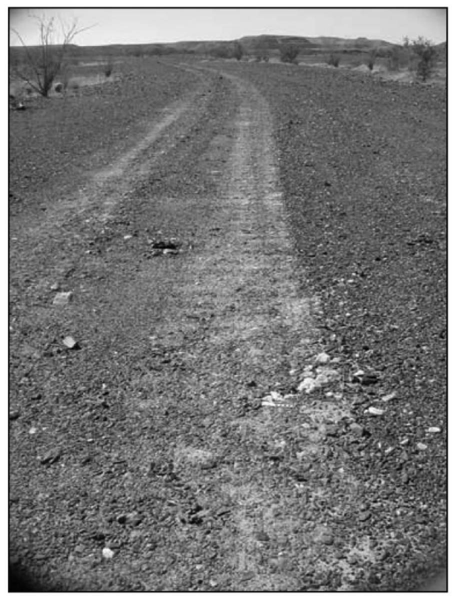
(H-4) Hypothesis 4 is that the quartz shatters were natural in origin, but had acquired cultural significance. Hypothesis 3 in this study confirms that the quartz shatters are the result of natural processes. But, could these natural phenomena have acquired cultural significance as well? Evaluation of H-4 is inconclusive based on available data, primarily because it addresses comparative models that have not previously been considered for the prehistory of southern California.

The presence of the concentration of quartz shatters may have attracted prehistoric peoples to the area. Or it is possible that the presence of the concentrations of quartz shatters created sterile zones that were “off limits”

Photograph 5: *View of fractured Miocene olivine basalt block.*

Photograph 6: *View of two foot diameter shattered quartz boulder. All internal fractures and external surfaces have manganese oxide stains or light patina of desert varnish, indicating antiquity of fracturing.*

Photograph 7: *View of fresh breaks on quartz block are bright milky white.*



Photograph 8:

View of quartz block with fresh white breaks showing tracked vehicle impressions.

for habitation and exploitation. Ethnographically, such off-limits locations exist on the Costa Arriba of Panama and protohistorically they existed among the chiefdoms of the southeastern United States. In both the Costa Arriba and the southeastern United States the uninhabited areas served two main purposes: to avoid conflicts between neighbors and to create island of ecologically stable plants and animals that were not subject to human depredation. Could the same behaviors have pertained to the quartz shatters at MRL? Perhaps. Hayden (1967:338) wrote about the Sierra Pinacate, 185 miles southeast of the Chocolate Mountains:

It may here be apropos to note that an almost superstitious fear of the Pinacate as a forbidden area full of danger and hardships and peopled by hostile Indians, has come down almost to the present among the Mexicans and Papagos of the surrounding country.

The Sierra Pincate was on the main salt gathering route between Pima–Papago country and the Gulf of California. We know from other ethnographies that certain trails and areas were avoided because of social taboos and fears of conflict.

Further research will be necessary to explore such possibilities. The first step will be to remap the known concentrations of quartz shatters with relationship to recorded trails, trail segments, and associated features.

Conclusions

Erosional deflation creates a flat, imbricated desert pavement surface where large, relatively resistant cobbles sit at elevated positions. All large cobbles that have passed through the deflating soil profile are fractured when they reach the surface, regardless of their lithology. All clasts show dark patination on the upper surfaces and orange on the lower surfaces. Quartz cobbles do not readily accept the dark desert varnish patination, but all quartz cobbles examined showed dark stains on exterior surfaces and interior fractures. The only difference between the quartz cobbles and cobbles of other lithologies on the pavement surface is their lighter color. This pronounced color difference caused previous studies (von Werlof 1984) to focus preferentially on light-colored cobbles while disregarding the darker cobbles that blended with the varnished desert pavement.

This study suggests that all large cobbles rising to the surface by deflation of the pavement terrace have been fractured along zones of weakness, perhaps early in the many-thousand-year development of the soil profile. At the surface, large, resistant cobbles sit at elevated positions, and are coated by dark brown varnish on upper and interior surfaces, and by orange stains on buried, lower surfaces. Recent disturbance of the fractured cobble components can be determined if the colored surfaces are out of place, or if fresh, unpatinated surfaces are developed by recent impacts.

This research also presents the possibility of a cultural landscape. The confirmation of H-3 clearly demonstrates the value of archaeologists and geologists collaborating on research involving cultural artifacts or problems, and lithic formation processes and source areas.

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Trade in the Mojave: taking a stab at flaked glass

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Background

Recent studies have indicated that examination of the routes by which historic items were traded by Native Americans, and the analysis of these items in archaeological contexts with prehistoric items, may bring to light a longer persistence of Mojave Desert prehistoric culture than previously understood. This study was prompted by a late prehistoric site recorded on a property within the high desert region of the Mojave Desert in traditional Serrano territory (Beck and Haase 1974; Heizer 1978; Kroeber 1925) near Apple Valley, California (temporary site number LSA-CBH531-H1). A record search of the surrounding area revealed that, although seven reconnaissance surveys had assessed over 2000 surrounding acres, this small lithic scatter represents the only recorded archaeological resource of any kind within an approximate two-mile radius. Examination of Government Land Office maps indicated that the site is located approximately six miles east of the Salt Lake Road (also called the Old Mojave Road, which eventually became US Route 66) which fronts the Mojave River (Survey General Office 1857). At its point nearest the site, this road occupies the same approximate alignment as an ancient trade route that locals used to cross the desert for trade and communication.

A systematic survey revealed that individuals had salvaged some shattered amethyst bottle glass from a historic refuse pile, and knapped it into seven informal tools (artifacts 1-7). These were found near two andesite flakes, manufactured from naturally occurring material (artifacts 8 and 9).

The artifact assemblage is given in Table 1 (from field notes taken by Kyle Karnes, August 10, 2005).

The sun-colored amethyst flaked glass artifacts were located on the surface within an area measuring approximately four square meters, approximately 20 meters west of their presumed source (refuse piles). The andesite flakes were located approximately 30 meters south of the flaked glass scatter. Four surface shovel scrapes yielded no amethyst micro-flakes in the immediate vicinity, and indicated very little depth (≥ 8 centimeters). No midden was observed in the vicinity. The area surrounding the lithic scatter contained several historic refuse piles indicating sporadic use over a period of approximately 70 to 100 years. The earliest of these piles contained shattered sun-colored amethyst bottle glass containing seed bubbles, identical to the above detailed flaked glass.

William Andrefsky Jr.'s work on prehistoric lithics states:

> The best kinds of stones for knapping are those that can be cracked in a reliable and predictable manner; such stones are brittle, homogeneous, and isotropic. In other words, the stones most suitable for flint knapping are those that are brittle and do not have direction-dependent properties such as bedding planes, fissures, cracks, or inclusions. Natural glass or obsidian is probably the best example of this material because it can be manipulated to crack in any manner the tool maker desires (Andrefsky 2005:24).

Glass containing manganese, which becomes amethyst in color when exposed to the sun, was commonly manufactured in the United States between 1880 and 1914 (Polack 1994:25). Contemporaneous metal and other types of glass present within the same refuse pile showed no signs of alteration or use. The andesite, sparsely scattered over the surrounding terrain in cobble form, was coarse-grained and contained phenocrysts. In short, according to Andrefsky's above prescription for preferable materials, the

Table 1. LSA-CBH531-H1 Artifact Assemblage

Artifact #	Material*	Type	Size (cm)
1	SCA glass	Unifacially flaked curved blade	6 x 1.5
2	SCA glass	Unifacially flaked scraper with seed bubble	7 x 6
3	SCA glass	Bifacially flaked and retouched glass scraper	5 x 3.2
4	SCA glass	Triangular fragment, point unifacially sharpened (drill?)	3.5 x 3.5
5	SCA glass	Unifacially flaked curved piece (end scraper)	2.5 x 2.3
6	SCA glass	Unifacially flaked/serrated curved piece	2.5 x 1.9
7	SCA glass	Unifacially flaked piece, slightly curved	3.5 x 2.5
8	Andesite	Grayish green flake with phenocrysts; platform intact, termination missing	5 x 4
9	Andesite	Grayish green flake w/phenocrysts; no platform, feathered termination	3.5 x 2

*SCA=sun-colored amethyst

andesite was less than ideal, and was therefore probably locally used for expedient purposes, not as a source of wealth or a trade item. A higher occurrence of knapped glass may indicate a preference for fine-grained material due to previous experience with higher quality crypto-crystalline silicate materials such as chert, chalcedony, and jasper, or obsidian, none of which were naturally present in the immediate vicinity.

The correlation of flaked historic glass salvaged from a historic refuse pile and poor quality lithic material with relatively close proximity to a major prehistoric and historic trans-Mojave Desert thoroughfare has prompted the following speculation on a sequence of possibilities.

Assuming that little fine-grained native lithic material was available for the production of chipped stone tools, or that access to traditional quarries had been cut off, one might be delighted to happen upon such high quality glass and, immediately recognizing its superiority, begin happily knapping away to produce a high quality, and uniquely (and enviably) colored stone tool, or better yet collect quantities of the material and trade it along the Old Mojave Road. Although the date range of the artifacts in question (i.e. 1880-1914) is after most facets of traditional prehistoric Serrano economy had collapsed (Heizer 1978:573), rudiments of Serrano populations still practiced traditional forms of trade and communication within the surrounding area (Heizer 1978:573), and were not incorporated into a reservation system until the passage of the Act for Relief of Mission Indians in 1891 created the San Manuel Reservation (San Manuel 2006). Could similar flaked glass demonstrate rudiments of prehistoric trade and economy persisting into the late 19th and 20th centuries?

Research into previous work conducted in California on this topic has indicated that trade and economy were rarely significant considerations where historic era lithic industry was concerned. Stephen Silliman's research at the Petaluma Adobe State Historic Park has indicated that "the production and use of stone tools in a colonial world indicates 1) a material necessity to overcome a scarcity of tools, 2) a maintenance of material comfort and familiarity and 3) a political social statement about identity and gender" (Silliman 2004).

Rebecca Allen's 1998 study of Indian Neophytes (baptized Indians) at Mission Santa Cruz has found that Neophytes continued to manufacture stone tools within their living areas throughout the mission era. She found that informal flaked stone tools outnumbered formal tools from this era by a greater margin than during prehistory. Fewer formal tools would seem to indicate that fewer unsupervised traditional specialized tasks were being performed. It therefore seems likely that formerly specialized tasks using traditional tools and techniques (i.e. stone tools used for hunting, and hide and vegetal processing) had been somewhat replaced by historic tools and techniques. New methods and tools have been associated with a combination of new and traditional subsistence strategies and economies (Allen 1998), and no evidence has been suggested to associate the new behavior with a reversion to the use of traditional trade routes.

Michael Sampson and Jill H. Bradeen presented Historic-Period Lithic Technologies in Old Town San Diego at Sacramento's 2006 Society for Historical Archaeology Annual Conference on January 13, 2006 in which they compare historic period post-contact lithic tool use with prehistoric lithic tool use. They answer questions related to the persistence of prehistoric tool use in a "19th century commodities-oriented economy" (Sampson and Bradeen 2006:1). In addition to exploring topics related to general and specific changes in subsistence strategies and residence patterns, the authors ask whether "traditional pre-Contact trading networks broke down in historic times…" (Sampson and Bradeen 2006:7). They offer evidence pointing to significantly higher percentages of chipped stone tool material sourced within or next to their study area (i.e. Old Town San Diego) during post-contact, compared to samples during pre-contact times. In other words, a higher diversity of pre-contact lithic material sources indicates a vibrant trade and traditional economy. They conclude that this indicates a breakdown in traditional trade routes that made a more diverse lithic tool assemblage available before European contact.

I propose that techniques described by Sampson and Bradeen may most aptly be applied to the question of historical cultural persistence by the Serrano. This will be explained in greater detail, below.

Although there exist some examples of native groups practicing traditional subsistence strategies near their aboriginal homelands long after most Californian Native Americans had succumbed to disease, massacre, relocation, and acculturation, their trade routes and economies had long since been almost completely destroyed or adapted from as a result of contact with Anglo economies (see also studies of Kumayaay in San Diego County [Shipek 1991] and Yahi in Lassen County [Kroeber 1961]). While little ethnographic information is available for late prehistoric and early historic subsistence strategies and residence patterns of Serrano people in the western Mojave, the archaeological record can help illuminate the subject.

Preliminary Results

While Silliman and Allen's earlier studies have helped organize frameworks in which post contact lithic studies can take place, I have found Sampson and Bradeen's techniques most useful to illustrate a temporal framework for the breakdown of traditional trade regionally within the Mojave, and specifically at LSA-CBH531-H1. A temporal association between the glass and andesite artifacts in a surface context, such as the one represented here, can be problematic, though considering the total lack of any other archeological material in the vicinity, the relationship appears likely. Given the site's close proximity to a well-documented trade route, people using prehistoric technology to manipulate items that were not traded probably did not have easy access to trade. In other words, by the time these artifacts were manufactured, it seems likely that trade had broken down, although some adherence to traditional prehistoric lithic technology remained. While the sample is admittedly miniscule, the higher number of glass artifacts than andesite indicates a preference for glass working. Since glass, obsidian, and good fine-grained cryptocrystalline silicates do not occur naturally in the vicinity, though they do appear to be preferable, it may be inferred that the individuals that produced these artifacts had experience with and, formerly, access to these materials. The late date range of the historic glass (i.e. 1880-1914), and the ability to work with a range of materials (including andesite) is indicative of a very late prehistoric and accomplished craftsman that must have been enculturated by a well functioning, complex and integrated cultural system. It is therefore postulated by this sequence of possibilities that fully integrated Serrano Indians (or other displaced Native Americans) were practicing traditional methods of lithic processing into the late 19th and possibly the early 20th centuries in the Mojave. Although the lack of other resources in the vicinity indicates that the current site's deposits probably occurred after traditional economy and trade had mostly broken down, the range of craft exhibited by the producer of these artifacts shows that the methods may have been recently learned, certainly within one generation, or may concurrently remain intact elsewhere in the Mojave.

Further Research

This sequence does not serve to propose a new system of absolute dating within the Mojave Desert. It does, however, point to a new method of analyzing late prehistoric sites in the Mojave that have been demonstrated to contain stratigraphy. I propose a re-examination of existing data available at Serrano village sites, beginning with the villages of Amutskupiabit, Guapiabit, and Atongaibit. Although these were visited by the Spanish into the early 1800s (DeBarros 2004), and considered subsequently abandoned, they were situated in areas of high importance to the Serrano and may have been seasonally visited much later. The research will focus on studying prehistoric lithic items recovered in an archaeological context with known, and ideally easily and accurately dateable (i.e. flaked glass, metal tools, etc.), historic items. The lithics that are found in the same deposits as historic items should then be quantified by lithic source, for comparison with sources of squarely prehistoric lithics. If Sampson and Bradeen's methods apply, we should expect to find a reduction in the diversity of lithic sources somewhere during the historic period. Correlating a lack of diverse lithic sources with dateable historic items will allow for a closer temporal approximation of the breakdown of traditional trade. Due to the late historic entry into the Mojave by Europeans on a permanent basis, these dates may indicate a later breakdown, and longer adherence to, traditional cultural practices than has been found for non-Mojave prehistoric populations of California. Furthermore, if a later occupation of traditional Serrano economy is inferred, interviews of living Serrano, supplemented by earlier ethnography, could help crystallize a temporal framework.

Other implications of this research may be applied to questions of when prehistoric trade broke down between coastal, Mojave and Southern Desert, and Colorado River groups. Where and when, for instance, in an archaeological context, do coastal items that were at once common trade items at desert sites disappear? Significant documentary evidence exists for Californian prehistoric coastal people practicing some trade with European explorers as early as 1542 (see Lightfoot and Simmons 1998). What are the earliest European items that make their way into the Mojave, and were these items numerous enough to leave a shadow in the archaeological record? If found, can any of these historic items, dateable by manufacturing techniques rather than chemical means, actually alter dates of significant European influence in the American interior prior to actual colonial exploration? Finally, can the application of Sampson and Bradeen's 2006 methods be applied to disparate groups in similar remote locations, to make statements about persistence of traditional economies and trade? These questions can largely be addressed by a reexamination of documentary sources as well as existing museum collections, resources that should be exhausted before any new excavation takes place.

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Abstracts of Proceedings

The 2007 Desert Research Symposium

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A primitive Early Miocene platanistoid dolphin (Cetacea: Odontoceti) from Cajon Pass, San Bernardino County, California

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An unusual and very significant dolphin fossil has been discovered in an outcrop of the Early Miocene marine Vaqueros Formation in Cajon Pass, southern California. The fossil was collected from a section of the Vaqueros Formation that unconformably overlies crystalline basement rocks, is in fault contact with the Paleocene San Franscisquito Formation, and is unconformably overlain by the Middle Miocene (Hemingfordian to Barstovian) Cajon Valley Beds. The dolphin was associated with fossils of elasmobranchs, the echinoderm Scutella fairbanksi, and mollusks Crassatella granti, Ostrea titan subtitan, and Pecten sespeensis. The presence of Turritella inezana is important, and establishes an Early Miocene age for the marine sediments. The dolphin is an extremely long-snouted member of the archaic superfamily Platanistoidea. Platanistoids, today represented only by the rare and much-endangered fresh water Ganges River Dolphin, were in early Tertiary time more diverse and relatively widespread, being found as fossils in marine and estuarine deposits all ocean basins of the World. The fossil specimen from the Cajon Pass is a partial skeleton of an animal that had a skull that was nearly a meter long, a body that was approximately four meters long, a very flexible neck, and relatively long forelimbs (flippers). It appears to be most closely related to a poorly known platanistoid, Allodelphis pratti Wilson, 1935, which has been reported from earliest Miocene sediments that crop out north of Bakersfield, Kern County, central California. It differs from that species by having some more highly evolved characters of its skull and by having larger forelimb bones. The lineage of dolphins to which these animals belonged is known only from the North Pacific Ocean basin, including Japan, Washington, Oregon, California, and Baja California, and they have a geochronologic range from Early Miocene to Late Miocene. The specimen is housed in the collections of the San Bernardino County Museum.

The use of cosmogenic nuclide dating to determine the age and erosion rates of alluvial fans at the Calico Site, Mojave Desert, California

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The use of cosmogenic nuclide dating of surface boulders and alluvial sediment often can be used to solve dating problems in desert environments. Preliminary measurements of the abundance of cosmogenic nuclides from surface boulders and depth profiles of alluvial fans located at the Calico Early Man site indicate a surface age of 70,000–250,000 years, and an erosion rate of approximately 6 mm per 1,000 years. Analyses of boulder samples and new depth profiles recently obtained in November 2006 should serve to narrow this age range. Additionally, cosmogenic nuclide concentrations of bulk sediment samples collected from washes contained within the alluvial fan basin should give an independent assessment of the average erosion rate and thus confirm both the surface age of the alluvial fans, as well as the erosion rate.

Late Miocene–Early Pliocene transition from lacustrine to fluvial deposition: inception of the lower Colorado River in Southern Nevada and Northwest Arizona

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Several basins in southern Nevada and northwestern Arizona record a transition from lacustrine to fluvial deposition between ~5.6 and 4.2 Ma that represents local inception of the Colorado River (CR). In the Lake Mead (LM) region near the modern CR, a series of lakes dominated the late Miocene landscape. In the eastern LM region, the Hualapai Limestone (HL) accumulated in such lakes in the Grand Wash trough, just west of the mouth of the Grand Canyon, and in northern Detrital basin. Geochemical, petrographic, and paleontologic data demonstrate a lacustrine origin for the HL. $^{40}Ar/^{39}Ar$ and tephrochronologic data bracket the age of the HL between 11 and 6 Ma. It is one of the youngest deposits formed prior to integration of the eastern LM region into a through-flowing CR. Sr isotopic data from the HL ($^{87}Sr/^{86}Sr$ to 0.7196; data from J. Patchett, 2002) imply that source waters for the HL lake differed significantly from modern CR water and may have been derived largely from groundwater issuing from the western Colorado Plateau and/or central Nevada carbonate aquifer. In the Grand Wash trough, a 4.4 Ma basalt flow is intercalated in CR gravels. In the western LM region, a lacustrine limestone just north of Frenchman Mountain rests on the ~5.6 Ma Wolverine Creek tephra and interfingers eastward with a gypsum deposit that extends to near the modern CR. These relations bracket CR inception in the LM region between ~5.6 and 4.4 Ma. To the south in the Lake Mohave and Laughlin-Bullhead City area, alluvial fans dominated the late Neogene landscape until ~5.6 Ma when small lakes formed. Near Bullhead City, the ~5.6 Ma Wolverine Creek tephra lies directly below a thin limestone. Nearby, younger CR gravels contain the 3.6-4.2 Ma "lower Nomlaki" tephra (House et al., 2005). Thus, inception of the CR in the Lake Mohave area is bracketed between ~5.6 and 4.2 Ma, similar to that in the LM region. These constraints are compatible with initial deposition of CR-derived sediments between 4.5 and 5.3 Ma in the Anza-Borrego Desert region (see Dorsey, 2006). Furthermore, these relations indicate relatively rapid, regional inception of the CR in the early Pliocene. Available data suggest, however, that lacustrine deposits in the western LM and Lake Mohave regions are slightly younger than the HL in the eastern LM region. This may imply that the lakes partly served as temporary sinks for CR water and were progressively formed and breached in a cascading downstream sequence.

Cenozoic evolution of the northern Colorado River extensional corridor, southern Nevada and northwest Arizona

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The northern Colorado River extensional corridor (NCREC) is a 70 to 100 km-wide region of moderately to highly extended crust along the eastern margin of the Basin and Range province in southern Nevada and northwestern Arizona. It has occupied a critical structural position in the western Cordillera since Mesozoic time. In the Cretaceous through early Tertiary time, it stood just east and north of major fold and thrust belts and also marked the northern end of a broad, gently (~15o) north-plunging uplift (Kingman arch) that extended southeastward through much of central Arizona. Mesozoic and Paleozoic strata were stripped from the arch by northeast-flowing streams. Peraluminous 65 to 73 Ma granites were emplaced at depths of at least 10 km and exposed in the core of the arch by earliest Miocene time.

Calc-alkaline magmatism swept northward through the NCREC during the early to middle Miocene, beginning at ~22 Ma in the south and ~12 Ma in the north. The most voluminous volcanism occurred prior to major east-west extension in early Miocene time but was accompanied by mild north-south extension. Major east-west extension followed the initiation of magmatism by 1 to 4 m.yr., progressing northward at a rate of ~3 cm/yr. Large-magnitude east-west extension engulfed nearly the entire region in middle Miocene time, beginning in most areas ~16 Ma and ending by ~9 Ma. A change from mainly intermediate to bimodal volcanism roughly coincided with the onset of major east-west extension. Tilt rates commonly exceeded 80o/m.yr. during the early stages of east-west extension. South of Lake Mead, volcanism generally spanned the entire episode of extension. Thus, thick volcanic sections, rather than sedimentary rock, accumulated in many growth-fault basins.

During middle Miocene extension, strain was partitioned into a west-dipping normal-fault system in the north and an east-dipping system in the south. The two fault systems and attendant opposing tilt-block domains overlap and terminate within the generally east-northeast-trending Black Mountains accommodation zone. Major east-west extension was contemporaneous on either side of the accommodation zone. The west-dipping normal fault system in the north is kinematically linked to major strike-slip faults along the northern margin of the corridor, where a complex three-dimensional strain field, involving both east-west extension and north-south shortening, characterized the middle to late Miocene. The east-dipping system in the south linked southward with major east-dipping detachment faults that flanked several metamorphic core complexes (e.g., Chemehuevi and Whipple Mountains).

The transition between the Colorado Plateau and the Basin and Range is unusually sharp along the eastern margin of the NCREC and is marked by a single west-dipping fault zone, the Grand Wash fault zone. Subhorizontal, relatively unfaulted strata on the Colorado Plateau give way to moderately to steeply east-tilted fault blocks across this zone. Topographic and structural relief across this boundary developed during middle Miocene extension and was established by 9 Ma. The location and abruptness of the Colorado Plateau-Basin and Range transition in this region may have been controlled by an ancient north-trending crustal flaw, inasmuch as it follows a diffuse boundary between Early Proterozoic crustal provinces.

The Colorado River arrived in the NCREC in latest Miocene-early Pliocene time following a long period of internal drainage. Thick sections of alluvial fan, continental playa, and lacustrine deposits accumulated in some basins until ~6 Ma. Widespread late Miocene lacustrine and evaporite deposits in the Lake Mead region suggest a large influx of fresh water. However, available age constraints indicate that a through-flowing Colorado River developed between ~5.6 and 4.4 Ma.

Tectonic elevation of Pliocene–Pleistocene debris flows and the underlying Calico Member of the Barstow Formation in the Yermo Hills: work in progress

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The study area is located ten miles East of Barstow, CA at the South East corner of the Calico Mountains. The Yermo Hills are a truncated set of two fanglomerate structures, the East Fan and the West Fan, that are separated by an unnamed northwest trending fault that runs diagonally between them. Together they are called the Yermo Fanglomerates. They appear to have been truncated from their older erosional sources by right-lateral movements along the Tin Can Alley Fault Zone. The two structures differ in form and some elements of their composition. The fanglomerates are an exposed set of debris flows and secondary erosional debris derived from Mesozoic intrusive and Miocene extrusive volcanics and reworked Miocene Barstow Formation materials that arise in the eastern Calico Mountains. The Fans extends considerably eastward and southeastward beyond the Calico Mountains range front to form the Yermo Hills. The eastern margin of the Eastern Fan abuts the former channel where the Mojave River flowed into Coyote Lake. The Eastern Yermo Fan has probably been incised by high-stand shorelines of Lake Manix. The Yermo Fanglomerates unconformably overlie the green lacustrine siltstones, limestones and water-laid ash deposits of the Miocene Calico Member of the Barstow Formation. In the porous Eastern Fan deposits, the Miocene siltstones may act as an aquatard, causing precipitation of pedogenic carbonate and other minerals. This paleosol layer is only rarely encountered in the Western Fan. In addition, in the Western Fan, perched springline barium mounds and dispersed hydrothermal replacement deposits along the Fanglomerate-Barstow contact indicate the presence of likely faults or fault splays that run through the Western Fan near the archeological site. Previous mapping by others suggests a Pliocene/Pleistocene age for deposition of the Eastern Yermo Fanglomerate debris flows. The antiquity of the Yermo Fans is suggested by the truncation and beheading of the fans, dendritic shape due to erosion, varnished, rubified and pavemented surfaces, and Lake Manix shoreline incision. Previous U/Th dating of the carbonates in the Western Yermo Fanglomerate from the excavation pits at the Calico Early Man Site of 200 ka suggests a late Pleistocene age for the base of those deposits. This leaves unanswered the question of where the debris from erosion of the topographically high Calico Mountains went since late Miocene and Pliocene times, during the Mojave regional Pliocene erosional event. Perhaps the Calico Mountains were not as elevated as a clastic source during that time interval. Since its deposition, the western portion of the Western Yermo Fan where we see the exposed contact of fanglomerate and Barstow Formation in the Green Ash Hills, appears to have been elevated above the current erosional sources in Mule Canyon. The Eastern Fan is eroding in a radial fashion off the central high at the present time while the eroding lobes of the Western Fan extend in primarily a southerly direction. Evidence for Western Fan elevation can be seen along the north-south Powerline Road at the Green Ash Hills where the contact of the Fanglomerate and the Barstow Formation in a faulted plunging anticline sits at 2,400 feet elevation, 80 feet or more above similar contacts one mile to the west. This research looks at the following questions:

Is the higher elevation of the top of the Western Yermo Fan relative to a presumed source the result of erosional events in the Calico Mountains that post-date fan deposition?

Have the Yermo Fans been tectonically elevated by dynamics of the Manix, Calico and Tin Can Alley fault zones and regional clockwise vertical-axis rotation of the fault-bounded blocks in the region?

Current erosional dynamics would make it difficult to deposit fan debris on the top of the elevated Green Ash Hills, yet the base of the Western Yermo Fanglomerate exists there unconformably atop the Barstow Formation. Currently the younger source debris washing out of Mule Canyon in Coyote Wash is deflected around the north edges of the Yermo Fans. The Mojave River was also deflected to the south and east of the Yermo Fans as it flowed into Coyote Lake. The Western Fanglomerate may have been, and likely was, much thicker originally, and the western portion reduced by erosion to its present lower elevation. But that is not typical of local fans. Other, shorter, alluvial fans that skirt the southeastern Calico Mountains do not have an elevated medial or distal portion. The short fans suggest that the baseline may have been dropped along there west of the Tin Can Alley Fault Zone. The exposed base at the Fanglomerate-Barstow contact in the Green Ash Hills also suggests that portions of the exposed baseline of the Yermo Fans east of the Tin Can Alley Fault have been elevated since the time that sediments were laid down. This study is exploring a tectonic geomorphology model for uplift on portions of the Western Yermo Fanglomerate east of the Tin Can Alley Fault. The authors will describe lithologies of the Yermo Fanglomerates and document structural elements, especially the internal faults and folds, to map details of the Western Fan. In addition we are working to establish provenance for the debris flow materials, whose clasts include exotic granites, gneiss and schist which point to a more northern Mesozoic sources in the Calico Mountains west of the Tin Can Alley Fault that has since been translated along the fault zone. These exotic clasts are presently found in the fanglomerate materials in both fans but with a much higher percentage of the exotic granitic clasts in the Eastern Fan. The granitics and metamorphics are present in addition to the Pickhandle and Yermo dacitic volcanics and reworked Barstow Formation clasts that are the major components of the matrix-supported debris flows. Mapping of the basal contacts of the debris flows and exposed lower Barstow Formation marker beds will help define the topography and indicate structural changes. Attitudes on debris flows and erosional bedding planes, including those exposed within excavation of the Calico Early Man Site will be plotted to describe the internal structures of the Western Fan. If these attitudes are more extreme than those typical of undeformed course debris flow deposits, compression might be suggested between the movements of the sinestral Manix Fault and the dextral-olique Calico Fault and dextral-oblique offset with "apparent vertical components" along the Tin Can Alley Fault Zone and their related fault splays. Uplift of the Yermo Hills may be the expression of the combined dynamics at the extremely complex intersection of the three major fault zones plus the effects of vertical-axis clockwise rotation of the local fault-bounded block in the Eastern Mojave Shear Zone since the beginning of the Pleistocene.

The state of the Colorado River ecosystem in Grand Canyon: lessons from 10 years of adaptive ecosystem management

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The year 2005 marked the 10th anniversary of the completion of the Final Environmental Impact Statement (EIS) on the Operation of Glen Canyon Dam on the Colorado River. A decade of research and monitoring provides an important milestone to evaluate the effects of dam operations on resources of concern and determine whether or not the desired outcomes are being achieved, or if they are even compatible with one another or not. A comprehensive effort was undertaken to assess the state of scientific knowledge about the resources of concern, as identified in the EIS. The result was the first systematic attempt by scientists to conduct an assessment of the changing condition of Colorado River ecosystem resources in Grand Canyon over a decadal timeframe (http://www.gcmrc.gov/products/score/score_reports.htm). In the EIS, 30 resource attributes are listed along with predictions for how those resources would respond under the Secretary of the Interior's 1996 Record of Decision, an operating prescription based on the preferred alternative of Modified Low-Fluctuating Flows (MLFF). Because of a lack of data or subsequent analyses to confirm whether some predictions stated in the EIS were correct, or not, 14 or 47 percent of the outcomes, are essentially unknown. Excluding outcomes that are unclear, then the remaining predictions in the EIS were correct in 7 out of 16 outcomes, or 44 percent of the categories listed. Mixed outcomes occur in 4 out of 16, or 25 percent of the categories, and failed predictions, occur in 5 out of 16, or 31 percent of the categories. As such, less than 50 percent of the outcomes were predicted correctly, underscoring the uncertainties associated with working in a large complex system with few to no long-term data sets. Similar uncertainties are faced by all resource managers charged with ecosystem restoration globally. The acceptability of this kind of uncertainty is influenced by interpretation, societal values, agency missions and mandates, and other factors. However, failure to correctly predict the future, in and of itself, is not deleterious under the paradigm of adaptive management where large uncertainties provide opportunities for learning and adjustment through an iterative process of "learning-by-doing" (Walters and Holling, 1990). Although recent studies have documented a continued decline of environmental resources of the Colorado River below Glen Canyon Dam, it has also identified options that might still be implemented by managers to achieved desired future conditions in Grand Canyon have also been identified.

Field guide to the San Andreas Fault

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Everyone has heard of the San Andreas Fault but few people outside the geology community know where it is. Nor do they know what to look for and how to see the fault. Until recently, there were no field oriented books for the general public that provided detailed information on the fault's location and accessibility. To remedy this situation, I have written 12 one-day driving trips along the fault reaching from Cape Mendocino to the Salton Sea. There are over 1100 miles of mile-by-mile road logs, and GPS coordinates for hundreds of fault features (scarps, pressure ridges, sag ponds, offset streams, etc.). Four-wheel drive is not required for any of the routes. The routes are accompanied by maps, color photographs and side trips to seldom-visited locations. Several introductory chapters review plate tectonics, fault-related landforms and they emphasize that the fault is part of a much larger fault zone that is associated with the plate boundary. An appendix lists the fault coordinates at intervals of one arc minute of latitude so that people can make their own maps. In this presentation intended for educators, leaders of geology field trips, and the amateur geologist, I will review highlights from the southern California desert routes.

Why overflow best explains the Colorado River's course

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Transverse drainages can originate in only four ways: antecedence, superposition, piracy, and overflow. All four hypotheses have been used to explain the course of the Colorado River through the Grand Canyon vicinity. Today, there is plentiful evidence that antecedence and superposition can be discounted, leaving only piracy and overflow as possible explanations. Piracy has been the favored explanation in most geologic publications during the latter half of the 20th century.

The course of the Colorado River downstream of the Grand Canyon argues strongly for ponding and overflow as the preferred hypothesis. In fact, ponding and overflow was the first scientific hypothesis proposed for the river's path by Newberry, who studied both the upper and lower reaches of the river. Subsequently, the issue became muddled by some of the biggest names in geology who worked primarily on the Colorado Plateau.

Piracy (i.e., headward growth of the Colorado River beginning at the Gulf of California) can be discounted for many reasons. Some of these reasons include: 1) the process of headward drainage growth is exceptionally slow (and in this case insufficient); 2) all streams and tributaries in a drainage basin should exhibit similar headward growth vigor; 3) it is highly improbable that a single stream will grow headward across multiple elevated bedrock structures to capture internally drained basins; 4) if headward growth occurred, it did not take any of the routes that a headward growing stream should have taken in this region; 5) and the existing sedimentary record is not consistent with a headward-growing stream.

Using geographical and geological evidence downstream of the Grand Canyon, I will present a strong case that the Colorado River and Grand Canyon must have originated by ponding and overflow, and the available evidence is entirely consistent with such a hypothesis.

The reproductive consequences of variable sex expression in desert holly (Atriplex hymenelytra).

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The common desert shrub, Atriplex hymenelytra, is considered to be a dioecious species. However, occasionally a few male plants will also produce some female flowers and fruits. The expression of both staminate and pistillate flowers among these "inconstant males" is thought to be related to genotype and/or environmental conditions. Preliminary results will be presented relative to the following questions: (1) Are seeds and bracts produced by females and inconstant males morphologically similar, (2) are seeds produced by females and inconstant males equally likely to germinate, and (3) are resources allocated similarly to the production of reproductive (versus vegetative) structures in females, males, and inconstant males?

Evaluation of Late Pleistocene ground-water discharge deposits in the Mojave Desert, eastern California

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Ground-water discharge deposits, also called "paleowetland" or "spring" deposits, form in arid environments as water tables rise and breach the ground surface during periods of enhanced effective precipitation. In addition to providing an important water source for local fauna, emergent water tables support hydrophilic vegetation, which in turn acts as a natural catchment system for eolian sediments. The interplay between emergent water tables, ecological and biological systems, and wind-derived sediments results in a unique and complex depositional environment that contains information on the timing (age of deposits) and magnitude (faunal and ostracode assemblages, isotopic data) of past climate change. Spring deposits also clearly demarcate the position of past water tables on the landscape, which provides direct evidence of past hydrologic conditions.

Although much of the Mojave Desert is extremely arid today, during the Late Pleistocene the region supported a large, interconnected lake system, as well as numerous springs and wetland systems. Approximately 130 localities that contain fossil spring deposits have recently been identified in the Mojave Desert by David M. Miller and colleagues during mapping efforts supported by the U.S. Geological Survey. Three of these localities, Valley Wells, Piute Valley, and Fenner Wash, have been chosen for investigation in the initial phase of this project. The primary goals of this initial phase are to (1) establish a stratigraphic framework for the deposits at each locality, (2) determine the ages and rates of deposition of the deposits, and (3) reconstruct local paleoenvironmental conditions using sedimentology, fossil ostracodes and gastropods, vertebrate remains, geochemical indices, and/or stable isotopic analysis.

These data and interpretations will then be evaluated to better understand the paleohydrologic response of local spring systems to regional and/or global climate change. The results of this study will also allow us to better understand the magnitude of natural variability in water table elevations in desert biomes in the American Southwest through time, and to improve our understanding of desert response to past climate change in order to facilitate management of this fragile ecosystem in light of future climate change.

Sand, sand lizards, sand people and sand management

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The increase in human population and the availability of expendable income has resulted in an increase in the demand for recreational areas in southern California. The growth of the Los Angels area to the west, Las Vegas to the northeast and the Phoenix area to the southeast has contributed to the increase use of the east Mojave Desert as a prime area for OHV activities.

Due to actions by Congress, federal land management agencies have prepared resource management plans for all federal lands in California. These plans try to structure the use of the land for multiple uses. As a result of these plans many species are currently being studies to learn how to best manage the species and their habitats .This presentation discusses one such study on the Mojave Fringed-toed lizard (FTL) in northeastern San Bernardino County.

The results of the first year of a five year study to establish basic population dynamics of the FTL will be presented. Modifications and plans for the 2007 season year's effort will be discussed.

Pliocene angiosperm hardwoods and recycled Cretaceous palynoflora of the ancestral Colorado River, Anza-Borrego Desert State Park, California: a review

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The Vallecito-Fish Creek Basin of Anza-Borrego Desert State Park includes a continuous time-stratigraphic sequence of lower Pliocene pro-delta/delta-front (Coyote Mountain Clays and Yuha Formation) and delta-plain (Palm Spring Formation) sediments sourced from the Colorado Plateau and deposited by the ancestral Colorado River. Exposed along Fish Creek Wash, these fine-grained, extralimital sediments yield locally-derived silicified angiosperm fossil wood (dicotyledons and monocotyledons) of the Carrizo Local Flora within the Palm Spring Formation, and reworked Cretaceous pollen zoned throughout the stratigraphic section. Based on tracheid cell structure, seven families of fossil wood are recognized in the paleoflora, representing eleven taxa. Eight are new for the area. Families include the Lauraceae, Salicaceae, Oleaceae, Hippocastanaceae, Arecaeae, Juglandaceae, and the Cupressaceae. Climate data inferred from this riparian association and on tree-ring growth analyses suggest that the paleoclimate of the northern Gulf of California was more temperate and wetter than now with winter rainfall dominant. The presence of Cretaceous pollen and microscopic foraminifers in time- constrained Pliocene sediments of Anza-Borrego is significant and contributes to the understanding of the biogeographic history of Anza-Borrego. The stratigraphic occurrence and distribution of the microfossils Proteacidites, Mancicorpus, and Aquilapollenites along Fish Creek Wash, for example, indicates that erosion of their source rock, the Mancos Shale, in the southern part of the Colorado Plateau began about 4.5 Ma, and in the northern part of the plateau at about 3.9 Ma. Thus, the presence of these microfossils testifies that the erosional events on the Colorado Plateau, including the extensive down cutting of the Grand Staircase and Grand Canyon, and the achievement of a through-flowing drainage by the perennial Colorado River to the northern Gulf of California, is a relatively young geomorphologic phenomenon.

A detailed strontium isotope study of the Bouse Formation, lower Colorado River, USA: implications for Colorado River history

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The mechanisms and timing for the integration of the Colorado River are the subject of much controversy, and the origin of the Bouse Formation has important implications for the Colorado River debate. The upper Miocene to Pliocene Bouse Formation, exposed along the lower Colorado River valley, was originally interpreted as an estuarine deposit within a Miocene embayment of the Gulf of California on the basis of paleontology; however, recent isotopic work suggests that the Bouse Formation is of lacustrine origin. Rather than a simple single-lake model, these data suggest a lake-overflow model involving a chain of three separate lake basins for the integration of the lower Colorado River. Past studies have established a broad estimate for the overall geochemical character of the Bouse Formation, with data from a wide variety of sample types and locations. Strontium isotope data from Bouse Formation carbonates yield a 87Sr/86Sr ratio (0.7108) close to the strontium ratio of the present-day Colorado River (0.71075) and well above the late Neogene marine strontium ratio (0.7090). However, these previous studies were undertaken in order to establish general isotopic values for the Bouse Formation, and lack clear age and stratigraphic relationships. To obtain Bouse Formation samples with a discernable relationship to one another, a more detailed and systematic sampling was performed. The Bouse Formation exists at elevations ranging from 280 to 1760 ft. AMSL, from Lake Mohave south to Cibola in three distinct paleo-lake basins. By sampling the basal 10 cm of marl at a range of elevations in each of the three basins, it is possible to establish changes in lake water isotopic chemistry through time. The up-elevation deposition of the basal marl layers tracks the filling of the basins. Comparisons are also made with the Sr isotope values of samples of the Hualapai Limestone from the Grand Wash Cliffs area. The

Hualapai Limestone has a significantly higher 87Sr/86Sr ratio (0.7145), and the transition to the lower strontium values that characterize the Bouse Formation is under investigation. Samples were collected from the base of carbonate outcrops in the Las Vegas Basin which have not previously been analyzed for Sr isotopes. A sample taken from near Frenchman Mountain, Nevada within this Las Vegas Basin shows a Sr isotope ratio similar to that of the Bouse Formation rather than to the nearby Hualapai Limestone. Additionally, the Sr isotope results from the Bouse Formation in the northern-most, or Mohave, basin corroborate the general range of Bouse Formation strontium values established by previous work. These data illuminate lake-overflow and mixing processes as well as the evolution of the lower Colorado River drainage system by utilizing the Sr isotope data from the Bouse carbonate and related lake deposits.

Playas, dune fields, and global aeolian processes: the paradox of the "dust-free" Mojave Desert

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On a global scale, deserts are the dominant source of atmospheric soil dust. As in other desert regions, aeolian landforms are common in California deserts. Surprisingly, the Mojave Desert has unusually low atmospheric dust loading compared to other desert and semi-desert regions of Western North America. Moreover, the dominant source of fine soil dust in the air over the Mojave Desert appears to be Asia. Contributing factors for this seeming paradox are discussed, including the structural geology of the Mojave, its vegetation and land use history, and its position relative to global atmospheric circulation patterns.

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