

Landscape Evolution at an Active Plate Margin

edited and compiled by
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In Memoriam

This volume is dedicated to Bobbe Adams who, along with her late husband Bob, helped found the Mojave Desert Quaternary Research Conference that has evolved into the Desert Research Symposium. Bob and Bobbe's influence continues through the scholarship they established for the best student paper presented at each year's symposium. Bobbe also donated her husband's geologic field notes, maps, and photographs to the Mojave River Valley Museum in Barstow, where they are available for research.

*Front cover:
Hot Creek, view toward the Sierra Nevada crest and detail of hot springs. Reynolds photographs.*

*Back cover:
Convict Lake. David R. Jessey photograph.
Terminal and lateral moraines in the Sierra Nevada. Reynolds photograph.*

*Title page:
1872 Lone Pine earthquake scarp. David R. Jessey photograph.*

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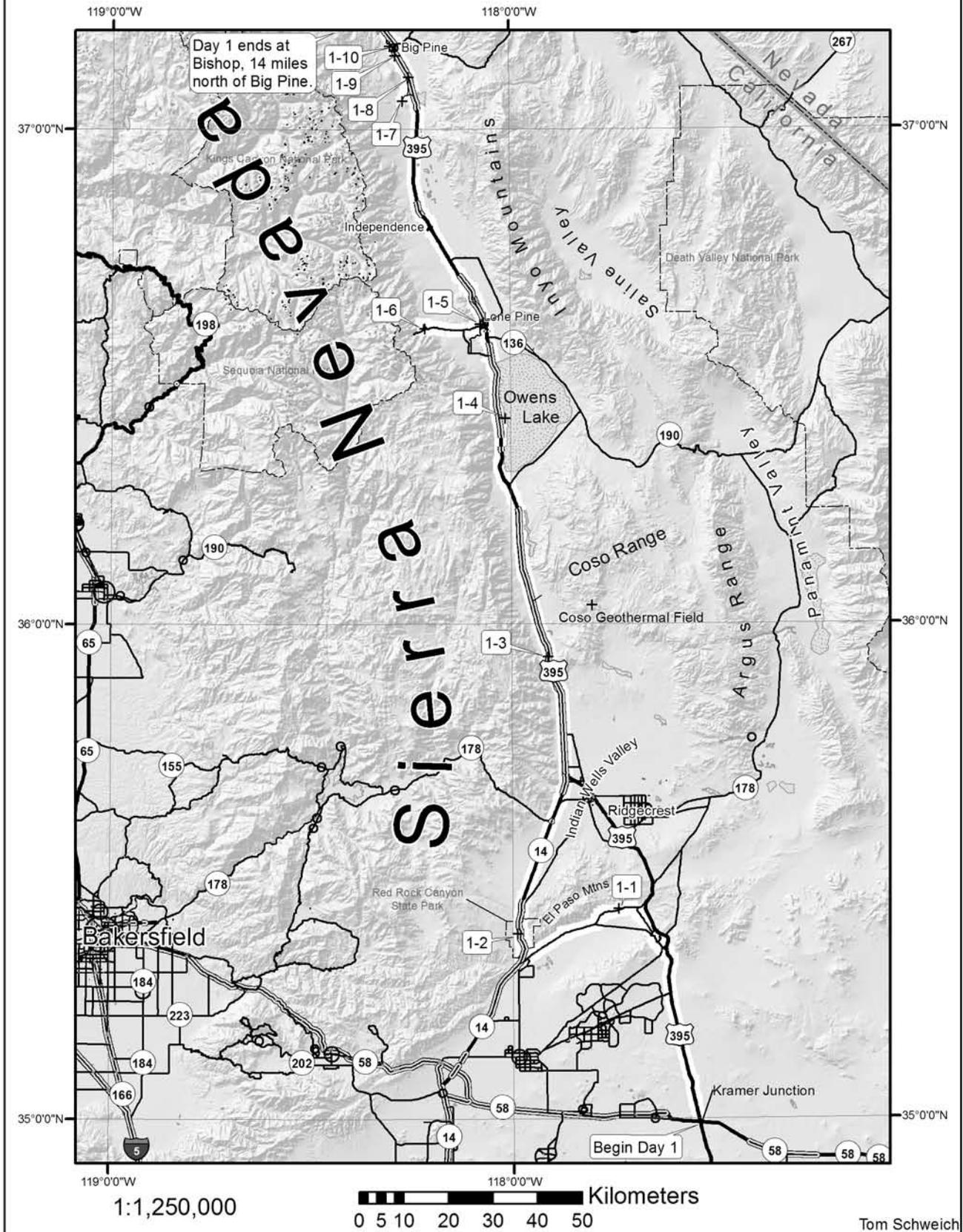
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2009 Desert Symposium Field Trip

Day 1: Mojave Desert and Basin & Range



Landscape evolution at an active plate margin: a field trip to the Owens Valley

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Day 1

Theme: Viscous volcanics and associated tectonic structures

with contributions from D. R. Jessey, R. E. Reynolds, Kim Bishop, Janet Westbrook, and Tom Schweich

Start Day 1

Convene at Zzyzx. Wear sturdy shoes and dress in layers; bring water, hat, and sunscreen. On the night before the trip, fill your vehicle with gas and check fluids and tire pressure.

Drive north from the Desert Study Center to I-15. ENTER I-15 WESTBOUND toward Barstow. Proceed through Barstow. EXIT RIGHT on SR-58 toward Mojave/Bakersfield.

What We Will See: Our 300 mile route will cross the Mojave Desert Province and enter the southwestern Basin and Range Province. In the southern Owens Valley portion of the Basin and Range Province we inspect basalt flows that range in age from middle Miocene to mid Holocene time. Their flow on older surfaces helps describe tectonic activity and local landscape topography. The geomorphic features in Owens Valley are those of an active plate boundary between the North American continental plate to the east and the evolving Sierra microplate to the west.

The Mojave Desert geomorphic province is a wedge-shaped area between the San Andreas and Garlock faults, eastward to southern Nevada and northwestern Arizona. Biological boundaries are not as distinct as those for the fault-bounded geologic province; many Mojave species extend northward into the Basin and Range province. Joshua trees (*Yucca brevifolia*) do not occur naturally outside the Mojave biologic province, and we will pass them as far north as Olancho. Creosote bush (*Larrea tridentata*) is the most common low-elevation shrub in the Mojave; the long-lived plant is closely associated with burro bush (*Ambrosia dumosa*). Bunch grasses are recovering from heavy grazing pressure of the 1850s–1980s. Spring rains can

turn the western Mojave into a solid sea of colorful annuals (yellow desert dandelion, gold coreopsis, blue lupine, and red paintbrush). Look for kestrels, falcons, hawks, and eagles on power line towers (Darlington, 1996; Schoenherr, 1995).

The flat surface of Mojave Desert hides thick, buried fossil soil horizons called “paleosols” (Ponti, 1985) that contain the remains of mastodon, camel, horse, elk, and antelope as well as smaller fossils such as cotton rat and ancestral wood rat. The rodents suggest the middle to late Irvingtonian North American Land Mammal Age (NALMA) of the Pleistocene (0.8–1.5 Ma; Reynolds, 1989, 1991) for paleosols of the Mojave west of Barstow. Late Pleistocene faunas (Rancholabrean NALMA, < 200 ka) are reported from inter-fingering dune–lake sediments around Rogers and Rosamond lakes (Reynolds and Reynolds, 1991). *Bison*, the Rancholabrean NALMA indicator fossil, has recently been reported from the Rosamond Lake area (Buena Vista Museum Natural History collections).

0.0 (0.0) **Kramer Junction.** Intersection of CA Hwy 58 and US Hwy 395. Fill gas tanks. TURN RIGHT (north) on Hwy 395.

1.5 (1.5) View left (west) of FPL Energy. The Kramer Junction Company is the managing general partner for the five 30-megawatt solar electric generating facilities just north of Kramer Junction. The plants normally operate on solar power. However, on cloudy days or early evenings an auxiliary natural gas-fired heater is available to keep the oil warm to supplement the solar power. The Kramer Junction project has a 30 year contract to provide energy to Southern California Edison. The solar generators provide the major portion (over 80%) of their output during periods of peak demand, particularly on hot afternoons.

6.2 (4.7) View left (west) of the Boron Federal Prison Camp, site of the old Boron Air Station. This prison, closed in 2000, was a minimum security camp that housed 540 male inmates. The domed structure at the hilltop remains in use by the Air Force and the FAA.

6.6 (0.4) The “Ball Dome” east of 395 is a doppler radar, EYX.

12.0 (5.4) Continue past a right turn.

14.0 (2.0) Continue past a road leading east to the former siding of Fremont on the Randsburg Railroad, running parallel to Hwy 395. This railroad was active from 1887 to 1933 (Myrick, 1991), serving rich mines in Randsburg and Johannesburg. Because of the 1,200-foot drop into Fremont Valley, a spur to Goler was never attempted. Processing of the ores from mines discovered in the Rand Mountains initially proved difficult due to the lack of water. For this reason Garlock, with good wells, was the first site of ore milling operations. By 1896 there were 4 short-lived mills at Garlock, one at Mesquite Spring, and Charley Koehn’s mill at Kane Spring. After the Randsburg Railroad was completed to Johannesburg in 1898, Yellow Aster ore was processed by a 50-stamp mill in Barstow. A water line from Garlock to Randsburg allowed a 30 stamp mill at the Yellow Aster mine to become operational in 1899. In 1901 the Yellow Aster added 100 stamps to the existing battery.

17.1 (3.1) The Rand Hills are at 10:00; the 1990s dumps of the Yellow Aster mine are at 11:00; and Red Mountain is at 2:00.

18.8 (1.7) Cross 20 Mule Team Parkway. This road also leads from Mojave northeast to Blackwater Well, Granite Well at Pilot Knob, through Panamint Valley and down Wingate Wash to the Furnace Creek borate deposits in Death Valley (Faye, 1999).

21.6 (2.8) Look ahead to excavations for scheelite placer mining for in the Atolia Tungsten District. Scheelite (calcium tungstate, CaWO_4) initially bedeviled dry placer gold miners. They were troubled by fragments of tungsten since it was heavy (specific gravity 6.0) and wound up with the gold concentrate. Scheelite was recognized as valuable in 1903. Claims purchased by De Golia and Atkins, with a contraction of names, became “Atolia.” Ore from lode deposits was initially hand-sorted but a mill was constructed in 1907. The peak tungsten production, nearly \$10 million, was between 1916 to 1918, when Atolia boasted a payroll of \$60,000 per month. The Atolia Tung-Sun Mining Company began placer mining in Baltic Gulch in 1942.

21.8 (0.2) Enter the Atolia Mining District. The headframe of the Union Tungsten Mine can be seen to the northwest (Fig. 1-1). The Paradox claims are on both sides of Hwy 395.



Figure 1-1. Union Tungsten Mine headframe. [J. Westbrook photo]

The district was a major producer of tungsten during the early 20th century. At one point, Atolia had a population in the thousands (Clark, 1997). Interest in tungsten deposits continues to the present but little mining has occurred for the past 40 years. Scheelite, the main tungsten ore, occurs with quartz and carbonates along faults in the Jurassic Atolia Quartz Monzonite (Peck, 1953). The veins strike northwest and dip steeply to both the north and south. They consist of coarse grained scheelite, quartz, and calcite, deposited as open space filling in the fractures. Wall rock alteration has caused pervasive alteration of biotite to chlorite and feldspar to white mica, carbonate, and clays. The district was referred to as the “spud patch” because of the round, light brown nuggets of tungsten.

22.9 (1.1) Enter Atolia.

24.7 (1.8) Continue past a left turn to Randsburg. View northwest of the 1990 Yellow Aster Mine dumps.

25.1 (0.4) Sign: “Entering Red Mountain.” The Kelly silver mine structures are to the west, although the headframe has been dismantled. Initial samples from the mine assayed 300 ounces of silver with three ounces of gold per ton. It became California’s largest silver deposit with over 20 million ounces produced between 1919 and 1942. The ore occurs in quartz-carbonate-chalcedony veins. Silver mineralization consists mainly of Ag-Cu-Sb sulfosalts (miargyrite and pyrargyrite) and Ag chlorides with minor native silver. Arsenopyrite is the chief gangue mineral. Due to the difficulty of milling, native silver was often rejected during processing and large nuggets can still be found on the dumps.

25.3 (0.2) Continue past Osdick Rd.

25.5 (0.2) View west of remaining Kelly Mine control building.

26.5 (1.0) Continue past Trona Road on the right to Searles Lake (see Reynolds, 2002). Steam Wells Valley is to the north. Some of the pumps on the wells that provided Johannesburg and Randsburg with water were operated by steam, an early use of geothermal power (Mendenhall, 1909; Reynolds, et al., 1998).

28.2 (1.7) Enter Johannesburg and Kern County. Randsburg and Johannesburg were the twin metropolises of the Rand Mining District. Randsburg was the company town, while Johannesburg was a planned community for miners with families (Clark, 1997). They functioned as major trans-shipping points for rail and stage lines. Ore from the district was initially shipped south to Barstow for milling, but with the construction of a 100-stamp mill in 1901 the district began to thrive. Large-scale gold mining continued until 1918. Gold production from the district was substantial into the 1930s and early 1940s, and there has been intermittent mining since, most recently from 1989 to 1995 at the Yellow Aster. Total estimated production from Randsburg mines up to 1940 is estimated at \$40 million. The mine was shut down in 2004 and the dumps and heap leach piles have been “reclaimed.”

The principal rocks underlying the Rand district are Precambrian Rand Schist and Jurassic Atolia Quartz Monzonite. The Rand Schist is chiefly biotite schist with smaller amounts of amphibolite and quartzite. Most of the lode-gold deposits are in veins along faults, except at the Yellow Aster mine where the gold occurs in a series of closely spaced veinlets in areas of intense shearing. The veins do not have a well defined strike; high grade ore zones most commonly occur at or near vein intersections or in sheared and brecciated zones. The ore consists of iron oxide-stained, brecciated and silicified rock containing fine-grained gold with varying amounts of sulfides. The sulfides increase at depth, but the gold values decrease. Mining stopped at a depth of 600 feet where unoxidized sulfides were found in the veins.

29.2 (1.0) Prepare for left turn across oncoming traffic.

29.4 (0.2) **TURN LEFT** (west) to Randsburg via the Redrock–Randsburg Road. Randsburg has a museum and a recently active single stamp mill.

29.9 (0.5) Disposal Site signs. **TURN RIGHT** (north) on Goler Road, and proceed north 4.2 miles to Garlock Road. View right (northeast) of Summit Diggings district. Placer gold deposits to the northeast are called Hard Cash Gulch and Summit Diggings (Reynolds, et al, 1998). Auriferous gravels of Summit Diggings may have been derived from Goler Gulch, three miles to the west. Right lateral restoration of movement on the left lateral Garlock fault would place auriferous gravels at the mouth of Goler Gulch. Farther east, the trace of the Garlock fault on the south shore of Searles Lake at the 2,250 foot shorelines is marked with tufa dated at 50,000 ka. Since the dated shorelines are not cut by the Garlock fault, that date constrains the latest movement on that portion of the fault to greater than 50,000 ka (Smith, 1964). We are entering Fremont Valley.

31.3 (1.4) Pass gravel dumps from hydraulic mining operation.

32.9 (1.6) Pass through low hills of Pleistocene alluvium elevated above axial drainage of Fremont Valley by the south-side up Cantil Valley Fault.

33.4 (0.5) Cross Fremont Wash and Cantil Valley Fault.

33.8 (0.4) Railroad crossing, watch for trains. Prepare to turn left (west).

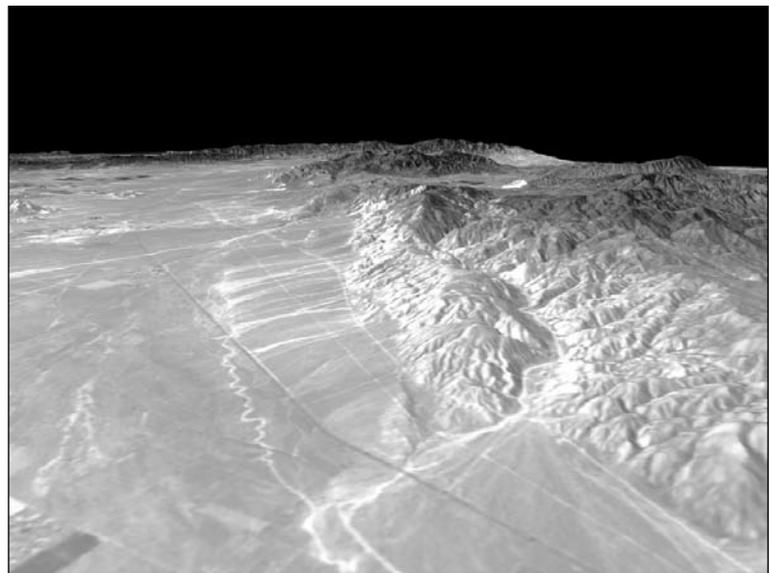


Figure 1-2. Shuttle Radar Topography Mission (SRTM) image looking west along the Garlock fault. El Paso Mountains are in the foreground and Tehachapi Mountains in the background; San Gabriel Mountains on the distant skyline.

34.2 (0.3) **Stop** at Garlock Road. Look for oncoming traffic. **TURN LEFT** (west) onto Garlock Road. The trace of the Garlock fault is to our right (Fig. 1-2). The first mining for lode gold, silver, and copper ore in the El Paso Mountains was in 1863 on Laurel Mountain, east of the El Paso Peaks.

John Goller, survivor of the 1849 Death Valley party, found gold nuggets on his trek across the desert. Although he regularly searched for his lost “mine” placer gold was not rediscovered here until 1893. Miners named one of the prospects, Goler Gulch, after Goller. By 1896, a half-million dollars of gold had been recovered (Vredenburg, 2009; Vredenburg and others, 1981).

35.4 (1.2) Continue past Goler Road on the left. The Goler Formation lies to the north, in the north-central portion of the El Paso Mountains. The Goler Formation consists of a two mile (3 km) thick section of fluvial sediments capped by fossiliferous marine sediments (Lofgren and McKenna, 2002; Cox and Edwards, 1984; Cox and Diggles, 1986; McDougall, 1987). Intense prospecting for fossils by the “Goler Club” has collected a suite of small mammals that represent four NALMA that span the Paleocene (Lofgren, et al., this volume). This is the best section west of Wyoming and Colorado for studying Paleocene faunas (60 Ma). No fossils have been recovered from the lower members (1 & 2) of the Goler Formation, but if rates of deposition can be considered constant, lower strata project as Cretaceous in age (Cox, 1998). Clasts in the Goler Formation tell us that the Mojave Block to the south was elevated in the early Tertiary (Cox, 1998). Prepare for a right turn on Charley Road.

38.0 (2.6) **TURN RIGHT** on Charley Road. Proceed north 0.2 mile.

38.2 (0.2) **PARK** along the road and **WALK NORTH** to the fault scarp.

STOP 1-1 (N 35°25'26.2"; W 117°44'15.5"). Cabins at the base of this 25–30 foot high scarp (Fig. 1-3) are owned by the BLM as part of their Adopt-A-Cabin program.



Figure 1-3. Charley Road (Stop 1-1) View east along Garlock Fault, the skyline notch above road, and playa-filled graben on north side of fault scarp. [J. Westbrook photo]

To the north from this escarpment is the southern flank of the El Paso Mountains. The rocks making up this portion of the El Paso Mountains are generally undifferentiated Paleozoic roof pendants and Mesozoic granitoids. The east–west striking El Paso fault (south side down) lies at the base of the slope and the 160 mile long Garlock fault along the base of this north-facing scarp (Fig.1-2). Look southeast across Cantil Valley to old alluvial ridges that mark the trace of the Cantil Valley fault that parallels the Garlock fault on the south side of Koehn Lake.

The Garlock fault zone marks the geologic province boundary between the Mojave Desert to the south and the Sierra Nevada and Basin and Range to the north. Movement along the Garlock fault is thought to have begun at about 10 Ma (Burbank and Whistler, 1987). While no surface rupture has occurred on the Garlock fault in historic times, there have been a few sizable quakes. The most recent was a magnitude 5.7 near the town of Mojave on July 11, 1992 (SCEDC earthquake database). It is thought to have been triggered by the Landers earthquake, two weeks earlier (Jones, Mori and Hauksson, 1995). Trenching suggests other ruptures occurred in 1050 A.D. near Tehachapi and 1500 A.D. near our present location. At least one section of the fault has shown movement by creep in recent years. Numerous studies have been published citing slip rates along the fault (LaViolette, et. al., 1980; Clark and Lujoie, 1974; Carter, 1980, 1982; Smith, 1975; McGill and Sieh, 1993; Petersen

and Wesnousky, 1994). Reported rates of sinistral slip vary from 2 to 11 mm/yr. The western portion of the fault, near its junction with the San Andreas fault, appears to be the most active with slip rates exceeding 7 mm/yr. In contrast, the eastern terminus of the fault, south of Death Valley, has been characterized by little or no Holocene slip (Smith, 1998). Despite decades of study, the Garlock fault remains an enigma and puzzling questions remain:

- How is strain partitioned across the Garlock fault from the Eastern California Shear Zone into the Mojave?
- Why does the eastern segment of the fault appear to be aseismic, and why does it terminate in a thrust fault at the Avawatz Mountains?
- What is the relationship of the Garlock fault to the San Andreas fault and the larger tectonic picture of southern California?

Return to paved Garlock Road.

39.0 (0.2) **Stop** at pavement, watch for cross-traffic and **TURN RIGHT**. As we proceed west, note the depression on the north side of the road is a left-laterally offset portion of Goler Gulch, the next drainage west (Reynolds and others, 1998). The cement foundation of an ore mill remains on the north scarp the depression. One-half mile west, tamarisk trees mark the site of Cow Wells, where Garlock found water at 30 feet, and where the Yellow Aster mill was built in 1898. The Garlock - Cow Wells area was not reached by the Trona Railway until 1914, when the line was extended from Mojave to Searles Lake and Owenyo to meet the



Figure 1-4. The 1898 Yellow Aster mill site at Cow Wells, where the water table was reached at 30 feet. [J. Westbrook photo]

Carson and Colorado (C & C) Railroad in 1907–1910 (Myrick, 1991). View west shows El Paso fault on right, Garlock fault left, and terrace between.

42.9 (2.9) The town of Garlock (el. 2175 ft). The town was a water source for cattlemen and traders avoiding a trek through the El Paso Mountains. When gold was discovered in a nearby canyon in 1887, an arrastra was constructed to process the ore. In 1893 a nugget valued at approximately \$2000 from Goler Canyon caused a gold rush and in 1895, Eugene Garlock from Tehachapi constructed an eight stamp mill (Hensher, 1998a). Miners would talk of going down to the “Garlock mill” and finally the town name Garlock just stuck. By 1898 the population of Garlock reached over 200 with five saloons, three boardinghouses, two restaurants, a post office, and a school. Seven mills were in operation. However, the town suffered a serious economic blow with the opening of a processing mill in Barstow and two large mills in Randsburg near the principle producing mine, the Yellow Aster (Fig. 1-4). By 1904, Garlock was largely abandoned. Although there was a renewal of activity with construction of rail lines from Mojave to Summit to Owens Valley, the town never regained its former status.

44.0 (1.1) Red Rock–Randsburg Road joins from the left. Continue west toward SR 14. Scarps of the Garlock fault are on the right. Green vegetation marks springs along the fault.

49.9 (5.9) Continue past a left (south) turnoff to Saltdale (El: 2000 ft.). The Consolidated Salt Company constructed a crushing and screening plant and narrow gauge railroad track onto the playa of Koehn Lake in 1914 on sixty placer claims filed by Thomas Thorkildsen and Thomas H. Rosenberger (Hensher, Vredenburg and Wilkerson, 1998). The plant produced around 20,000 tons of salt annually. The company pumped well water onto the lake floor where it was allowed to partially evaporate. The resulting brine was drained into several ponds where a 6-inch layer of salt would form after a few months. The salt was saw cut into cakes that were cleaned by hand and hauled to the mill where they were ground, sized, sacked, and shipped to Los Angeles. By 1918 a second company was producing salt and

in 1922 the companies were merged. Production continued until 1972 when a lawsuit pursued by the BLM found the original claims to be invalid.

45.3 (0.4) Continue past the right (north) turnoff to Last Chance Canyon. Once the main access into the El Paso Mountains, this road now requires 4WD since flash floods have re-engineered the road.

47.7 (2.4) The site of Gypsite (south), operated by Charles Koehn in 1909, produced gypsite (gypsum and clay) used for plaster and agricultural soil amendments (Hensher, 1998).

49.1 (1.4) Cross the channel of Red Rock Canyon Wash.

49.6 (0.5) View of Honda test facility (see below).

49.9 (0.3) Continue past Cantil Road to the left.

50.7 (0.8) Cross a drainage.

53.0 (2.3) Neuralia Road is on the left just before we turn off on CA 14.

Stop at the intersection with CA 14. Look to the south to see a curving line of vegetation. This marks the edge of a 3-mile-wide oval (clearly seen in Google Earth imagery) test facility owned by Honda Motor Corporation. The facility is well guarded and off limits to travelers. At night, car headlights can often be seen as they circle the track. This has fueled conspiracy theories that the test track is the locus of “alien” activity from the nearby “vortex” in the El Paso Mountains. **TURN RIGHT** (north) toward Red Rock Canyon State Park.

55.4 (2.4) A trace of the El Paso fault is visible ahead to the left and right. The tan sediments to the west contain a fossil fauna that is significantly younger (late Hemphillian) than faunas in the Dove Spring formation. The sediments are bounded by the El Paso fault (north), the Sierra front fault (west) and the Garlock fault (southeast) and higher degree of clockwise rotation (35 degrees) than the Dove Spring Formation (15 degrees; Whistler, 1991, p. 111). This suggests that the block has a history independent from the Dove Spring Formation, and may have been left-laterally translated from the Bedrock Spring Formation in the Lava Mountains (east of Hwy 395); a distance of approximately 30 miles in the last 7–5 Ma.

56.1 (0.7) Cross the trace of the El Paso fault, which here juxtaposes Mesozoic (Jurassic?) rocks and undifferentiated Ricardo Miocene Group on its down-

thrown side (Whistler, 2005). Although motion along this fault has been characterized as predominantly dip-slip, its proximity to the Garlock fault suggests it may have a left-slip component. Despite paleoseismic studies on the nearby Garlock fault (McGill, 1992), little is known about slip rates for the El Paso fault. The SCEDC website suggests last movement in the Late Quaternary, but this presents a structural conundrum. The Garlock fault near this location has moved within the past 500 years, suggesting it is quite active. As such, its close proximity to the south-dipping El Paso fault should result in the Garlock fault buttressing dip slip movement along the El Paso fault. Perhaps Quaternary movement along the El Paso fault has been largely left-slip?

56.8 (0.7) Basement rocks (largely granite) are unconformably overlain by the Miocene Ricardo Group. This is best seen driving south on CA 14; when driving north, look back at the unconformity after passing a ridge of Mesozoic basement rock.

57.4 (0.6) **Move** to left lane and prepare for a left turn. Watch for oncoming traffic.

57.6 (0.2) **TURN LEFT** (west) on Abbott Road to the Ricardo Campground and Red Rock Park headquarters.

58.5 (0.9) Note the basalt outcrop on the left side of the road.

58.8 (0.3) Red Rock Canyon visitors center (El: 2175 ft.). **PARK** on pavement before kiosk.

STOP 1-2 (N 35°22'23.3"; W 117°59'18.3"). This outcrop (Fig. 1-5) lies within the west-dipping Dove Spring Formation of the Miocene Ricardo Group. Loomis (1984) divided the Dove Spring Formation



Figure 1-5. Stop 2: outcrop of Miocene Dove Spring basalt exposed along Abbott Road. [D. R. Jessey photo]

into six units which range in age from 7 to 13 Ma. The Dove Spring Formation lies unconformably on the Cudahy Camp Formation. The Cudahy Camp is thought to have been deposited from 15-19 Ma (Loomis and Burbank, 1988). The base of the Dove Spring is marked by conglomerate grading upward into arkose. Overlying layers are comprised of a variety of differing lithologies including lacustrine and fluvial sandstones, siltstones and conglomerates, felsic tuffs and basalt flows. The uppermost unit (Unit 6 of Loomis, 1984) is a nearly flat-lying sedimentary sequence that lies disconformably on the units 1-5 of the Dove Spring.

Recent ages published by Frankel, et al., (2008) indicate the three lowermost basalts were extruded between 11.7 and 10.5 Ma. No age is available for the uppermost basalt flow, but Whistler (1987) quotes an age of 8.1 Ma for a rhyolite tuff above the basalt. This brackets basalt extrusion between about 8-12 Ma. Loomis and Burbank (1988) examined the evolution of the El Paso basin and concluded that extrusion of the basalts was coincident with the onset of left-slip motion along the Garlock fault and the east-west extension characteristic of Basin and Range topography. The subsequent emergence of the Sierra Nevada Mountains began at 8 Ma. Frankel (Frankel et al. 2008) argues that the Summit Mountains south of the Garlock fault are correlative with the Dove Spring Formation, but paradoxically they contain no basalt flows. This suggests a very local source for the basalts. Unfortunately, evidence for nearby faults of the proper age to provide a conduit for the basalt flows is absent. (Fig. 1-6)

Anderson (2004) sampled four basalt flows within the Dove Spring Formation (Tba₁-Tba₃ of Loomis (1984) and a fourth undifferentiated flow within unit 4 of Loomis). Figure 1-6a is a total alkalis vs. silica diagram for the basalts of the Dove Spring Formation. Basalts from the Coso volcanic field (Stop 3) are shown for comparison. Note that Dove Spring (DS) basalts average approximately 53% total SiO₂ and 3% total alkalis. This is 3-5% more silica than basalts from other Owens Valley and Mojave basalt fields and 1-3% more alkalis. Figure 1-6b is a basalt tetrahedron, again comparing the DS and Coso basalts. DS basalts are generally tholeiites while those from Coso are predominantly alkali basalts. The latter are far more characteristic of southern California basalt fields. While some fields; i.e., Big Pine and Darwin, span the entire range of basalt compositions, most like Coso, Cima, Pisgah and Amboy are distinctly alkaline. Tholeiitic basalts are uncommon.

Anderson (2004) also noted the presence of iddingsite in many DS basalts. Iddingsite, a cryptocrystalline mixture of hydrated iron oxides and Mg-clays rims, veins and in many cases totally replaces olivine. In other cases, olivine has altered to a calcium-rich siderite. Unaltered olivine is extremely rare, as the CIPW norm depicted in Figure 1-6b would suggest. Iddingsite is a controversial mineral/alteration product originally thought to form by weathering of olivine. However, Edwards (1938) suggested that iddingsite was more likely a deuteric alteration product formed during magma mixing. Baker and Haggerty (1967) concluded the formation of iddingsite required only an increased oxygen fugacity and could be achieved simply by allowing a basaltic magma to incorporate circulating meteoric water. Furgal and McMillan (2001) reached a similar conclusion, but cautioned the alteration required elevated temperatures, not typical of circulating groundwater.

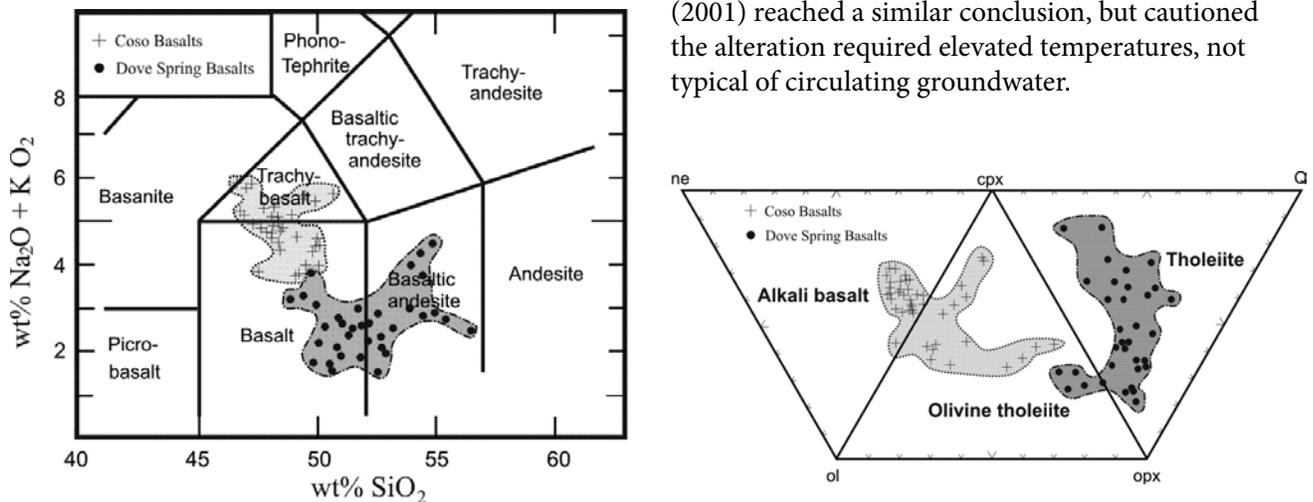


Figure 1-6. (a) Total alkalis vs. silica (TAS) diagram for the Coso and Ricardo (Dove Spring) volcanic fields; (b) basalt tetrahedron for the two fields [modified from Anderson, 2004].

Kushiro (1968) was the first to offer an explanation for differences in composition of mantle-derived basaltic magma. He suggested that depth of melting was an important variable and that magmas from the shallow mantle would tend to be quartz normative tholeiites while magmas produced in the deep mantle would be olivine normative and alkaline. Anderson and Jessey (2005) suggested that one possible explanation for the DS tholeiites, therefore, was that they represented a shallower partial melt. This could be related to their age and/or geographic position to the west of other Mojave/Owens Valley basalt fields and nearer shallow remnants of the east-dipping subduction zone and East Pacific Rise (Yang, 2002). However, Brown, Bruns and Jessey (2008) pointed out that this does not account for the presence of iddingsite which requires preexisting phenocrysts of olivine. Therefore, initially, DS basaltic magmas may have been as alkaline as those of other Owens Valley fields and melting depth might not be an important variable in compositional difference.

The Cudahy Camp and the Dove Spring formations are biostratigraphically important because they contain North American Land Mammal Age (NALMA) transitions of middle and late Miocene faunas. The late early Miocene Cudahy Camp Formation is non-fossiliferous, while the Dove Spring faunas show the transition from the Clarendonian through Hemphillian land mammal age (13.4–7.3 Ma; Burbank and Whistler, 1987; Loomis and Burbank, 1988; Whistler, 1991). Research in the local stratigraphic sequence presents biostratigraphic, magnetostratigraphic and tephrochronologic age parameters (Whistler and Burbank, 1992) that can be compared to other sedimentary sections across the continent (Tedford and others, 2004). This stratigraphic section contains lithologic clasts that relate it to nearby tectonic events. Previous work suggested that clasts from the rising Sierra Nevada Mountains were present in an upper portion of the section that dates between 8.5–8.4 Ma (Whistler and Burbank, 1992). Current studies (D. P. Whistler, p. c. to Reynolds, 2008) suggest that “*Sierran-looking granitic clasts in place*” have been found in a portion of the section above Ash 15 along the Powerline Road west of the campground that dates to 9.4 m.y. associated with the Dove Spring Local Fauna. This suggests that granites of the Sierra Nevada geologic province were exposed by 9 Ma.

Return to vehicles and **RETRACE EAST** to CA 14.

59.8 (1.0) Watch for cross traffic. **TURN LEFT** (north) on CA 14. Note the brightly colored outcrops of west-northwest dipping Ricardo Group lacustrine and fluvial sedimentary layers as you drive north on CA 14. The prominent ridge-capping black rocks are basalt flows within the Ricardo Group.

63.3 (3.5) Continue past a left turn to Dove Spring offroad vehicle area. Pleistocene terrace deposits crop out along Dove Spring Wash. These sediments represent deposition of a bajada between the Sierra Nevada Range and the El Paso Mountains, and were dissected when erosion breached the fault complex along the south side of the El Paso Mountains. Sediments contain small and large mammal species including *Bison*, the RanchoLabrean indicator mammal, and have produced C-14 dates between 19–10 ka (Whistler, 1990, 1991).

64.4 (1.1) Continue past a right (southeast) turn to Burro Schmidt’s Tunnel and northern access to the El Paso Mountains. Continue northeast on CA 14.

65.5 (1.1) Divide (el. 3300 ft.) separates Fremont Valley drainage system from Indian Wells Valley, the latter draining east toward Searles Valley. The Scodie Mountains of the southern Sierra Nevada are to the west-northwest. Indian Wells Valley lies at an elevation of 2300+ feet, only 300 feet above Fremont Valley. Presumably the elevation difference results from movement along the El Paso fault, but it could also be related to transpression caused by a left step in the Garlock fault (Smith, 1998).

73.4 (7.8) Pass Robbers Roost granitic spires at 10:00 (Reynolds, 2002).

77.3 (3.9) Continue past a left (west) turn on SR 178 to Walker Pass and Bakersfield. Walker Pass, at 5250 ft, is the obvious break in the Sierra. Joseph R Walker “discovered” this pass in 1834 leading the Bonneville expedition, the first wagon train to CA; it was later used by Frémont in 1845. The pass became a major route from gold fields at Keyville and Kernville to Mojave and north to Reno (Pracchia, 1994).

77.7 (0.4) Continue past Bowman Road on the right (El: 3700 ft.). Look uphill at 9:00 to see creosote rings. These rings are clones of the original plant, and the largest rings is 60 ft. in diameter and the plant is estimated to be at least 9000 years old.

80.2 (2.5) Continue past a right (east) turn on SR 178 to Inyokern and Ridgecrest (El: 2310 ft.). Magnolia, siding 16 (Inyokern) and Brown, siding 18, on the

Nevada & California railroad were used in building the aqueduct. The town of Ridgecrest grew around the China Lake Naval Ordnance Test Station.

82.2 (2.0) Continue past a left turn to Indian Wells, a spring on the Sierra Nevada frontal fault (SNFF) that provides water for the Indian Wells Brewery and Lodge. Historically, rescuers of the Bennett/Arcane party stopped for water going to and coming from Death Valley. The wells serviced trade route travelers thereafter. Owens Peak (8453 ft.) to the northwest was named for Richard Owens, a member of Fremont's 3rd expedition.

84.5 (2.3) Junction of CA 14 and Hwy 395. Continue north on 395.

90.0 (5.5) Continue past Brown Road. View right (east) of China Lake and at 1:30 to the White Hills, both of which contain late Pleistocene faunas (Day 3, MP 138.8).

91.4 (1.4) Pearsonville on the right (east) is billed as the "hubcap capital of the world" since Lucy Pearson has a collection of hubcaps rumored to number in the tens of thousands. Cross the Inyo/Kern County Line at Pearsonville.

92.6 (1.2) Continue past Sterling Road.

94.5 (1.9) Continue past a left (west) turn to Kennedy Meadows via Nine Mile Canyon Road. The Coso volcanic field is to the east.

The Owens River has cut a canyon between the Coso Range (east) and the Sierra Nevada (west). The oldest volcanic flows lie to the south. The Upper (prominent columnar jointing) and Lower Little Lake Basalt are 130,000 and 400,000 years old respectively. The Basalt of Red Hill, named for the prominent cinder cone just to the east of Hwy 395 at the Fossil Falls turnoff, overlies the Little Lake Basalts. It has not been dated, but is younger than 130,000 years and older than 10,000 years. The flow has followed the drainage of the Owens River and appears at first glance to lie below the older Little Lake Basalt. The most recent volcanic activity at Red Hill is dated at about 10,000 years although there is evidence for a more recent flank eruption from a small vent on the northwest flank of the cinder cone.

100.7 (6.2) **Move left** and prepare for left turn. Watch for oncoming traffic.

100.9 (0.2) **TURN LEFT** (cautiously) onto the Little Lake Road and bear north. The Southern Pacific



Figure 1-7. Columnar-jointed Upper Little Lake Basalt at Stop 3. [D. R. Jessey photo]

Railroad (SP) is on the west. The SP Lone Pine branch commenced in 1908 to facilitate building of the LA Aqueduct. This is the site of the Little Lake Hotel and gas station. The only building left is the green post office to the west.

101.1 (0.2) **PARK** and **WALK** to the outcrop of columnar basalt.

STOP 1-3 (N 35°56'00.2"; W 117°54'29.9"). The Coso volcanic field lies to the east of the Sierra Nevada at the western edge of the Basin and Range province. It consists of Pliocene to Quaternary rhyolite domes and basaltic cinder cones covering 150 square miles. The youngest eruptions are bimodal, with basaltic lava flows intruded by 38 rhyolite lava flows and domes.

The rocks exposed in this outcrop are columnar-jointed basalts of the Upper Little Lake Basalt (Fig. 1-7). In general, the basalt flows and cones on the west side of the Coso volcanic field are Pleistocene in age while those to the east are Pliocene. The volcanism occurred in two distinct pulses; one from 4.0-2.5 Ma, that resulted in 31 km³ of basalt, rhyodacite, dacite, andesite and rhyolite, and a second from 1.1 Ma to the Holocene with nearly equal amounts of basalt and rhyolite (Duffield, et. al., 1980). These basalts contain fresh, unaltered phenocrysts of olivine in a fine-grained groundmass of plagioclase and both clinopyroxene and orthopyroxene. No iddingsite has been observed or reported by previous researchers (Grove, 1996, Bacon, et. al., 1981, DePaolo, et. al., 2000). Figure 6a reveals that Coso basalts are generally alkaline with K₂O+Na₂O averaging 5%. The majority of samples plot in the alkali basalt triangle of the basalt tetrahedron (Fig. 1-6b).

The structural setting of the Coso volcanic field remains controversial. Monastero, et. al. (2005) conclude the Coso Range lies at a right step or releasing

bend in a dextral shear system that extends from the Indian Wells Valley northward into the Owens Valley. This results in northwest-directed transtension, which is accommodated by normal and strike-slip faulting. This necessitates that the Owens Valley fault must extend southward from Diaz Lake in the Alabama Hills, beneath the sediments of Owens Lake, to the eastern side of the Coso Range where it steps west to the Little Lake fault. If so, the tectonic setting would be analogous to southern Death Valley, where basalts have been emplaced along the Black Mountain detachment formed by a right step in the Furnace Creek-Southern Death Valley fault zone.

Return to the vehicles and Hwy 395, bearing left at the cattle guard.

101.3 (0.2) **Stop** at pavement, watch for cross traffic. **TURN LEFT** (north) on Hwy 395.

101.5 (0.2) Little Lake on the right is a spring-fed lake augmented by a low dam.

101.9 (0.4) Continue past the turnoff to Little Lake Ranch (private duck hunting club) to the right. Red Hill (the scoria cone just north of the turnoff) is straight ahead. Red Hill is one of the youngest volcanic features in the Coso volcanic field. It is thought to be no older than 10,000 B.P. and may have last experienced a flank eruption within the last 1000 years. The flank eruption is marked by a small crater that lies on the northwest side of the cinder cone and is best seen traveling south on Hwy 395. CEMEX currently operates the Red Hill Quarry on the south flank of the cinder cone. They have a small crusher plant that produces a fine scoria aggregate used for road cinders and in the manufacture of cinder block, as well as larger-sized material utilized for landscaping.

104.6 (2.7) Continue past a right (east) turn at Cinder Road to Fossil Falls BLM campground. Continue north on Hwy 395. The Fossil Falls campground and interpretive nature walk lie a short distance along this road. Fossil Falls gets its name from the basalt outcrop that marked an active waterfall or rapids in the Owens River during times of a wetter climate. The last major overflow of water from the Owens River into China and Searles Lakes probably occurred about 4,000 years ago (Bates, this volume). The potholes in the basalt attest to the water's velocity. The basalt here is typical of the Little Lake Basalt, with prominent olivine phenocrysts.

109.4 (4.8) Coso Junction Rest Stop. Trees and greenery to the west at 9:00 mark the trace of the SNFF.

Three lines to the west are: upper, old LADWP aqueduct (1913); center, Southern Pacific railroad to Lone Pine; lower, new LADWP aqueduct (1967).

112.3 (2.9) Northeast at 1:30 you can see an ancient landslide that blocked the drainage of the Owens River. The river subsequently cut through the toe of the slide and resumed its normal channel.

112.4 (0.1) Dunmovin.

112.9 (0.5) Continue past a right turn to the Haiwee Power Plant.

116.2 (3.3) Continue past a right turn on Haiwee Road leading to the reservoir. Haiwee Reservoir, to the east, marks the southernmost "lake" in the LADWP system for the eastern Sierra. Two intakes, one constructed in 1913 and the second in 1970, draw water from the reservoir. From this point southward to Los Angeles, the Owens River flows only in aqueducts. The high peak on the Sierra crest at 10:00 is Olancho Peak (el. 12,122 ft).

117.1 (0.9) Haiwee Canyon Road is on the left.

119.2 (2.1) Continue past Lakeview Road. Vegetation marks the SNFF scarp (west).

120.9 (1.7) Sage Flat Road junction. Continue on Hwy 395.

122.1 (1.2) North Haiwee Road crosses spurs of the S. P. railroad that served during construction of the LADWP aqueduct.

122.8 (0.7) Cross the LADWP aqueduct.

124.1 (1.3) Continue past a right turn to Cactus Flat.

124.5 (0.4) Continue past a left turn to Walker Creek Road and a wild burro refuge.

125.5 (1.0) The historical marker commemorates the Farley Olancho mill site. Look east to the Plio-Pleistocene basalt of the Darwin volcanic field that drapes the south end of the Inyo Mountains. The basalt covers Paleozoic sedimentary rocks and metamorphosed Mesozoic volcanics and sediments.

126.0 (0.5) Continue past the junction with CA 190 and continue north on Hwy 395 toward Lone Pine. CA 190 heads east across the Darwin Plateau and Panamint Valley before descending into Death Valley.

127.5 (1.5) The Crystal Geyser bottled water plant is on the right. This plant began producing bottled water in 1990. Since then, two additional plants have

been built to meet the growing demand. Their website proclaims “Our water really is from a spring. CRYSTAL GEYSER® ALPINE SPRING WATER™ actually locates the perfect spring source first, and then builds our bottling plant there.” A mystery surrounds the actual location of this “spring.” Crystal Geyser publicity suggests the water is from a “protected source” below the summit of Olancho Peak. A visit to the bottling plant failed to resolve this issue as employees would not divulge the location. A report in the High Country News (Nov. 1996) states that Crystal Geyser has been involved in a dispute with Anheuser-Busch over the groundwater beneath their properties in the Owens Valley. They believed that increased pumping of groundwater by Anheuser-Busch immediately north might allow saline water into local aquifers. The Crystal Geyser water comes from a well inside the blue building - not exactly from the “pure mountain spring” their advertising suggests. Inyo Co. regulations only allow Busch to export water by truck-load. Most goes as drinking water to the Briggs Mine in Panamint Valley.

128.5 (1.0) Point of Historical Interest commemorating town of Cartago (see Day 3, MP 31.7), a shipping point for silver ingots from Cerro Gordo south to Los Angeles. Springs at the base of the Sierras mark the frontal fault.

130.3 (1.8) View northeast of the Permanente Soda Works on Owens Lake (See “soda” Day 3, MP 36.6).

134.8 (4.5) The Los Angeles Aqueduct Cottonwood Treatment Plant is west of the highway. At this location, ferric chloride is injected into the water and precipitates ferric arsenide. The resultant heavy precipitant is deposited in the inlet channel to Haiwee Reservoir. The primary source of arsenic is thought to be hot springs at Hot Creek (Stop 2-5, this volume) near Mammoth Lakes (Campbell, 2007).

135.3 (0.5) **TURN RIGHT** (east) to the 1873 Cottonwood charcoal kilns.

136.3 (1.0) **PARK** at the Cottonwood charcoal kilns (Fig. 1-8).

STOP 1-4. Colonel Stevens built a sawmill in upper Cottonwood Canyon to saw wood for the kilns. Timber was delivered to the kilns by a Y-shaped flume and “roasted”

with minimum oxygen to produce charcoal, then carried by paddle-wheel steamship across Owens Lake to Swansea (see Day 3, Stop 3-8 MP 80.1). The Southern Pacific Mojave to Owens Valley railroad grade 500 feet to the east is marked by a culvert under the railroad grade. Retrace to CA 395.

137.3 (1.0) **Stop** at 395, watch for cross traffic, and **TURN RIGHT** (north).

139.1 (1.8) Continue past a left turn (west) to the Cottonwood Power Plant, built in 1910, which supplied hydroelectric power to shovels that dug the aqueduct channel from Aberdeen (Big Pine) to Haiwee. Two original 1906 Pelton water wheels at the plant are still working. Later, the plant supplied electricity to tungsten mills at Red Mountain and Atolia. Cottonwood Power Plant produces about 6,000 megawatt hours of power per year, enough for 5,000 homes. The average flow is 16 cfs, about 11,000 acre-feet per year. LADWP has similar plants at Division Creek and Big Pine Creek that take Sierra water to the aqueduct. Between the DWP and SCE, almost every creek with fairly constant flow from the mountains is hooked to a hydroelectric plant.

139.9 (0.8) View right (northeast) at 1:00 of the abandoned potash works associated with the Pittsburgh Plate Glass plant.

141.1 (1.2) Pass the abandoned Pittsburgh Plate Glass sodium carbonate plant. Sodium carbonate was extracted from “dry” Owens Lake. PPG used the chemical to prepare specialty glass. The property was subsequently purchased by a doctor who made his fortune by inventing a heart valve. He died before realizing his intention of turning the two-story office building into a museum highlighting his invention. View northeast of recessional shorelines. Lone Pine Peak (12,944 ft.) dominates the Sierra skyline.



Figure 1-8. Stop 1-4. The Cottonwood charcoal kilns. [Reynolds photo]

144.1 (3.0) View right (northeast) of recessional shorelines. Historically the lake was 30 to 50 feet deep, deeper in the Pleistocene.

145.7 (1.6) Continue past a left (west) turn to Lubken Canyon Road.

147.0 (1.3) Left (west) is Diaz Lake, a sag pond created between the Lone Pine and Owens Valley faults during the 1872 earthquake. The sag pond was empty in 1955 and did not hold water until 1969. The name Diaz comes from Eleuterio and Rafael Diaz, who were the unfortunate owners of the ranch that became the site of the lake following the earthquake (Clark, 1997). The Alabama Hills west of Lake Diaz form a prominent, 10 mile long, north-trending ridge of Cretaceous granite and Jurassic metavolcanic and metasedimentary rocks separated from the main Sierra massif by a frontal fault (the Independence fault). As such, they are only one of two areas of arc-related rocks within the valley. The other arc-related rocks are the Poverty Hills, 35 miles to the north.

148.3 (1.3) Continue on Hwy 395 past a junction on the right with Hwy 136 to Death Valley.

149.6 (1.3) Enter Lone Pine.

149.8 (0.2) A movie museum is on the left (west).

150.1 (0.3) **TURN LEFT** (west) at the traffic light onto Whitney Portal Road.

150.5 (0.4) Intersection of Tuttle Creek and Fairbanks Road. Portagee Joe Creek campground is 0.1 mile south (left).

150.7 (0.2) **TURN RIGHT** just beyond the LADWP aqueduct.

150.9 (0.2) Cross two tree-lined creeks, immediately **TURN RIGHT**, and **BEAR LEFT** (north) on a dirt road.

151.1 (0.2) **PARK** and **WALK** to the fault scarp (Fig. 1-9).

STOP 1-5 (N 36°36'19.2"; W 118°04'28.9"). White granite debris against dark rock debris. This scarp, on the Lone Pine fault, preserves evidence of slip during the Lone Pine earthquake of March 26, 1872. The Owens Valley fault generated a magnitude 7.5 to 7.8 earthquake considered to be the third largest historic earthquake in the conterminous United States (Ellsworth, 1990; Beanland and Clark, 1994). This rupture had both oblique-normal and right slip displacements. Dextral displacement averaged 6.0



Figure 1-9. View looking to the north along the Lone Pine fault. [D. R. Jessey photo]

± 2.0 m with 1.0 ± 0.5 m of normal slip (Bacon and Pezzopane, 2007) Surface rupture length has been measured at 110 km (Beanland and Clark, 1994). This particular scarp has been extensively studied (Oakeshott, et al., 1972, Beanland and Clark, 1987, Lubetkin and Clark, 1988). It lies near the southern extent of the 1872 surface break.

The Owens Valley fault near Lone Pine consists of the main trace and a secondary branch. The “main” fault extends along the western side of the town of Lone Pine forming the east side of Lake Diaz to the south. The discontinuous secondary trace lies approximately 1.0-1.5 kilometers to the west at the base of the Alabama Hills. This secondary fault was termed the Lone Pine fault by Lubetkin (1980). While both faults ruptured during the 1872 earthquake, the Lone Pine fault experienced right-oblique slip and the Owens Valley fault only dextral shear.

Following the 1872 earthquake researchers initially focused on vertical displacement reporting 12–18 feet of offset (Hobbs, 1910). It wasn't until the 1960s that the right-slip component was fully documented and recognized as the dominant component of motion (Bateman, 1961). Offset of roughly 18 feet vertically and 40 feet horizontally was ascribed to the 1872 earthquake.

In the 1980s it was determined that this scarp was, in fact, the product of three separate Holocene events. The dip-slip component from the 1872 earthquake was estimated to be 3–6 feet with horizontal slip of 12–18 feet. Lubetkin and Clark (1988) proposed three

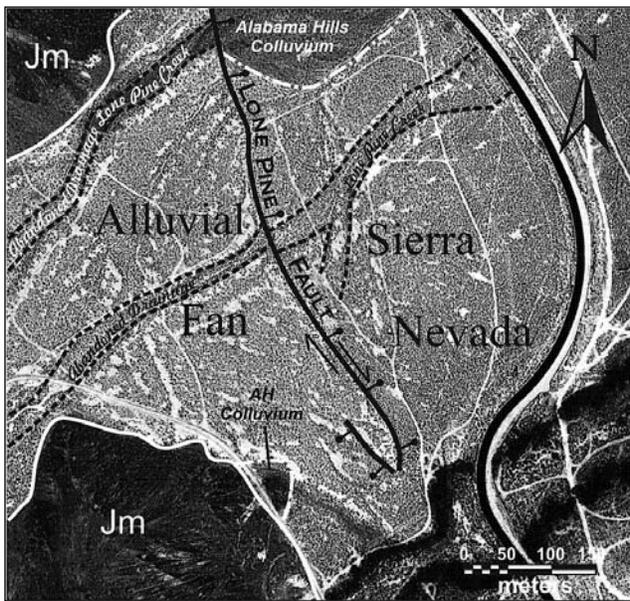


Figure 1-10. Offset drainages of Lone Pine Creek by the Lone Pine fault. Jm = Jurassic metavolcanics of the Alabama Hills [Google Earth, 2008].

separate events; the antepenultimate event between 10,500 and 21,000 cal yr BP, and the penultimate event at 5000 cal yr BP. Subsequently, Bacon and Pezzopane (2007) reported ages of 14,000–24,000 and 8800–10,200 for the two events. Calculated slip rates vary from 0.4 to 1.33 mm/yr horizontally and 0.12–0.25 mm/yr vertically, with a recurrence interval of approximately 10,000 years (Lubetkin and Clark, 1988, Bacon and Pezzopane, 2007).

Figure 1-10 is a recent aerial photo available through Google Earth. It reveals the presence of an abandoned stream channel southeast of the main drainage of Lone Pine Creek that was not previously recognized. This portion of the channel has been offset 70–80 meters to the south along the fault. Assuming a horizontal displacement of 1 mm/yr this channel would represent ~75,000 years of fault motion. Interestingly, the channel also appears to be offset ~25 meters in the direction of dip. The reason for this offset is less clear. If one postulates a 45° dip on the Lone Pine fault, the ratio of horizontal to vertical offset would be similar to that reported by other researchers. However, Lubetkin and Clark (1988) state the Lone Pine fault dips 75–90° and as such the resulting separation cannot be accounted for simply by the dip-slip component of motion. It appears more likely to result from incision of the channel and deposition of debris flows in the abandoned channel. Just when this would have happened is problematic as Lubetkin and Clark (1988) suggest this channel has been abandoned for at least 10,000 to 20,000 years. See also Bishop (this volume)

and Copenhagen (this volume).

Lone Pine Creek supports classic riparian woodland with cottonwood trees, willows, and grasses. Jeffrey pines come from cones washed down Lone Pine Creek from the Portal area.

Return to the Whitney Portal Road.

151.4 (0.3) **TURN RIGHT** (west) on Whitney Portal Road.

154.2 (0.1) Continue past the right turn to Movie Flat Road, with an information plaque about the Alabama Hills as a film and television location. The Alabama Hills are a popular location for television and movie productions; often Westerns. Since the 1920s, over a dozen television shows have filmed here, including *The Gene Autry Show*, *The Lone Ranger*, and even episodes of *I Love Lucy*. Over 250 movies including classics such as *Gunga Din*, *Springfield Rifle*, *A Star is Born*, and *How the West Was Won*, as well as more recent productions such as *Star Trek V* and *Tremors*, were also filmed here. In *Gladiator*, Russell Crowe rides a horse in front of the Alabama Hills, with Mount Whitney in the background, for a scene presumably set in southern Europe (Internet movie database, 2009; Wikipedia, 2008).

151.5 (0.1) Continue past the Alabama Hills visitor turnout on the right. Mt. Whitney, ahead, reaches 14,496 ft.

153.5 (2.0) Pass through Movie Flats.

153.9 (0.4) Continue past a left turn to Horseshoe Meadows.

156.3 (2.4) Continue past Olivas Ranch Road to the left.

157.2 (0.9) Continue past Lone Pine campground to the left (southwest).

159.0 (1.8) **PULL RIGHT** into interpretive overlook on the north side of Whitney Portal Road. (*Caution—watch for downhill traffic!*)

STOP 1-6 (N 36°35'45.5"; W 118°12'46.7"). As we look to the east across the Owens Valley we can see some of the tectonic features shown on Figure 1-11. We are standing on or very near the scarp of the Independence fault. The valley floor lies at an elevation of 4000 feet while the Sierra crest averages 12–14,000 ft. and the Inyo Range averages 9,000, making the Owens Valley one of, if not the “Deepest Valley” in the United States. The Lone Pine fault (previous stop)

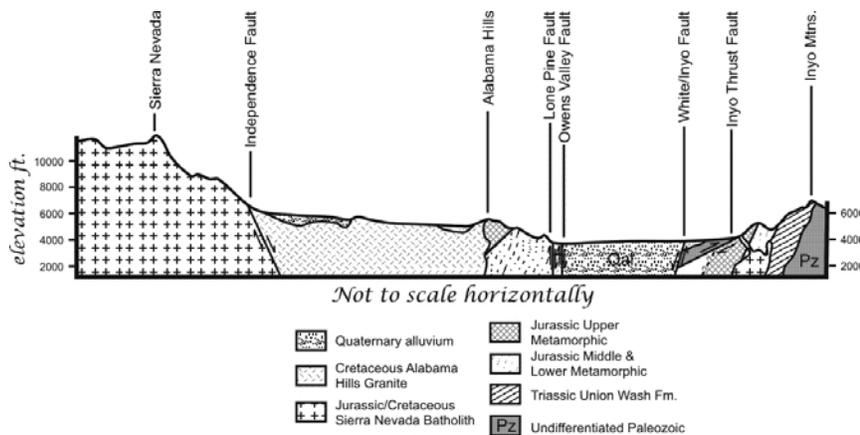


Figure 1-11. Generalized geologic cross section of the central Owens Valley. [D. R. Jessey]

lies along the east edge of the Alabama Hills and the Owens Valley fault about 0.6 miles to the east of that. The scarp of the Inyo Mountains is created by the White-Inyo fault.

The Alabama Hills lie 0.6–1.2 miles west of the town of Lone Pine, California. They form a prominent ridge of Cretaceous plutons and Jurassic metavolcanic and metasedimentary rocks separated from the main Sierra massif by the Independence frontal fault. Across the valley is the “famous” Union Wash ammonite locality and the geologically complex Inyo Mountains. The Owens Valley and Inyo Mountains lie near the western margin of the North American craton and at the center of the central Sierra Nevada segment of the Cordilleran arc (Sorensen et al, 1998). The emplacement of the Sierra Nevada batholith was a result of the oblique subduction of the Farallon oceanic plate beneath the North American plate from 210–80 Ma.

The late Cenozoic tectonic history of eastern California is dominated by the Eastern California Shear Zone (ECSZ). The ECSZ is the name applied to the southern segment of the Walker Lane belt, a zone of dextral shear that accommodates about 20–25% of the total relative motion between the Pacific and North American plates (Gan, et al., 2000). North of the Garlock fault the ECSZ is 60 miles wide and comprised of four separate north-northwest striking fault zones; from east to west the Death Valley-Furnace Creek, Fish Lake Valley, Hunter Mountain–Panamint Valley, and Owens Valley (Frankel, 2008). The Owens Valley fault is the only one of the four that has recorded a major seismic event in the historic past.

Le, et. al. (2007) published a comprehensive study of Sierra Nevada Frontal Fault (SNFF) system including the Independence fault, 5 miles to the west of Lone

Pine. Their study reported that movement along the Independence fault was dip-slip, east side down, at a rate of 0.2–0.3 mm/yr. This rate is about one-half to one-third that necessary to accommodate the 5000–7000 feet of vertical displacement between the Sierra Nevada crest and the Owens Valley. They suggested this discrepancy indicates the Sierra Nevada Mountains are older than the generally agreed upon figure of 3–5 Ma (Wakabayashi and Sawyer, 2001), perhaps as

old as 10 Ma.

The Lone Pine fault lies immediately to the east of the Alabama Hills. The Lone Pine fault appears to be right, oblique-slip fault. Rates of vertical motion are estimated at 0.12–0.25 mm/yr. Horizontal slip rates are less well constrained, but Lubetkin and Clark (1988) suggest rates ~1 mm/yr. The relationship between the Lone Pine fault and the Owens Valley fault to the east is enigmatic. Since both were displaced by the 1872 Lone Pine earthquake they are clearly interrelated, however, Le et. al (2007) suggest that motion along the Owens Valley fault is almost pure dextral shear. Estimates of slip rate are variable from a high of 4.5 mm/yr (Kirby, et. al., 2008) to a low of 0.7 mm/yr (Lubetkin and Clark, 1988). No estimates are available for the vertical component of movement along the Owens Valley fault, which manifests itself in the scarp created by the 1872 earthquake, but it is undoubtedly small; < 0.1 mm/yr.

The White-Inyo fault strikes along the east side of the Owens Valley, 5–7 miles east of Lone Pine. Its history is complex. Bacon et. al. (2005) suggest that Holocene motion along this fault is predominantly right oblique-slip at a rate of 0.1–0.3 mm/yr. However, Beiller and Zoback, (1995) argue that motion along the White Mountains/Inyo Mountains fault system underwent a change from predominantly dip-slip (west side down) to oblique slip at 288 ka. This seems reasonable as the vertical component in an oblique-slip fault moving at a rate of 0.1–0.3 mm/yr would be insufficient to account for the relief between the White/Inyo crest and the valley floor.

Richardson’s (1975) Master’s thesis from the University of Nevada–Reno was the first publication on the stratigraphy of Alabama Hills. The Jurassic sedimen-

tary and metavolcanic units are largely nonmarine arc rocks. Similar rock units are found in the southern Inyo Mountains, where they are underlain by unmetamorphosed to weakly metamorphosed lower Mesozoic to Paleozoic strata. A more recent paper by Dunne and Walker (1993) subdivides the Jurassic Alabama Hills into a lower interval comprised of ash-flow tuff and volcanogenic sedimentary strata in nearly equal amounts and an upper unit consisting of rhyolitic, ash-flow tuff. In the southern Inyo Mountains, the lower interval is described as having a thick basal conglomerate, overlain by volcanogenic conglomerate, sandstone and siltstone, capped by a basaltic lava flow. The middle interval is comprised of about 65% silicic tuff, 25% andesite and rhyolite, and 10% volcanogenic sedimentary units. The upper interval is composed of roughly 95% volcanogenic conglomerate, sandstone, siltstone and about 5% welded and ash flow tuff. The volcanic and sedimentary units were subsequently metamorphosed during emplacement of Mesozoic granites (148 and 82 Ma) (Chen and Moore, 1979).

Despite extensive study some intriguing questions remain unanswered:

- How is strain partitioned across the Owens Valley? This question deals with the observation that both strike-slip and dip-slip faulting have occurred within the past 5 Ma. Stockli et al. (2003) suggest that vertical and horizontal motion are episodic; the current regime of dextral shear began about 3 million years ago. Bellier (1995) suggests that the Owens Valley fault transitioned from dip-slip to dextral slip motion at 0.288 ka and that this transition may not have been the first. In contrast Monastero et al. (2005) argue the Coso Range was created by transtension due to a right step in the Owens Valley fault. Such a tectonic setting could accommodate simultaneous dip-slip and strike-slip motion.
- What is the age of the Owens Valley fault? Most believe that ECSZ has been active only since Pliocene-Pleistocene time (Lee et al., 2001). Based upon measured rates of Holocene slip, displacement on the order of 6 miles would be indicated (Moore and Hopson, 1961). Others, however, believe that the Owens Valley segment of the ECSZ may have been active since 83 Ma with displacements of about 40 miles (Kylander-Clark, et al., 2005). Bartley et al. (2007) even propose that 10–40 miles of additional offset occurred prior to 83 Ma.
- How do we reconcile the slip discrepancy between geodetic measurements and paleoseismicity? Paleoseismic studies suggest slip rates across the Owens Valley fault at 1–3 mm/yr (Beanland and Clark, 1994; Lee et al., 2001; Bacon et al., 2002; Bacon and Pezzopane, 2007). This is at odds with geodetic measurements suggesting motion of 6–8 mm/yr (Peltzer et al., 2001). The discrepancy between satellite measurement and paleoseismic studies is difficult to reconcile. Suggestions include an elastic upper crust overlying a viscoelastic lower crust (Dixon et al., 2003) and a heretofore undiscovered component of oblique slip on many of the dip-slip faults (Bacon and Pezzopane, 2007).
- How is strain partitioned across the Garlock fault? Much controversy exists as to fate of the ECSZ south of the Garlock fault. The Mojave is dominated by dextral shear, but no convincing evidence has been presented of ECSZ faults cross-cutting the Garlock. The Blackwater fault and the Calico fault have been proposed as southern extensions of the Little Lake (Owens Valley) fault (Peltzer et al., 2001), but this is difficult to reconcile with sinistral shear along the Garlock.
- What role does Sevier thrusting play in the evolution of the Owens Valley? Stevens and Olsen (1972) mapped a fault, termed the Inyo thrust, near Tinemaha Reservoir and suggested rocks were carried eastward along this fault a distance of approximately 20 miles. A major problem was timing of the thrust. Stevens originally correlated the thrust with the Permian–Triassic Last Chance thrust, a hypothesis that was untenable for the fault as mapped to the east of Lone Pine. Here, Middle Jurassic rocks lie in the hanging wall of the thrust and hence the fault can be no older than Jurassic. Dunne and Gulliver (1978) even questioned the existence of the thrust, suggesting instead that the fault was a normal fault. Stone et al. (2004) provide some resolution to this controversy suggesting Stevens' Inyo thrust lies near the crest of the Inyo Mountains at Cerro Gordo and that a younger thrust fault exposed along the base of the Inyo Mountains is, in fact, either late Jurassic or early Cretaceous in age. These stacked thrusts suggest a Sevier style of deformation. However, the dearth of outcrops has made definitive conclusions difficult.

We are at elevation 6000 ft. and vegetation has changed (see Gardner, this volume). Ahead at Whitney Portal (8200 ft.) are dense stands of Jeffrey pine and mixed coniferous plants, along with blue grouse, deer, black bear, Steller's jays, and golden-mantled ground squirrels. Downslope, precipitation decreases and temperatures increase, supporting piñon–juniper woodland in turn replaced at lower elevations by Great Basin plant communities of mountain brush associations including bitterbrush (*Purshia* sp.) and Brigham tea (*Ephedra viridi*). Black brush (*Coleogyne ramosissima*) is found on interfluves of the alluvial fans. Lower, Great Basin sagebrush (*Artemisia tridentata*) is dominant. Along the road is a thick stand of rabbit brush (*Chrysothamnus* sp.), a “pioneer” in disturbed areas that crowds out natives. California buckwheat (*Eriogonum fasciculatum*) may be native or introduced. Scattered needlegrasses (*Achnatherum* sp.) and native bunch grasses are making a comeback after grazing.

Return to vehicles, turn around and **RETRACE** to Lone Pine on Whitney Portal Road.

164.2 (5.2) Slow for curves ahead.

167.1 (2.9) **Stop** at the traffic light and **TURN LEFT** (north) on Hwy 395.

167.2 (0.1) Cross Bush Street. The Southern Inyo Museum is on the left; Stratham Hall senior center is on the right.

167.4 (0.2) Continue past Russell Spasshouse Park on the left.

167.9 (0.5) Continue past Narrow Gauge Road on the right leading to site of Owenyo.

168.1 (0.2) The cemetery on the hill to the left (west) side of Hwy 395 contains a mass grave site for victims of the 1872 Lone Pine earthquake. Twenty-seven Lone Pine residents were killed in the quake, but only 16 were buried in this grave. The peak to the east in the Inyo Range is New York Butte (el. 10,668 ft); farther to the north is Keynot Peak (el. 11,101 ft).

169.9 (1.8) First set of northwest trending powerlines.

171.2 (1.3) Cross the Owens Valley fault scarp after the second set of powerlines. The LADWP aqueduct lies behind the low fence at the base of the Alabama Hills. The vast riparian woodland of willows, tules, and grasses is supported by springs along the fault and leaks from the aqueduct.

172.0 (0.8) The Alabama Gates spillway from the LADWP aqueduct.

173.0 (1.0) Continue past a left turn to Moffat Ranch Road, the northern exit from the Alabama Hills.

176.6 (3.6) The Manzanar National Historic Monument Interpretive Center is to the west. Mt. Williamson (el. 14,389 ft) is on the Sierra crest. Manzanar is Spanish for apple orchard. During the early 20th century, this part of the Owens Valley supported a thriving agricultural community that produced some of the finest apples and pears in California. When LADWP acquired the water rights, the supply of agricultural water was reduced and the land was abandoned. Following the attack on Pearl Harbor the Federal government appropriated the land for one of five major Japanese–American Relocation Centers. At its peak Manzanar housed nearly 18,000 men, women and children. WWII ended, the camp was closed, and Inyo County utilized parts of it for storage and maintenance of county equipment. It fell into a state of disrepair until the National Park Service undertook an extensive renovation and reopened Manzanar as an interpretive center in 2004. The former school multipurpose building now houses extensive exhibits documenting the events that lead to the relocation, and detailing the everyday lives of the internees. One guard tower has been reconstructed, original housing units have been returned and renovated, and gardens refurbished.

177.2 (0.6) Continue past a right turn (east) on Manzanar–Reward Road leading to the Union Mining District.

182.1 (4.9) Enter Independence. Continue past a right (east) turn to Kearsarge and Mazourka Canyon (see Day 3).

182.5 (0.4) Market Street. The Inyo County Courthouse is on the right; the Eastern California Museum is 0.25 miles to the west. The original courthouse was destroyed by the 1872 earthquake and a later one by fire. This structure dates from 1921 (Clark, 1997). Note the large block of Bishop Tuff in front of the courthouse.

182.7 (0.2) Deny Park (Inyo County) is on the left (west).

183.8 (1.1) The gravel plant at 9:00 produces aggregate for the CalTrans Hwy 395 improvement project. When widening of the highway from Independence to Bishop is complete, the plant will move south.



Figures 1-12 and 1-13. East of Goodale Road, 1970s diversions allow water to flow into the Owens River channel. [J. Westbrook photos]

184.2 (0.4) Continue past Fort Road South. The devastating mud flow of August 2008 will be visited on Day 3 (Wagner and others, this volume).

184.7 (0.5) Continue past Schabbell Lane.

185.1 (0.4) Continue past Miller Lane.

190.8 (5.7) Continue past Sawmill/Black Rock Road. Enter the Big Pine–Taboose volcanic field. This field consists of more than 30 cinder cones and flows covering 400 square miles. All were emplaced during several episodes of volcanism spanning 2 million years. The north–south alignment of the cones strongly suggests that existing faults acted as feeders for the basaltic magmas. (See discussion at Stop 1-6).

194.9 (4.1) Continue past Goodale Road (Figs 1-12 and 1-13).

196.4 (1.5) Continue past a left turn to Taboose Canyon. We are approaching the Poverty Hills at 11:00. Rising 100 to 300 m (900 ft) above the valley floor and covering 3 square miles, the hills expose Paleozoic metasedimentary rock and Mesozoic intrusive rock along the Owens Valley fault in the axial part of Owens Valley (Fig. 1-14). The hills earned their name from the fact that they yielded little gold to prospectors compared to the nearby Fish Springs Hills 1.2 miles to the northwest.

199.9 (3.5) Continue past a wildlife view east at Elna Road.

200.7 (0.8) Move to the left lane and prepare to turn left.

201.1 (0.4) TURN LEFT (west) on Fish Springs Road at Tinemaha Reservoir.

201.6 (0.5) BEAR LEFT on Tinemaha Road at Griffith Road.

203.1 (1.5) PULL RIGHT into roadside turnout.

STOP 1-7. Poverty Hills. Two models have been proposed for the origin of the Poverty Hills. The most recent model during the past couple of decades suggests the hills are a tectonic uplift created by horizontal shortening at a left step in the right-lateral Owens Valley fault zone (Taylor and Dilek, 2001; Taylor,

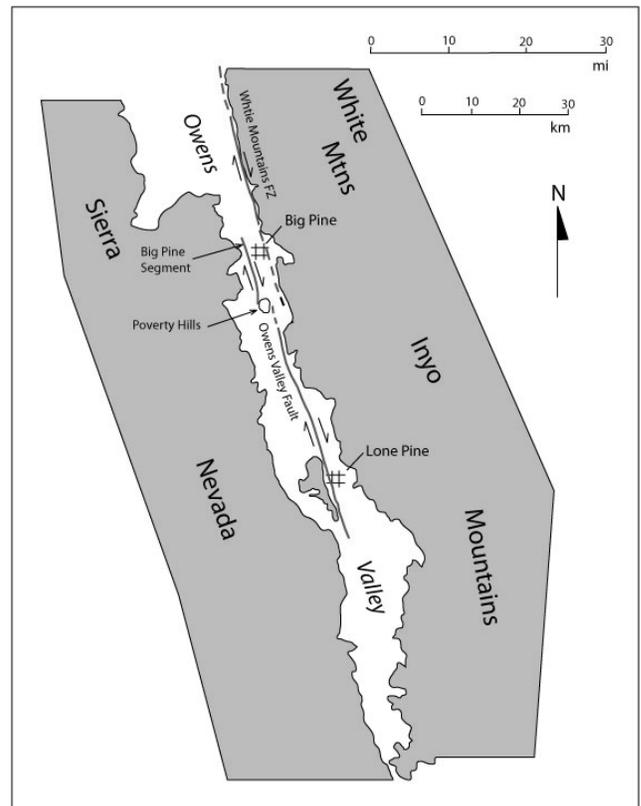


Figure 1-14. Location map showing Poverty Hills and Owens Valley fault. [K. M. Bishop]



Figure 1-15A. Light colored stratigraphic layer in Paleozoic Keeler Canyon Formation exposed along Tinemaha Road on the western edge of the Poverty Hills. The bed, shown outlined, was distorted by cataclastic breccia flow. Compare the distorted shape of the layer to that of a known rock avalanche deposit shown in Figure 1-15b. [K. M. Bishop photo]

2002). A recently advocated second model suggests the hills are a landslide mass derived from the Inyo Mountains (Bishop, 1999; Bishop and Clements, 2006), although the landslide model was first suggested in the 1960s by Pakiser et al. (1964) based on a gravity survey of the valley. If the hills are indeed a landslide, the long-run out character classifies it as a rock avalanche. Correct interpretation of the origin of the hills is important because it has implications for understanding the kinematics of the Owens Valley fault. The landslide model will be advocated here. The presence of widespread brecciation across the hills suggest a landslide deposit (Bishop and Clements 2006). At this stop we will hike down Tinemaha Road to view breccia exposures in road cuts.

Please watch for cars.

Of particular interest in the road cuts is the exposure of a brecciated light-colored stratigraphic unit within dark colored units near where we park. The color contrast of the units allows one to clearly delineate the results of cataclastic flow deformation of the breccia. The light-colored unit thickens and thins across the outcrop and has a contorted shape (Fig. 1-15A). It is proposed that the heterogeneous stress field required to create this pattern is the result of fluctuating internal stresses created by vibration and/or variable traction as the landslide mass moved rapidly across an irregular ground surface. The deformed geometry of the light-colored unit is reminiscent of the geometry of deformed stratigraphic units present in the Blackhawk landslide, a well-known rock avalanche deposit in the southern Mojave Desert (Fig. 1-15B).



Figure 1-15B. Light colored (outlined) stratigraphic layer in the Paleozoic Furnace Formation exposed in the Blackhawk rock avalanche in the southern Mojave Desert. [K. M. Bishop photo]

Advocates of the transpressional model for the Poverty Hills suggest the road cuts expose fault breccia (Taylor, 2002). In this hypothesis, a thrust fault accommodating uplift is present just below the road level. A counterargument to this idea is that the stress field from tectonic thrust faulting could be expected to be relatively homogeneous across the width of the outcrop. Thus, fault activity would not be expected to cause the contorted shape of the light-colored unit.

If the hills are a landslide mass, the landsliding occurred at least several hundred thousand years ago based on the degree of erosional dissection and the age of basalt flows that must post-date emplacement. The source area of the landslide is likely the Inyo Mountains where bedrock units of the same age and lithologies are exposed in the Santa Rita Flat pluton area.

Big Pine Volcanic Field. Basalt flows and scoria cones of the Big Pine volcanic field can be seen along both sides of U.S. 395 from the base of the Sierra Nevada Mountains to the west, to the Inyo Mountains to the east. The field encompasses an area of approximately 150 square miles from one mile south of the town of Big Pine (Stewart Lane, mp 217.5) to 8 miles north of Independence (Sawmill Road, mp 196.9). Volcanism dates from 1.2 Ma to as recently as 25 ka (Manley, et al., 2000; Moore and Dodge, 1980). Basaltic rocks dominate, but a small outcrop area of rhyolite is present to the west of the Poverty Hills.

Numerous active faults have been mapped within the Big Pine field. The Independence fault lies to the west of the basalt field and along the Sierra Nevada front. Le et al. (2007) reported that movement along

the Independence fault is dip-slip, east side down. The Fish Springs fault bisects the volcanic field offsetting several cones and flows. The alignment of flows and cones along the fault suggest it may have acted as an important conduit for ascending magmas. Beanland and Clark (1994) propose that motion along the Fish Springs fault is purely vertical, east side down. The right-slip Owens Valley fault lies less than a kilometer to the east of the Fish Springs fault. The White-Inyo fault lies along the east side of the Owens Valley. Bacon, et. al. (2005) suggest that Holocene motion along this fault is predominantly right oblique-slip, however, Bellier and Zoback, (1995) argue that motion along the White Mountains/ Inyo Mountains fault system underwent a change from predominantly dip-slip (west side down) to oblique slip at 288 ka. Alignment of cinder cones along the White-Inyo fault suggests it has also acted as an important channel for basaltic lavas.

Darrow (1972) published the first study of the Big Pine basalts, noting they were largely alkaline and attributing the alkalinity to thickened crust beneath the central Owens Valley. Waits (1995) sampled only a limited population of inclusion-bearing flows and stated that they were “highly alkaline”. Varnell (2006) examined the geochemistry and petrology of the Big Pine field in more detail. Figure 1-16 is a basalt tetrahedron compiled from chemical analyses of the Big Pine basalts.

Samples from the Big Pine field span the compositional range from alkali basalt to tholeiite. They do not show the restricted range of either the Coso or Ricardo fields or volcanic fields of the Mojave. Big Pine samples are spread almost equally between alkali basalt and tholeiite with only a small number lying in the olivine tholeiite subtriangle. Ringwood (1976) demonstrated that a thermal divide separated fractionating basaltic liquids and at low pressure it was not possible for an undersaturated *ne* normative magma to evolve into a *Q* normative tholeiite. This implies that the alkali basalts and tholeiites of the Big Pine field may have evolved separately and are the products of two differing events. Varnell (2006) also examined Big Pine basalts in thin section. She noted that olivine was present in some samples and absent in others. Nepheline was an occasional trace constituent. Darrow (1972) was the first to report the presence of iddingsite in Big Pine basalt, but unlike its

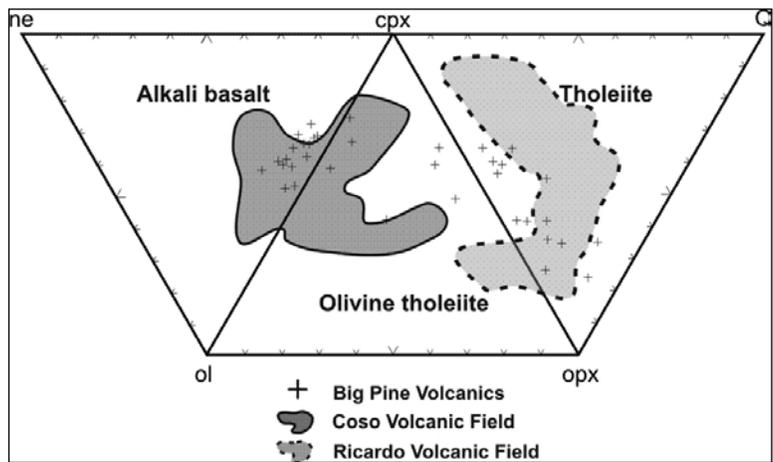


Figure 1-16. Basalt tetrahedron for the Big Pine volcanic field. Range of basalt compositions for the Coso (dark gray) and Ricardo (light gray) shown for comparison. Coso data from Grove (1996) and Ricardo from Anderson (2004).

ubiquitous occurrence in the Dove Spring (Ricardo) basalts, it is rare in the Big Springs basalts.

Compositional differences between the various Owens Valley volcanic fields have previously been attributed to differing melting depth (Anderson and Jessey, 2004; Wang et. al., 2002). However, the presence of iddingsite, an alteration product of olivine, lead Lusk and Jessey (2007) to propose that the initial composition of all Owens Valley magmas was alkaline and that the observed geochemical and petrographic differences are a function of differing evolutionary paths that involved an increase in oxygen fugacity for iddingsite-rich basalts. Reasons for the fugacity change are uncertain. Lusk and Jessey (2007) suggest-

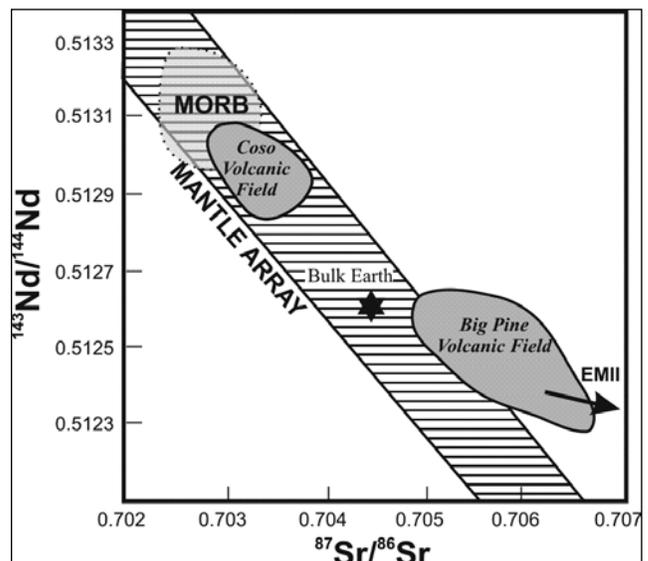


Figure 1-17. Epsilon neodymium diagram for xenoliths from the Big Pine and Coso volcanic fields. Data from the NAVDAT database.

ed it could be caused by assimilation of crustal rocks when the basaltic magma rose to shallow depth and became gravitationally stable. Heat from the ponded magma would melt the surrounding rocks increasing oxygen fugacity and changing the composition of the magma through assimilation, thus accounting for the significant compositional variation between the various Owens Valley basalt fields.

Evidence for this hypothesis comes from isotopes. Figure 1-17 is an ϵNd diagram for the Coso and Big Pine fields. Coso basalts lie wholly within the mantle array, suggesting Coso magmas were extracted from the mantle and extruded quickly after only limited interaction with the crust. Big Pine basalts, however, lie largely outside the mantle array and along an evolutionary trend (Zindler and Hart, 1986) requiring crustal contamination (Zindler and Hart, 1986). For Owens Valley basalt fields this contamination occurred when ponded basaltic magma assimilated crustal rocks. Proceed southwest to Tinemaha Creek Campground.

203.7 (0.6) **Turn around** at the intersection of Tinemaha Road and Fuller Road. **RETRACE** to Griffith Road. Red Mountain is the large volcanic cone to the west-southwest and the cinder cone of Crater Mountain is to the north.

205.3 (1.6) **TURN LEFT** on Griffith Road from Tinemaha Road. The view northwest at 10:00 is of a small red cinder cone with black basalt flows from Crater Mountain (elev. 6055') to the northwest.

205.7 (0.4) Join Fish Springs Road. Proceed north. The scarp of the Inyo/Owens/Fish Creek fault at the base of Crater Mountain is at 10:00.

207.0 (1.3) Fish Springs hatchery on the left is the largest in Owens Valley.

207.6 (0.6) **PULL TO THE RIGHT** and **PARK**.

STOP 1-8. Fish Springs Fault Scarp. Two scarps are exposed along the east face of Crater Mountain. Looking to the west, the Owens Valley fault (OVF) scarp is visible as an east-facing scarp. This part of the fault defines a 12 mile long segment of the OVF extending from the northwest side of the Poverty Hills to a termination north of Big Pine and referred to as the Big Pine segment. Almost due west of the stop is a tephra cone that is cut on its west side by the fault. Following with the eye northward from the cone, the scarp is first seen to cut alluvial fan material and then to cut basalt of Crater Mountain. The scarp cuts west

of a hill composed of basalt just north of the alluvial fan materials. Another scarp is developed on the east side of the hill. On the north side of the hill, the faults on either side of the hill merge. The portion of the Big Pine fault segment between the Poverty Hills and the point where the two faults merge is referred to as the Fish Springs fault.

From study of the tephra cone offset, Martel (1989) demonstrated that the Fish Springs fault is an east-side down dip slip fault with no strike-slip component. This poses a problem to the transpressional uplift model for the Poverty Hills. Without a right-lateral component of slip on the Fish Springs fault, there is no geometric reason for uplift of the western Poverty Hills.

Beanland and Clark (1994) proposed that the Big Pine segment contains a large strike-slip component north of the Fish Springs fault. As evidence, they listed the possible right-lateral offset of two hills along the fault at Crater Mountain, offset of a line of tree stumps near Big Pine, and offset of a stream channel north of Big Pine. Examination of each of these features suggests that the evidence is weak. The possible offset of the hills is questioned because it is not clear how the hills formed. They are localized in an unusual area of the fault, where the fault contains a graben structure. Until the origin of the hills is understood, it is arguable that the hills have been offset horizontally. The line of tree stumps is irregular and it is not clear the trees were not planted in a pattern that today suggests the possibility of right lateral offset. Furthermore, it is not even clear that the trees pre-date the most recent fault rupture. Finally, it is not clear that the proposed offset stream channels were ever directly adjacent to another.

In order to explain the lack of a strike-slip displacement on the Fish Springs fault, they suggested that strike-slip motion and dip-slip motion is partitioned at the latitude of the Fish Springs fault with dip-slip movement occurring on the Fish Springs fault and strike-slip movement occurring on the eastern fault. This suggestion is speculative in that there is no evidence for strike-slip displacement on the eastern fault. As an alternative in light of the weak evidence for strike-slip displacement on the Big Pine segment it is proposed that the entire Big Pine segment is a pure dip slip fault. We will consider evidence for this possibility at the next stop. Return to vehicles. Proceed north.



Figure 1-18. Upper—view to the north of the Big Pine fault segment near the town of Big Pine. Lower—view to the north of the Lone Pine fault east of the town of Lone Pine. The Lone Pine fault ruptured in 1872 and still has a steep, well-developed free face, whereas the Big Pine fault scarp is strongly degraded and does not contain a free face anywhere along its length. The lack of a free-face in the Big Pine fault is evidence that the fault did not rupture in 1872. [K. M. Bishop photos]

207.9 (0.3) **Stop** at southbound Hwy 395. Watch for traffic, cross southbound lanes, and **TURN LEFT** on northbound 395.

210.9 (3.0) Continue past a right turn to Steward Lane. Prepare for left turn onto Big Pine Dump Road.

211.2 (0.3) **TURN LEFT** on Big Pine Dump Road.

211.5 (0.3) **PARK** at waste transfer station. **WALK** up the gated dirt road to the Big Pine fault scarp.

STOP 1-9. Big Pine Fault Scarp. The current paradigm for this segment is that the fault last ruptured in 1872 and that it contains a predominant strike-slip component and a lesser dip slip component. The purpose of this stop is to examine the surface trace of the Owens Valley fault and consider evidence for the following two concepts: 1) the Big Pine segment did not rupture in 1872, and 2) the Big Pine segment's kinematics is pure dip slip.

Two lines of evidence suggest that the fault did not rupture in 1872. First, rock varnish is ubiquitously developed on rocks exposed in the fault scarp, as best seen in the basalt of Crater Mountain. Rock varnish takes approximately 2000 years to develop sufficiently to be visually discernible. If the Big Pine segment ruptured in 1872, then a zone of unvarnished rock disrupted by the fault movement should be present, but it is not. A second line of evidence is the state of fault scarp degradation where the fault cuts fanglomeratic deposits. Along its entire length, the maximum scarp gradient is no greater than approximately 30 degrees, the angle of repose for loose, sandy material. Nowhere is a “free face” (steep exposure of in-situ material cut by the fault) present. Wallace (1978) studied fault scarp degradation rates and suggests that in the Great Basin, free faces persist in fault scarps on the order of a couple thousand years following rupture. The lack of a free face anywhere along the Big Pine fault segment thus argues against 1872 rupture. This notion is well-illustrated by comparing the Lone Pine fault scarp, where a free face in fanglomerate created by 1872 rupture is still well-formed, and the Big Pine fault scarp, with its lack of a free face (Fig. 1-18).

Directly west of the Big Pine dump transfer station, the Big Pine fault cuts basalt to form a 150 foot wide graben. The west side of the graben is defined by two parallel east-facing scarps. The westernmost of the two faults is crossed by the channel of an intermittent stream. Channel erosion has developed in both the upthrown and downthrown fault blocks. There is no horizontal offset of the channel across the fault. It may be possible that the channel has been eroded since the last fault displacement. If so, the lack of horizontal offset offers no evidence of a horizontal component of offset. However, the scarp reaches 12 feet in height, which suggests it was created by multiple displacements. Given that the channel is incised approximately 6 feet into the downthrown fault block, it seems unlikely that the channel has been formed only since the last fault rupture.

Approximately 0.3 mile north of the incised stream along the main fault scarp is a line of boulders marking the edge of an ancient debris flow. This line of boulders crosses the fault scarp and is displaced downward across the scarp. There is no horizontal offset of the debris flow edge. This provides a second piece of evidence arguing against a horizontal component of displacement for the Big Pine fault.

The lack of evidence for 1872 rupture and for a right-lateral displacement component of the Big Pine fault segment suggests that the Big Pine fault segment may not be strongly linked kinematically to the Owens Valley fault south of the Poverty Hills. This, in turn, allows for the possibility that the Big Pine fault segment is sufficiently unrelated to the Owens Valley fault that it should be considered a separate fault. In addition, the Owens Valley fault south of the Poverty Hills may continue northward past the eastern edge of the Poverty Hills and into the Owens River floodplain, where it is buried by recent alluvium. The right lateral component of displacement along this fault, documented by Beanland and Clark (1994), rather than being transferred to the Big Pine fault segment as suggested by them, is proposed to be transferred to the right across the Owens Valley to the White Mountains fault zone along the east side of Owens Valley. This scenario is consistent with the presence of the "Tinemaha releasing bend" east of the Poverty Hills documented by Slemmons et al. (2008). Turn vehicles around and retrace to Highway 395.

211.8 (0.3) **Stop. TURN LEFT** on State Highway 395. Watch for traffic from both directions.

212.2 (0.4) Enter Big Pine. Continue past Bartell/Blake Street. Stay in the left lane

212.6 (0.4) Big Pine School is on the right.

213.1 (0.5) **TURN LEFT** on Crocker, between two gas stations with food marts.

213.6 (0.5) **PARK** on right. **WALK** west along the road 1/10 mile.

STOP 1-10. Big Pine fault scarp discussion. Return to vehicles, **RETRACE** to 395.

214.2 (0.6) **Stop** at Highway 395. **TURN LEFT** (north).

214.4 (0.2) **TURN LEFT** (west) toward Inyo County Parks campsite; Baker Creek campground.

214.7 (0.3) Stop sign.

215.2 (0.5) Cross the elevated Big Pine fault scarp and remember that this scarp was developed by the 1872 earthquake of magnitude 7.5+ within a mile of the communities of Big Pine, Independence, and Lone Pine! The Owens Valley earthquake is the third largest historic earthquake in the conterminous United States (Ellsworth, 1990; Beanland and Clark, 1994).

215.4 (0.2) Enter Baker Creek campground. **TURN AROUND** and **RETRACE** to Hwy 395.

216.6 (1.2) **Stop** at Hwy 395. Watch for cross traffic. **TURN LEFT** (north).

216.8 (0.2) Continue past the right turn to CA 168 east leading to Westgard Pass, Waucoba Canyon, Goldfield, and Tonopah, Nevada.

218.1 (1.3) Continue past a left turn to Reynolds Road. View across Klondike Lake to Owens Valley Radio Observatory.

218.8 (0.7) Continue past a right turn to Klondike Lake.

223.8 (5.0) Continue past a left turn to Keough Hot Spring (a resort with swimming pool, picnic area, etc.).

225.5 (1.7) Continue past a left turn (west) for Collins Road. Domestic water from the SNFF comes hot from the ground, and homes have to cool the water before drinking.

228.5 (3.0) Continue past a right turn (east) to Warm Springs Road.

230.0 (1.5) Continue past Schober Lane to Bishop.

230.5 (0.5) Enter Bishop.

231.0 (0.5) Continue past Line Street (Highway 168 west) at traffic signal.

231.6 (0.6) Yaney Street, with visitor center on the right.

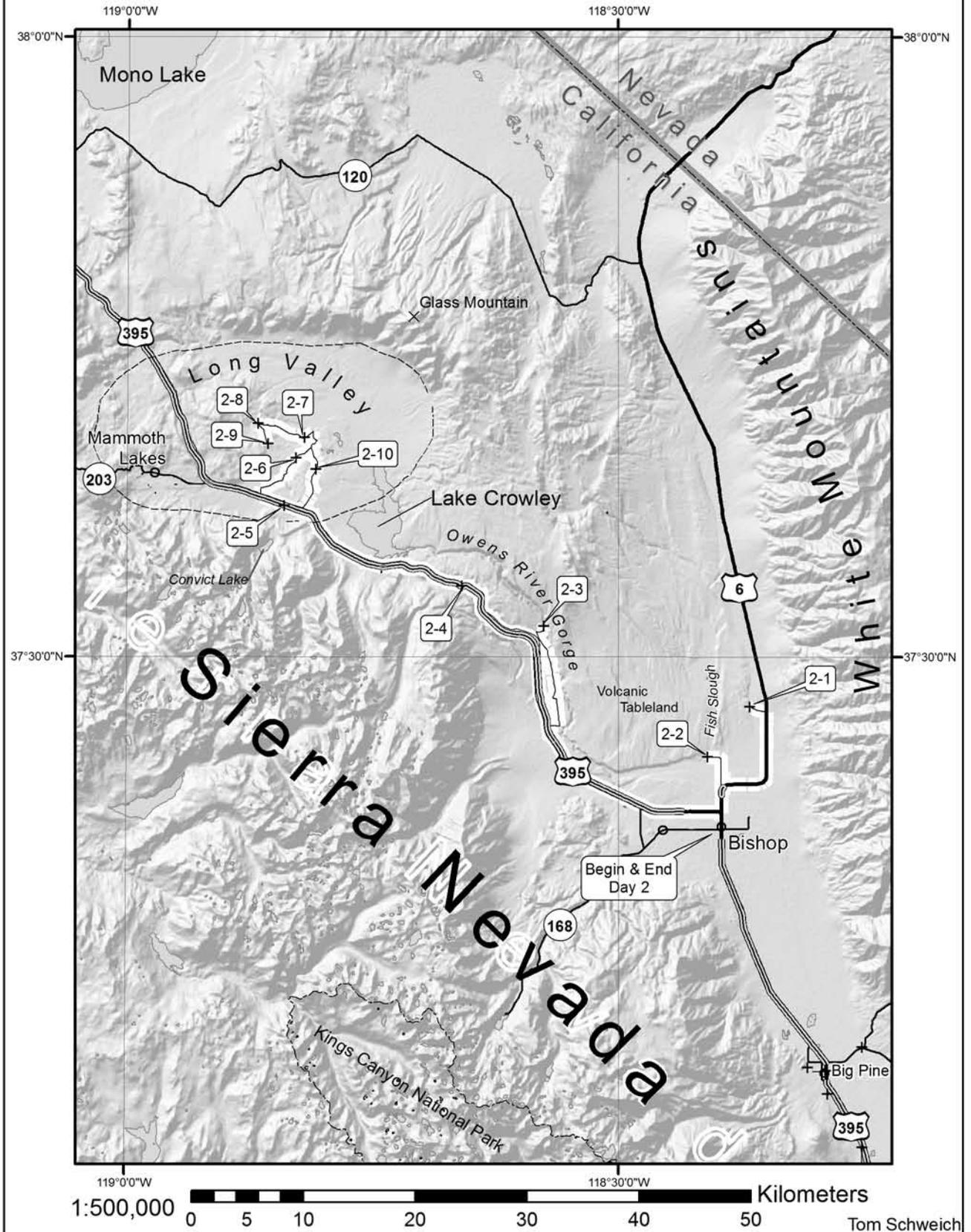
231.9 (0.3) We are at the junction of Hwy 395 and Hwy 6. Hwy 395, here called the North Sierra Highway, bends due west and continues to Sherwin Grade, the Long Valley Caldera, and Mammoth. Highway 6 turns northeast, passing the community of Laws and the Railroad Museum located on Silver Canyon Road, and continues toward Benton and Tonopah.

End of Day 1

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2009 Desert Symposium Field Trip

Day 2: Features Near And In The Long Valley Caldera



Day 2**Theme: Glaciers and snow where the hot springs flow**

with contributions from: D. R. Jessey, Steve Lipshie, R. E. Reynolds, Janet Westbrook, and Tom Schweich

Start Day 2

What We Will See: Our 130 mile route north of Owens Valley through Round Valley and into the Long Valley Caldera will reach elevations of 7,000 feet, as we travel parallel to the Sierra crest with average heights of 12,000 feet. The Long Valley Caldera contains numerous hot springs, evidence of fumaroles, and signs of hydrothermal alteration. Timber, minerals, rock commodities, and water—hot and cold—have been exploited over the last 160 years. Our route crosses eastern Sierra canyons where evidence of the powerful forces of Ice Age glaciers remains.

0.0 (0.0) Line St. in Bishop.

0.5 (0.5) Visitor center at Yaney Street. Move to the right lane.

0.8 (0.3) **CONVENE** at Giant Sequoia Tree and the 3 Flags at the intersection of highways 395 and 6. Sequoias were planted in 1918. There are seven in Big Pine, more than 25 in Bishop, three in Independence, and at least one in Lone Pine.

Bishop, CA, founded in 1864, has a population of approximately 4000. The town was named after Bishop Creek; the creek was named after Samuel Addison Bishop, a cattle rancher in the Owens Valley. The city of Bishop calls itself the “Mule Capital of the World” and hosts a festival called Mule Days over the Memorial Day weekend.

BEAR RIGHT at traffic light and **PROCEED NORTH** on Highway 6 when it intersects with Highway 395.

1.8 (1.0) Bear right on Hwy 6 at Dixon Lane.

2.1 (0.3) Continue past Five Bridges Road on Hwy 6.

4.6 (2.4) Cross the Owens River. Pass Silver Canyon Road, which leads to the Laws Railroad Museum, on the right.

6.1 (1.5) Continue past Jean Blanc Road. Highway 6 bears north.

8.5 (2.4) **Prepare for left turn.**

8.7 (0.2) **TURN LEFT** on Rudolph Road to the Chalfant Quarry. If you pass a Mono County sign, you’ve

gone too far.

9.5 (0.8) **BEAR RIGHT** as road forks (NW).

9.6 (0.1) **BEAR LEFT** to the quarry.

9.8 (0.2) **STOP 2-1. Chalfant Quarry** (N 37°27’36.1”; W 118°21’58.3”). The Bishop Tuff, dated at 760 ka, was produced by a 6-day-long eruptive event from the Long Valley Caldera (Hildreth and Wilson, 2007). Air-fall ash remnants have been found from western California to Nebraska, an area of 1×10^6 mi². Pyroclastic flows spread 40 miles to the southeast down Owens Valley, 20-30 miles east to the White Mountains and 30–50 miles north into Mono Basin. The caldera collapsed into an 8 mi x 14 mi elliptical depression. It has been subsequently enlarged by slumping and erosion to its current 11 mi x 20 mi dimension.

The Bishop Tuff consists predominantly of biotite–plagioclase–quartz–sanidine high-silica rhyolite. Ash fall deposits are plinian pumice lapilli and crystal-rich ash. Ash flows, as thick as 500 ft, range from nonwelded to eutaxitic, and from unconsolidated, or welded vitric zones to fully devitrified zones (Hildreth and Wilson, 2007). Eruptive volume is not well known but Hildreth and Wilson (2007) estimate that 150 mi³ of magma was erupted to produce the Bishop Tuff.

Two main major units of the Bishop Tuff deposit are visible here (Fig. 2-1). The lower 12 feet of the section consists of the poorly-sorted air-fall tuff that was deposited downwind from the



Figure 2-1. Two units of the Bishop Tuff exposed at the Chalfant Quarry. The lower 12 feet of the section consists of airfall pumice, and the upper 15-18 ft of the section is comprised of the basal portion of a thick pyroclastic flow. The dark layers just below the contact are manganese oxide stains caused by groundwater circulation. [Source: <http://lvo.wr.usgs.gov/>]

Table 2-1. Major Element Analyses of the Bishop Tuff (Cal Poly-Pomona)

	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅
Chalfant Ash Flow Tuff	77.88	0.08	11.80	0.93	0.05	0.07	0.45	3.52	5.21	0.018
Chalfant Air Fall Tuff	78.07	0.08	11.80	0.99	0.04	0.08	0.56	3.36	5.03	0.016
Owens River Upper BT	78.22	0.07	11.60	0.86	0.03	0.05	0.44	3.65	5.08	0.013
Owens River Lower BT	78.24	0.07	11.83	0.92	0.03	0.05	0.50	3.42	4.95	0.015

eruption of the Long Valley caldera. The upper 15-18 feet of the section consists of the basal portion of the pyroclastic flow that comprises much of the Volcanic Tableland to the west. At this location, it is remarkably well-sorted for a pyroclastic flow. The “dark layers” just below the contact between the two units are manganese oxide stains resulting from groundwater circulation. Table 2-1 summarizes the average major element content of 32 samples taken from the two units exposed in the Chalfant Quarry and the Upper and Lower Bishop Tuff of the Owens River gorge (Stop 2-3). Statistically, the four sample locations/units are indistinguishable. This supports the theory that the Bishop Tuff was emplaced in a single event over a limited span of time.

View east of ranches and trees along the scarp of the fault at the base of the White Mountains. Two strands show right lateral offset. **RETRACE** to Hwy 6.

10.9 (1.1) **Stop** at Hwy 6. Watch for traffic; **TURN RIGHT** (south).

13.6 (2.7) Continue past Chalk Bluff Road.

14.5 (0.9) **Bear right** (west) at Silver Canyon Road.

17.3 (2.8) **TURN RIGHT** (north) on Five Bridges Road just after a sweeping turn left.

18.9 (1.6) The highway bears left (west) at Jean Blanc Road.

19.7 (0.8) **TURN RIGHT** (north) on Fish Slough Road.

20.4 (0.7) Pavement ends.

20.7 (0.3) **PARK** at complex intersection.

STOP 2-2 (37° 25' 10.1" N, 118° 24' 33.6" W). **Fish Slough.** Fault-aligned springs at Fish Slough have created a unique biological oasis containing rare or threatened species of plants and animals. The East Side Bluff is five miles long and 300 feet high. The boulders and cliffs of the fault scarps offer excellent hunting, perching, and nesting sites for raptors.

During the 1860s ranching and mining home-

steads developed at “The Bend” of the Owens River (Bishop and Laws). Phillip Keough set up a stage stop in 1890 near the northwest spring of Fish Slough. Fish Slough Road became a main freight wagon and cattle route connecting Bishop and Laws to the prosperous mining camps of Benton Hot Springs, Bodie, Aurora, and as far north as Reno and Carson City. In the 1930s the City of Los Angeles bought riparian lands in Owens Valley, including Fish Slough, to acquire water rights and supply a growing city, thus precluding intensive agricultural uses and private land development.

The three perennial Fish Slough springs and their outflows are surrounded by wet marshlands with bulrushes, cattails, sedges, willows, and cottonwoods. These are bordered by seasonally wet alkali meadows, frosted with a white mineral crust. Beyond the meadows lies a saltbrush scrub community typical of the northern Mojave Desert. Outside the marsh, Great Basin plants survive extreme heat and cold and drought. The saltbush scrub gives way to Great Basin desert sagebrush communities at higher elevations (Forbes, et al., 1991; Odion, et al., 1992; BLM, 2007).

North of Fish Slough there is an abrupt loss of Mojavean species, and there is an abrupt loss of Great Basin species to the south.

Haller, et al. (1992) compared the phyto-geographic relationships of the wetland flora of Fish Slough to those of other desert wetlands east of the Sierra

Table 2-2. Plants and birds at Fish Slough

Plants common and rare	Birds endemic and migrant
bulrushes	sage sparrow
cattail	rock wren
sedges	mourning dove
willow	golden eagle
cottonwood	prairie falcon
shadscale	cinnamon teal
Parry saltbush	mallards\
four-winged saltbush	ruddy duck
spiny hopsage	northern pintail
bud sage	gadwall
Indian ricegrass	great blue heron
desert trumpet	American bittern
Venus blazing-star	northern harrier
Fish Slough milk vetch	
alkali mariposa lily	
alkali cordgrass	
Great Basin centaury	
hot springs fimbry	
hot spring fimbriylis	
King's mousetailivesia	
silverleaf milk vetch	

Nevada. Fish Slough, with 126 wetland taxa, has the richest flora of any of the nine areas surveyed. Fish Slough wetland flora is most similar to Fish Lake Valley and Deep Springs Valley to the east, and least similar to Saline Valley. Fish Slough is outstanding, since 34 of its species do not occur at any of the nine other wetland areas surveyed. Species with primarily Mojavean (8 percent) and Great Basin (10 percent) distribution are about equally represented in Fish Slough wetlands.

Several protected plants thrive here: Fish Slough milk vetch (*Astragalus lentiginosus* var. *piscinensis*) is a federally listed threatened species. You cannot find this low-growing perennial forb anywhere in the world except the alkali meadows of Fish Slough. Nearby are white flowers of the Inyo County star-tulip, the Alkali mariposa lily (*Calochortus excavatus*; BLM Sensitive Species). Five other rare plant species depending on the Fish Slough are alkali cordgrass (*Spartina gracilis*), hot springs fimbry (*Fimbristylis*), hot spring fimbri-stylis (*Fimbristylis thermalis*), Great Basin centaury (*Zeltnera exaltata* Syn: *Centaurium* (*Centarium exaltatum*)), King's mousetailivesia (*Ivesia kingii* var. *kingii*), and silverleaf milk vetch (*Astragalus argophyllus* var. *argophyllus*).

Fish Slough is also critical aquatic habitat for the Owens pupfish (*Cyprinodon radiosus*), Owens tui chub (*Gila bicolor snyderi*), and the Fish Slough springsnail (*Pyrgulopsis pertubata*). Owens pupfish was found throughout the Owens Valley from 1859 to 1916, but water diversions and wetland drainage by farmers caused the fish to drastically decline until they were thought to be extinct in 1948. Two of Fish Slough's species, the Owens pupfish and the Owens tui chub, are listed as endangered. In 1982, the Bureau designated Fish Slough as an Area of Critical Environmental Concern (ACEC) to preserve the unique aquatic ecosystem (BLM, 2002) – but a small population was rediscovered in Fish Slough. These pupfish feed on insect larvae and small aquatic creatures and can survive in warm, salty water with low oxygen levels (see Schoenherr, this volume). The males turn blue and silver during breeding season and vigorously defend small territories. The Owens speckled dace is a subspecies of speckled dace found in some places in the Owens River. Fish and Game hopes to reintroduce the Owens tui chub to the Slough.

RETRACE to Highway 6.

21.1 (0.4) Resume pavement.

24.2 (3.1) **Stop** at Highway 6. **TURN RIGHT** (south).

25.4 (1.2) Enter Bishop.

25.7 (0.3) **Stop** at signal and **TURN RIGHT** (northwest) on Highway 395.

27.2 (1.5) Traffic light at Barlow Road. Proceed on Highway 395.

31.7 (4.5) Sawmill/Pleasant Valley roads junction. Outcrops of Bishop Tuff are present on both sides of US 395. Ahead on the skyline is the escarpment of Wheeler Crest, composed of Wheeler Crest Granodiorite. Proceed on Hwy 395.

34.3 (2.6) Continue past Mill Creek Road.

35.6 (1.3) Continue past Pine Creek Road. If you want to visit the abandoned Pine Creek tungsten mine, turn left and drive 10 miles west along Pine Creek Canyon Road to a locked gate. The mine lies about 0.5 mile beyond the gate. The mine was closed in 1990 and more recently (2001) the existing mill was sold and dismantled. About 600 people worked at Pine Creek during its World War II heyday. The mine began production of tungsten in 1918 and was acquired by Union Carbide in 1936. Union Carbide sold the mine and mill to Strategic Minerals Corporation in 1986. Avocet Ventures bought the mill and 50 percent of the mine in 1995.

The mine was the largest tungsten supplier in the U.S. Between 1937 and 1990, 83,600 tons of WO_3 (tungsten trioxide), 12,700 tons of molybdenum byproducts, 18,300 tons of copper, 1.8 million ounces of silver, and 3600 ounces of gold were produced (Kurtak, 1997). The ore occurs as a contact metamorphic skarn formed when Paleozoic roof pendant rocks reacted with hydrothermal fluids generated by intrusion of the Sierra Nevada batholith. Continue straight on Hwy 395 around the western edge of Round Valley.

37.2 (1.6) Begin the climb up the Sherwin Grade. Highway 395 ascends 2500 feet over the next 10 miles. The grade is named for James Sherwin, owner of a ranch in Round Valley. In 1874, Sherwin built a toll road up the grade of Rock Creek to the vicinity of present-day Toms Place. For years, U.S. Highway 395 followed the winding route of Sherwin's toll road, until the existing four-lane highway was built in 1956–57 across the volcanic tableland.

37.5 (0.3) **Prepare to turn right**; look for “Paradise/Swale” sign.



Figure 2-2. Collapsed aqueduct pipe, 1980s. [J. Westbrook photo]

37.7 (0.2) **TURN RIGHT** at Paradise/Swale Meadows. Proceed east on Owens Gorge Road.

38.4 (0.7) **TURN LEFT** (north) at T intersection. The large pipeline that parallels the road is a conduit that carries Owens River water to Control Gorge Powerhouse No. 3 in the gorge.

39.0 (0.6) Roadcut through a fumarolic mound. The white, hydrothermally altered tuff consists primarily of microcrystalline cristobalite, tridymite, and alkali feldspar. The numerous low, rounded mounds on the surface of the tableland are believed to be the sites of formerly active gas vents. The tuff near fumaroles has been highly indurated and is less susceptible to weathering, leaving the more resistant rocks at sites of former gas vents as low mounds.

39.8 (0.8) LADWP surge tank straight ahead. Surge tanks are empty tanks designed to absorb the energy of the first surge of water that comes through the line when the valves are opened. Energy is dissipated by compressing the air in the tank.

42.4 (2.6) Continue past the right fork toward LADWP Middle Gorge and drive to the locked gate.

45.0 (1.6) The road crosses a small gorge cut into Bishop Tuff.

45.4 (0.4) The paved road forks. Take the right fork.

45.6 (0.2) **STOP 2-3. Owens River Gorge** (N 37°31'30.8"; W 118°34'31.6") and LADWP Power Plant Middle Gorge. For comparison with Day 1 Cottonwood and Division Cr. plants, the Owens Gorge Power Plants are “triplets,” each generating 37.5 megawatts. The three plants operate simultaneously, and generate 100 megawatts. They also operate on aqueduct flow minus the water used in the Owens Gorge Fishery. About 120,000 acre feet of water goes through the plants each year, generating 225,000 megawatt-hours, enough for 200,000 homes. The penstock in Owens Gorge is eight feet in diameter.

Why two colors of pipe? An LADWP workman made a very big mistake in the 1980s—he shut down an upper valve but forgot to shut down the lower one first, causing a huge water hammer which collapsed the entire pipe between the valves (Fig. 2-2). The tan pipe is a replacement from Germany. While awaiting the replacement pipe, water flowed down Owens Gorge, reestablishing the riparian woodland and the fishery.

The Owens River has eroded downward nearly 500 feet through Bishop Tuff at this locality. The tuff is comprised of two units. The upper unit (Ubt) is poorly indurated and has striking radial columnar jointing (Fig. 2-3). Column diameters typically range between 3 and 5 feet (Gilbert, 1938). Most columns



Figure 2-3. Bishop Tuff exposure in the Owens River gorge. Contact between poorly indurated Upper Bishop Tuff (Ubt) and welded Lower Bishop Tuff (Lbt) shown by the black line. [D. R. Jessey photo]

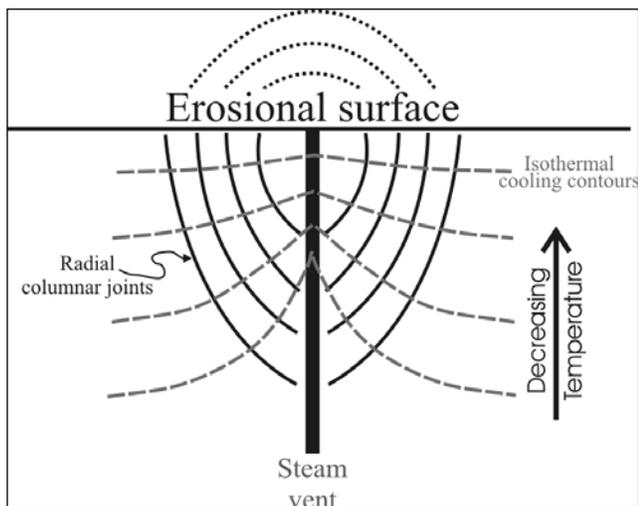


Figure 2-4. Computer-generated model for thermal cooling around a fumarolic vent. [D. R. Jessey]

are oriented in a radial pattern. The lower Bishop Tuff (Lbt) is a strongly welded, massive tuff with irregularly developed vertical jointing.

The Upper Bishop Tuff consists of pale pink, poorly welded, vitric pumice ash. It readily darkens to gray on weathered surfaces. The Ubt contains abundant pumice shards as well as phenocrysts of sanidine, quartz, and plagioclase. The Lower Bishop Tuff is more strongly welded with flattened and elongated pumice fragments common. (See if you can locate the contact as you walk down the DWP access road; in 2003 it was marked by Chinese writing!) Unweathered Lbt is pale red to gray and noticeably denser than the overlying Ubt. In all respects, the lower unit resembles a “textbook” ash flow welded tuff. Note from the table presented at the last stop that the two units are chemically very similar.

The origin of the radial joint sets remains controversial. However, Mike Sheridan, while a graduate student at Stanford, undertook a computer simulation of gas flow in geothermal systems. He suggested (Sheridan, 1970) that each joint set represents the locus of fumarolic activity (Fig. 2-4). The radial jointing is similar to the heat flow pattern developed during cooling around a gas vent. The joints form normal to isothermal surfaces. If

you look to the east, across the gorge you will see hummocky terrain with numerous surface bumps. Each “bump” would be the location of an inactive fumarole.

Turn around and RETRACE to Hwy 395.

45.8 (0.2) TURN LEFT (south) on Gorge Road.

51.9 (6.1) TURN RIGHT toward Hwy 395. View at 10:00 to the west-southwest of lateral moraines at Pine Creek.

52.6 (0.7) Stop at Hwy 395. TURN RIGHT. Proceed north on Hwy 395 up Sherman Grade.

53.9 (1.3) Mono County Line. Volcanic tableland extends from here back to Middle Gorge.

57.9 (4.0) Vista Point on the right. The town of Bishop lies 13 miles to the southeast, Round Valley six miles to the south, and the escarpment of the Wheeler Crest to the southwest. The summit of Mt. Barcroft is 20 miles to the east in the White Mountains.

60.3 (2.4) Crossroad.

62.2 (1.9) Summit of Sherwin Grade, el. 7000 ft.

63.1 (0.9) Prepare for left turn. Watch for oncoming traffic.

63.3 (0.2) TURN LEFT and proceed east to the turn-out on the left side of Lower Rock Creek Road.



Figure 2-5. The “Big Pumice Cut” along Highway 395. The bouldery Sherwin Till is overlain by Bishop Tuff. The basal air-fall tuff is overlain by a much thicker ash-flow. An angular unconformity separates the units. Several prominent “clastic dikes” cut the tuff layers. [Photo by S. Lipshie, 1976]

63.7 (0.4) **PARK** at turnout.

STOP 2-4. Big Pumice Cut (N 37°33.27.4"; W 118°39'31.9"). Blackwelder (1931) first described the Sherwin till. This till is one of the oldest Pleistocene glacial deposits in the Sierra Nevada, exceeded in age only by the McGee till. Blackwelder believed that the Sherwin till was younger than the Bishop Tuff. However, Gilbert (1938) stated that the tuff is younger than the Sherwin till. Putnam (1960) studied the till-tuff at several localities including the Big Pumice Cut and concluded, as had Gilbert, that the Bishop tuff overlies the Sherwin till.

However, Rinehart and Ross (1964) questioned whether the till that underlies the Bishop Tuff was really Sherwin till because of its apparent great age. Ash collected just above the till-tuff contact from the Big Pumice Cut was dated at about 710 ka (Dalrymple and others, 1965). That age was thought to be too old for the Sherwin glaciation, generally regarded as being equivalent to the Kansan and/or Illinoian glacial stages of North America. Sharp (1968) undertook a comprehensive study of relationships between the till and the tuff in the vicinity of the Sherwin Grade. He demonstrated conclusively that the Sherwin till is older than the Bishop Tuff and that the till exposed in the Big Pumice Cut is Sherwin till. A more recent $^{40}\text{Ar}/^{39}\text{Ar}$ date of 760 ka for the Bishop Tuff (Van den Bogaard and Schirnick, 1995) is the currently accepted age for the unit. Because the till appears to have undergone about 50,000 years of weathering prior to burial, it was thought to be around 800 ka in age by Sharp and Glazner (1997). This suggests that it correlates with the Kansan glacial stage.

When viewing the Big Pumice Cut (Fig. 2-5), note the contact between the boulder-laden till and the overlying, poorly indurated tuff. The basal 15 ft of the tuff is composed of air-fall material, while the overlying layer is an ash-flow. Within the tuff, layering of the air-fall ash is parallel to the Sherwin till erosional surface, whereas bedding in the ash-flow is horizontal. You should be able to see the angular unconformity between the two tuff units, although

recent weathering has greatly obscured that relationship.

Also note, the tuff is cut by a series of vertical clastic dikes, perpendicular to layering, that extend into the till. The dikes, derived from gravel deposited on the tuff, consist of unconsolidated sand and gravel fining downward. Wahrhaftig (1965) suggested that the gravel originated as glacial outwash that slumped into fissures formed in the tuff.

RETRACE to Hwy 395.

63.9 (0.2) **Stop** at 395. **Watch** for oncoming traffic! **TURN LEFT** and proceed north on 395.

58.6 (1.0) Pass Tom's Place/Rock Creek Road. Owens River Gorge Road is on the right. Toms Place is named for Tom Yerby, who built a general store and cabins here in 1917. The town, once considered a backwater trailhead, has become the "hot" property of the eastern Sierra. Lots sell in the six figures, and million dollar vacation homes are not uncommon.

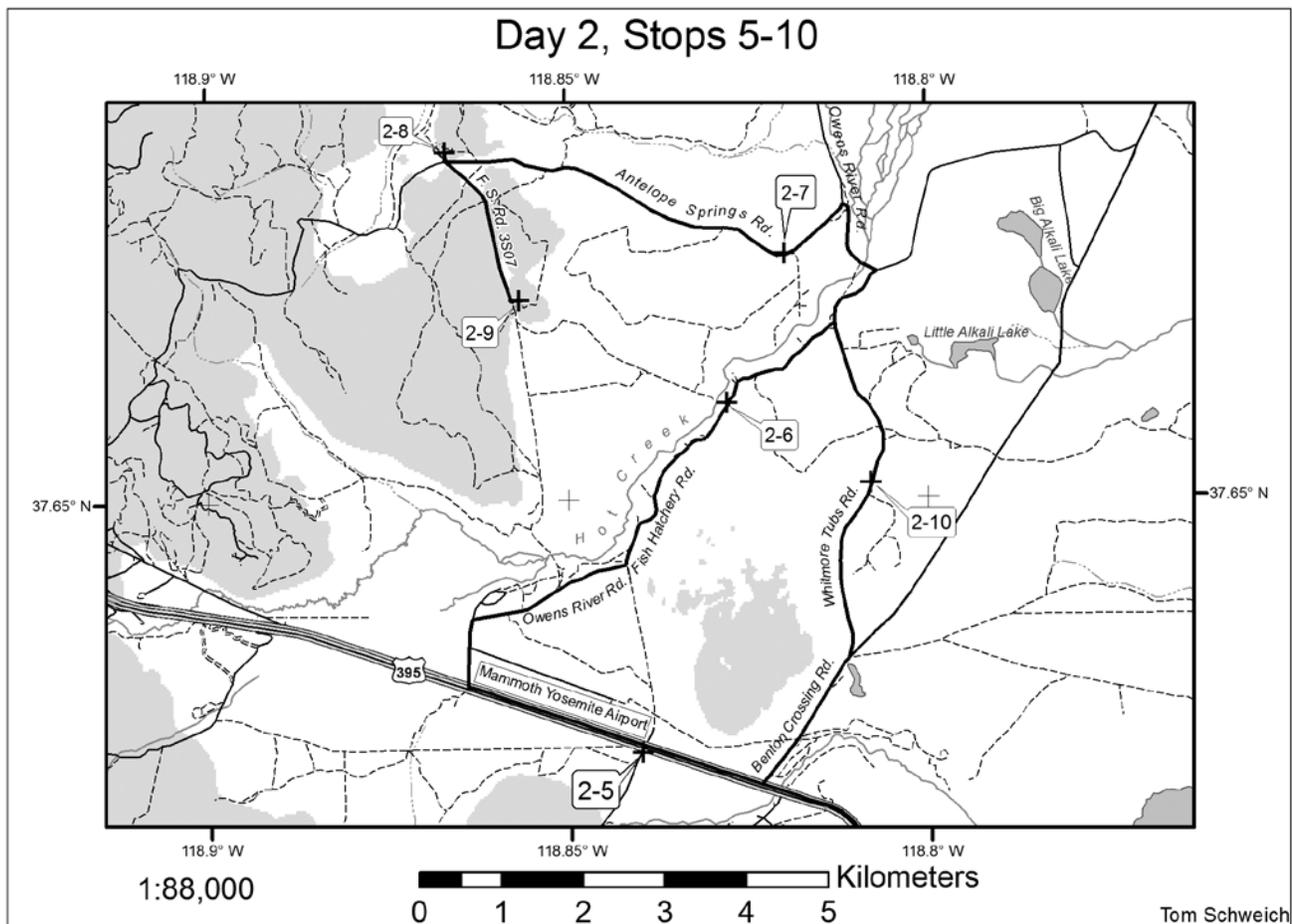
66.2 (2.3) Cross Crooked Creek Gorge.

66.6 (0.4) Continue past the turn to Hilton Creek and Crowley Lake on the right. Hilton creek is named for Richard Hilton, who had a dairy ranch along the creek in the 1870s.

67.9 (1.3) McGee Creek lateral moraines are to the



Figure 2-6. Scarp of the Hilton Creek fault crossing a Tioga lateral moraine. McGee Creek flows to the north through the trees at the bottom of the photo. McGee Creek Road lies at the base of the moraine. [Source: Lipshie, S., 2001, *Geologic guidebook to the Long Valley-Mono Craters Region of Eastern California*, 2nd Edition: South Coast Geological Society Field Trip Guidebook]



west.

69.4 (1.5) Pass McGee Creek Road on the left. Two miles up McGee Creek Road, near the McGee Creek Campground, is a 50-ft scarp within a Tioga moraine created by the Hilton Creek fault (Fig. 2-6). In June and July, 1998 two earthquakes of $M=5.3$ ruptured the Hilton Creek fault. The June quake had its epicenter (right lateral strike-slip motion) near the campground; the July epicenter (normal movement) was 1.2 miles to the west. During one of the two quakes a large boulder rolled down the hill from where it had been perched on the scarp of the Hilton Creek fault into the campground leaving a trail of flattened vegetation and shattered pavement.

Just to the north of this location are the epicenters of the $M=6.0$ earthquakes of 1980 that heralded the renewal of seismic activity along the Hilton Creek fault. Offset from those quakes has been as much as a foot in the vertical plane. The triggering mechanism for the earthquakes remains controversial, but one intriguing theory suggests that it is related to magma injection along fractures. This causes the fault blocks to be shoved aside. Seismic surveys suggest magma

may be only 1–3 miles below the surface in this area.

Long Valley Caldera. Enter the southern margin of the Long Valley caldera (see Jessey, this volume: p. 135, fig. 1), covering 180 square miles (4,000 square km) and stretching 18 miles east–west, 10 miles north–south, and forming the basin with Glass Mountain to the northeast, Bald Mountain to the north, and Mammoth Mountain on the west. Volcanic and tectonic activity over the past 4 million years formed spectacular landscape around the Long Valley caldera and the Mono Basin. The Sierra Nevada and White Mountain faults produced impressive relief of the eastern Sierran and White Mountain escarpments bordering northern Owens Valley starting 3 million years ago. Two distinct and related magmatic systems dominated the volcanic evolution of the Long Valley during this time (Bailey and others, 1976; Bailey, 1989). Compositions of lava started as basaltic and became progressively silica-rich through time, producing rhyolitic eruptions.

The older magmatic system produced widespread eruptions of basalt and andesite between 3.8 and 2.8 Ma over much of Long Valley and Mono Basin. The

Sierra Nevada fault system has offset some of these early lava flows as much as 1,000 m, leaving the Sierra crest high and Owens Valley low. Volcanic activity in the vicinity of the present Long Valley caldera produced rhyolite eruptions at 3.1 to 2.5 Ma followed by high-silica rhyolite from 2.1 to 0.8 Ma. Lava from the latter eruptions form Glass Mountain on the northeast rim of the caldera.

The Glass Mountain eruptions culminated in the catastrophic eruption of 600 cubic kilometers of rhyolite 760,000 years ago. This eruption deposited the Vishop Tuff over wide areas of the southwest, and caused the simultaneous 2- to 3-km subsidence of the magma chamber roof to form the present oval depression of Long Valley caldera. Subsequent eruptions were confined within the Long Valley caldera with accumulations of crystal-free rhyolite 700,000 to 600,000 years ago. The caldera floor warped to form a resurgent dome followed by extrusions of crystal-rich rhyolite at 200,000-year intervals (500,000, 300,000, and 100,000 years ago) in clockwise succession around the resurgent dome within the caldera (From Ewert and Harpel, 2000).

71.8 (2.4) Watch for 1950s Caltrans-installed snow fences to the west. Fences were constructed upwind to keep wind-driven snow off the highway. Snow drifts build up along the eastern side of the fences, helping keep the road open with minimal plowing. The effects of higher and lower amounts of precipitation over time can be evaluated by looking at shrubs on either side of the snow fences. Near Deadman Creek, Michael Loik (UC Santa Cruz) and graduate students are studying simulations of increased and decreased snow depth scenarios as envisioned by some climate change models. Fluctuating snow depth results in variable soil moisture available for shrub and tree growth in the subsequent spring and summer, effects resonating through processes in the ecosystem such as litterfall, root growth, soil carbon storage, and soil nitrogen content. Their work on carbon, water, and nutrient fluxes driven by snow depth/melt timing in this system helps understand community-level responses to snow depth changes. Such responses include natural recruitment of Jeffrey and lodgepole pines (*Pinus jeffreyi* and *P. contorta*), as well as Great Basin sagebrush and antelope bitterbrush (*Artemisia tridentata* and *Purshia tridentata*) (McNulty, 2004).

72.3 (0.5) Pass the turnoff to SNARL (Sierra Nevada Aquatic Research Laboratory) on the left. SNARL was established along Convict Creek by the U.S. Bureau of

Fisheries and Wildlife in 1935. When the Feds phased out its operation in 1973, SNARL was acquired by the University of California. At present, the facility provides research space and accommodations for small field groups and is administered as a part of the Valentine Eastern Sierra Reserve by UC. The site has a manmade stream system, consisting of nine meandering channels used for research on stream hydrology and ecology.

72.5 (0.2) Pass a right turn to Benton Crossing Road leading to Whitmore Hot Springs.

73.4 (0.9) **TURN LEFT** at Convict Lake Road.

73.5 (0.1) **PARK** along right shoulder of the road to view the Convict Creek moraines.

STOP 2-5. Convict Creek Moraines (N 37°37'19.5"; W 118°50'25.0"). Looking south toward Convict Lake, one can see a 900-foot-high ridge that is a gigantic lateral moraine separating Tobacco Flat to the south from Long Valley to the north. The moraine is thought to be of Tahoe age. On the right (west) side of the road is a massive Tioga moraine made up of a series of arcuate end moraines.

McGee Mountain is the flat-topped mountain to the left, beyond the Tahoe lateral moraine. On top of McGee Mountain is an old erosional surface. Age dating of andesite preserved on the erosional surface suggests a minimum age of about 2.6 Ma (Dalrymple, 1963). The McGee till overlies the Pliocene andesite. Boulders and cobbles in the till consist mostly of Round Valley Peak Granodiorite, with lesser amounts of metamorphic clasts (Rinehart and Ross, 1964).

From geomorphic evidence, Putnam (1960 and 1962) postulated vertical displacement along range-front faults at McGee Mountain amounting to 3000 feet between the Sherwin and Tahoe glaciations and an additional 1000 feet between the McGee and Sherwin glaciations (Table 2-3). Thus, the unusual perched glacial deposit atop McGee Mountain provides strong evidence for substantial Plio-Pleistocene faulting and growth of the eastern escarpment of the Sierra Nevada in the last 2.5 million years.

Further south along the Convict Lake Road, Mount Morrison is the dramatic pointed peak beyond the Tioga lateral moraine on the left (south) side of the Convict Lake cirque basin. Laurel Mountain is the high peak to the right (north), beyond the Tioga mo-

Table 2-3. Glacial Sequences of the Eastern Sierra Nevada

North American Sequence	Blackwelder (1931)	(modified from Lipshie, 2001) Sharp and Birman (1963)	Lipshie (2001)	Absolute Age (years B.P.)
			Matthes	0-700
Post-Wisconsin			Recess Peak	2000-4000
			Hilgard	9000-10,500
Late Wisconsin	Tioga	Tioga	Tioga	12,000-25,000
Early Wisconsin	Tahoe	Tenaya-Tahoe	Tenaya-Tahoe	45,000-75,000
Illinoian	-	Mono Basin	Mono Basin	About 130,000
Kansan	Sherwin	Sherwin	Sherwin	About 800,000
Nebraskan	McGee	McGee	McGee	1,500,000-2,500,000

rairie deposits. The Convict Lake basin lies between the two mountain masses. The basin is an excellent example of a U-shaped glacial valley.

As is the case with many glacial deposits throughout the Sierras, there has been significant disagreement as to the ages of the Convict Creek moraines. Blackwelder was the first to study and interpret age relationships of these deposits when he visited the area in the 1920s. Subsequently, others have examined the Convict Creek moraines with a variety of interpretations as to their age relationships (see Lipshie, 2001 for a discussion of these interpretations). **RETRACE** to Hwy 395.

73.6 (0.1) **TURN LEFT** (north) on 395.

75.0 (1.4) **TURN RIGHT** on Hot Creek/Fish Hatchery Road, the paved road to Hot Creek.

75.2 (0.2) Pass a right turn to Mammoth/June Lake airport (Airport Road).

75.5 (0.3) **TURN RIGHT** on Hot Creek Hatchery Road.

75.9 (0.4) Pass the fish hatchery, where they use heated water from Hot Creek to raise the fish. The Hot Creek Fish Hatchery is down the gravel road to the left. While not the most geologic of stops, children really love it! Near the hatchery, geothermal springs mix with the water from Hot Creek. The result is a series of ponds that maintain year-round temperature of 55 to 63°F, an ideal environment for incubating trout eggs and raising fish. The hatchery, run by the California Department of Fish and Game, has been in operation since the early 1930s. Children love to feed the fish (costs a quarter), but stick your hand in the water at your own risk!

76.1 (0.2) We are driving over a late

Quaternary fault scarp that had minor additional offset in 1980. This linear, low scarp cutting late Pleistocene basalt is one of several pre-existing faults that ruptured at the surface during the 1980 earthquake swarm; this one possibly is a splay of the Hilton Creek fault. Extension cracks appeared on Hot Creek

Road trending in a northwest-southeast direction at the base of the southwest-facing fault scarp. Extension varied from 0.5 to 3 inches, with individual cracks reaching 15 to 20 feet in length. Vertical displacement ranged from 6 inches to about one foot. Field mapping and fault plane solutions suggest extension, but the actual displacement mechanism is complex, possibly involving magma movement.

76.7 (0.6) Road bears left at T-intersection and pavement ends.

78.2 (1.5) **TURN LEFT** into the Hot Creek Geologic Site parking area and **PARK**.

STOP 2-6. Hot Creek (37°39'38"N, 118°49'40"W). From the interpretive overlook, note the steam rising from fumaroles and hot springs along the creek (Fig. 2-7). Hot springs also discharge directly into Hot Creek near the remains of the bridge that formerly spanned the creek. The mingling of hot spring water with snow-melt fed stream water produces extreme temperature gradients in the creek. The wide range of



Figure 2-7. View looking to the west along Hot Creek. Light-colored areas have been hydrothermally altered by geothermal waters. [Source: Farrar, Evans, Venezky, Hurwitz and Oliver, 2007]

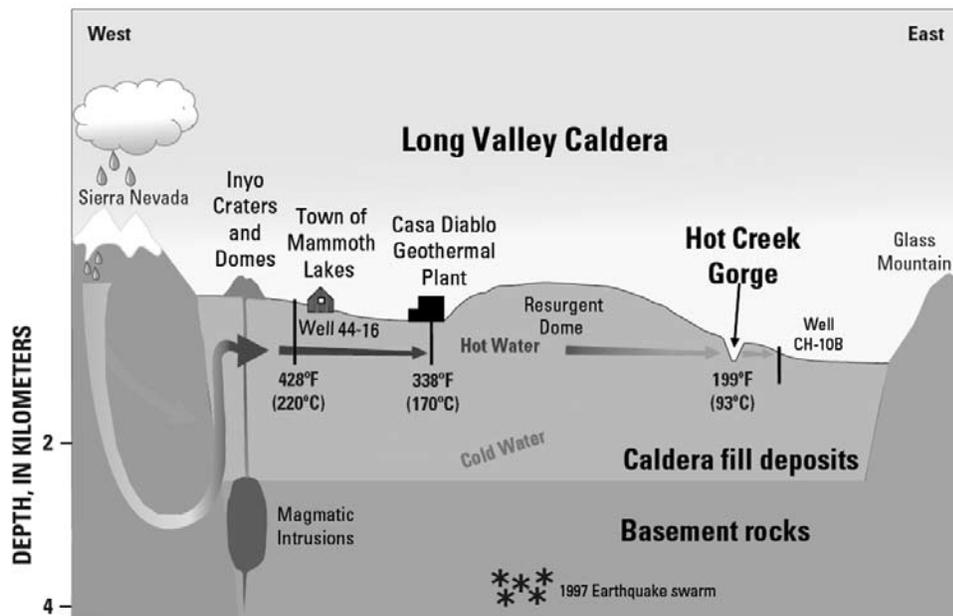


Figure 2-8. Cartoon cross section showing hydrologic recharge in the Long Valley Caldera [Source: Farrar, Evans, Venezky, Hurwitz and Oliver, 2007].

temperatures has made this area popular for swimming year-round so the Forest Service constructed change rooms for visitors. Unfortunately, in 2006 Hot Creek experienced an increase in geothermal activity, intermittently spurting hot, sediment-laden water as high as 6 feet above the stream surface. At times this geyser activity is vigorous enough to produce “popping” sounds audible from hundreds of feet away. The activity usually lasts a few seconds and occurs at irregular intervals, with several minutes between eruptions. The unpredictability of this hazardous activity led the U.S. Forest Service to close parts of the Hot Creek Geologic Site in June 2006. As of 2008 it remained closed to public access.

Note the altered rhyolite in the gorge. Hydrothermal activity has kaolinized and opalized the rock producing the white, bleached appearance (we will see a better example at the next stop). The rhyolite has been dated 300,000 years. The northeast trend of Hot Creek is consistent with that of the Hilton Creek fault, so many geologists have theorized the main fault or a branch are the conduit for hydrothermal waters. This hypothesis became more difficult to defend when recent strong earthquakes on the Hilton Creek fault (1998) had little impact on the thermal regime of Hot Creek. As you walk down the path to the creek also note the hummocky terrain on the northeast side of Hot Creek, reminiscent of the Owens River gorge.

The Hot Creek Geologic Site lies within the Long Valley caldera, created 760,000 years ago during the mas-

sive Bishop Tuff eruption. A subsequent eruption produced the 300,000 year-old Hot Creek rhyolite flow, known locally as Doe Ridge. The toe of that flow entered a lake. After the lake receded, Hot Creek eroded a steep-sided gorge through the toe of the solidified lava. The alteration to kaolinite/perlite and silicification visible along the trail to Hot Creek may have occurred when the rhyolite flow interacted with lake water or as a consequence of subsequent hydrothermal activity.

The geothermal system in Long Valley Caldera is recharged primarily by snowmelt around the western and southern rims of the caldera (Fig. 2-8). The meltwater infiltrates to depth where it is heated to at least 220°C by magma underlying the Inyo Craters and Domes, 10 miles west of Hot Creek. The heated water rises along steeply inclined fractures to depths of 0.3–1.25 miles. It then flows eastward to discharge at the surface along Hot Creek and around Crowley Lake. The water temperature decreases eastward because of conductive heat loss and mixing with cold water. In the springs near Crowley Lake, temperatures are at only about 50°C (Farrar, et al., 2007).

The springs in Hot Creek all emerge along a stream section between two faults and discharge a total of about 8.5 cubic feet per second of hot water (Farrar, et al., 2007). This water flow represents nearly 70 percent of the total heat discharged by all thermal springs in the Long Valley caldera. The thermal springs farther east all discharge less water and at lower temperatures.

Some of the current hot springs appeared suddenly in 1973. At least five hot springs formed, with the two largest starting as geysers that spouted water 10 feet into the air. Within weeks geyser activity had ceased, but the hot springs remained. The origin of these hot springs remains unclear, but it was noted that they appeared within hours of a relatively small ($M=3.5$)

earthquake 25 miles southeast of Hot Creek. Presumably, seismic activity altered the subsurface plumbing system, giving rise to the springs. Prior to the small earthquake, heated water was trapped below an impermeable horizon. The seismic event breached the impermeable strata and superheated water and steam rose rapidly, initiating geysers at the surface. After the initial pulse of superheated water reached the surface, the heat flux decreased and the geysers became hot springs. The increased rate of thermal activity continued along Hot Creek through the period of moderate seismicity in the 1980s. It began to wane in the 90s and continued at low ebb until 2006.



Figure 2-9. Kaolinite “ore” stockpile at the Huntley Mine open pit. [D. R. Jessey photo]

Geysering recommenced in May 2006.

However, at that time seismic activity and ground deformation were at the lowest levels seen in many years, and the reason for the increase in thermal activity was unclear. The change seems to be related to increased temperature in the shallow thermal ground water that supplies the springs (Farrar, et al., 2007). What caused the temperature increase in the aquifer is not known, but it may be a delayed response to an earthquake swarm in 1997 that could have opened new flow pathways for hot water. An alternative explanation suggests that increased pressure in the aquifer led to an increased flow rate at the springs. In 2006, following a winter of heavy snow, there was abundant snowmelt that would have increased pressure in the aquifer.

Leave the parking area by the east exit and **CONTINUE EAST** on Hot Creek Road.

79.4 (1.2) **BEAR LEFT** at the intersection with Whitmore Tubs Road.

80.0 (0.6) **TURN LEFT** on Owens River Road.

80.1 (0.1) Cross Hot Creek.

80.6 (0.6) **TURN LEFT** on Antelope Springs Road (3S05) toward Little Antelope Valley.

81.3 (0.6) **PARK**.

STOP 2-7. The low ridge to the right is capped by siliceous sinter deposits from fossil hot springs activity. This is a good place to view silicified plant remains. The fossilized plants are mostly a marshland flora.

You should be able to observe some silicified stalks of marsh plants that have weathered out of the sinter.

83.1 (1.8) Continue past a right turn.

83.5 (0.4) The road passes through Pleistocene lake sediments.

84.0 (0.5) Five-way intersection. **TURN 45° RIGHT** (not 90° right). Forest Service route 3S05 goes straight through the intersection, heading westward. **CAUTION:** road is clayey and vehicles can easily get stuck.

84.1 (0.1) **STOP 2-8. Huntley Kaolinite Mine** (N 37°41'18.2”; W 118°51'59.9”). The Huntley Kaolinite Mine is an active mine currently owned by Standard Industrial Minerals of Bishop, CA (Fig. 2-9). The mine does not operate on weekends, but mining equipment is parked at the site. Due to vandalism that occurred in 2005–2006 the mine property is now posted with prominent No Trespassing signs. However, the tailing piles can be accessed from Antelope Springs Road without entering the mine site. The company has been quite accommodating to geologists in the past. If you plan to enter the mine property, contact the company and let them know when you would like to visit and get their permission. If you visit the pit, respect their equipment and other property.

The kaolinite has been formed by hydrothermal alteration of Pleistocene lakebed sediments and the underlying rhyolite. The degree of alteration is a function of porosity, the more porous tuffaceous lakebeds being more pervasively altered than the rhyolite

(Cleveland, 1962). The rhyolite has been dated at 300,000 years making it correlative with that at Hot Creek. The alteration appears to be controlled by a north-trending fault system that created a graben. Kaolinization is best developed along the east side of the graben. Relict bedding from sedimentary layers and flow banding of the rhyolite can be seen in large boulders from the stockpiles and in outcrop in the pit. Opal veins are common, while alunite is rare. The genesis of the kaolinite and alunite will be discussed at the next stop, but both are common alteration products of acid-sulfate geothermal systems.



Figure 2-10. Outcrop of blue chert as seen from Forest Service road 3S07. View looking to the southwest. [D. R. Jessey photo]

The Huntley Mine began production of kaolinite clay in 1952, and in its first seven years of operation the mine produced 16,300 tons of clay (Cleveland, 1962). The mine, formerly owned by Huntley Industrial Minerals of Bishop, was acquired by Standard Slag of Warren, Ohio, in 1962. In 1987, the mine was resold to a local group in Bishop. The kaolinite mine has enough reserves to last until 2055 at the current rate of production (Richard Harmon, oral communication, 2005). Kaolinite is hauled by truck from here to the company mill at Laws, near Bishop, where it is ground into powder and shipped to customers. Kaolinite is important in the production of ceramics, porcelain, and glossy paper.

84.2 (0.1) Leave the Huntley Mine and **RETRACE** south to the 5-way intersection. Go straight (south) at intersection. This road is Forest Service route 3S07. Keep your speed up on this road to avoid getting stuck in the clay.

85.5 (1.3) **TURN LEFT** on an unimproved dirt road (look for the rusted water storage tank).

85.6 (0.1) **STOP 2-9. Blue Chert Mine** (N 37° 40'19.2"; W 118°51'23.6"). Drive up the road a few hundred feet and **PARK** on the small hill near the outcrop of blue chert.

This small hill appears to be the top of a fossil hot springs system (Fig. 2-10). Numerous other small knolls occur throughout this area. Each hill represents the locus of a hot spring or fumarole. As hot water rises in the vent conduit and cools, it deposits

microcrystalline silica along the walls of the fissure. This often seals the vent system, which remains dormant for decades until seismic activity reopens the fissures. When that happens, the trapped geothermal waters rise rapidly to the surface and flash to steam. The explosiveness of the erupting geyser often overcomes the tensile strength of the rock, shattering it and creating a breccia pipe. The rapid temperature drop deposits a silica sinter blanket around the vent opening.

The chert from this locality has been dated at approximately 300,000 years B.P. This is consistent with ages for the rhyolite at the Huntley Kaolinite Mine and that at Hot Creek. Presumably, the chert formed during a major period of hot springs activity that developed following volcanism associated with rhyolite emplacement. One intriguing question is the source of the blue color. It appears restricted to this hill, and has not been explained adequately. The source of all color in cherts, according to many sources, is impurities incorporated during chemical precipitation. The most common impurity is amorphous iron oxide resulting in the red variety of chert known as jasper. The nature of impurities causing other colors of chert is uncertain and the subject of ongoing research.

Vista Gold Inc. of Littleton, Colorado holds a lease on the Blue Chert property. Through a purchase agreement with Standard Industrial Minerals, owner of the Huntley Kaolinite Mine, they currently control approximately 1800 acres. An extensive drilling project was completed in 1998 which outlined 68,000,000 tons of ore grading 0.018 oz/ton gold. The company green-lighted property development in 1998, but en-

vironmental opposition has delayed plans. During a visit in 2008 no evidence of recent activity was noted.

Mine Development Associates (2008) reports that the gold deposit is underlain by rock units related to caldera formation and subsequent resurgence. Lithologies include volcanoclastic siltstones and sandstones deposited in a lacustrine setting within the caldera, debris flows with local intercalated silica sinter, and rhyolite flows and dikes. All lithologies have been altered and/or mineralized to various degrees.

The north-south trending Hilton Creek fault defines the eastern limit of the resurgent dome within the central part of the Long Valley caldera and extends beyond the caldera to the south. Vista believes that this fault system controls the distribution of gold mineralization at the Blue Chert deposit.

The epithermal (hot springs) gold and silver mineralization falls within the low sulfidation (quartz-adularia) deposit type. Several favorable exploration targets, termed the North, Central, South, Southeast, and Hilton Creek zones, are mineralized with low grades of gold and silver along north-south trending zones up to 8,000 ft in length with widths ranging from 500 ft to 1500 ft. The tabular bodies are generally flat-lying or have shallow easterly dips. Mineralized zones are typically from 50 to 200 ft thick and exposed at (or very near) the surface. These zones correlate with areas of intense argillic alteration and/or silicification. The predominant clay mineral is kaolinite, while the silicification varies from chalcedony to amethystine quartz or opal. Multiple periods of brecciation and silicification are evidenced by cross-cutting veinlets and silicified breccia zones.

Mineralization consists of poorly crystalline pyrite, often with framboidal texture, and lesser euhedral pyrite. Investigations have shown submicroscopic gold occurs within the framboidal pyrite. Where gold grains have been observed, the grains are small (1 to 6 microns) and have low amounts of contained silver. A significant portion of the gold resource is present in material which has been at least partly oxidized. The pyrite is altered to iron oxides (goethite), releasing the gold. See Jessey (this volume), for discussion of ore genesis and a model for ore deposition.

This fossil hot spring hill is in the lower reaches of a Jeffrey pine (*Pinus jeffreyi*) forest covering much of the Long Valley resurgent dome. Individual pines can be seen in sagebrush scrub of Long Valley south and

east to Benton Crossing Road. Jeffrey pine extends from Baja California to the Siskiyou Mountains in northern California, where it is found on ultramafic-derived soils. Typically the Jeffrey pine is found on well drained soils at elevations from 6000 ft. to 8000 feet. At higher elevations (Mammoth Lakes), Jeffrey pine is replaced by California red fir (*Abies magnifica*) and then Sierra lodgepole pine (*Pinus contorta* ssp. *murrayana*). Highest elevation trees are the whitebark pine (*Pinus albicaulis*), which form krummholtz in windswept areas. Logging transported timber by rail to Bodie, where it became mine timbers, stamp mills, and siding for bars and brothels.

Great Basin sagebrush to the east gives way to seasonally wet alkaline meadows in Long Valley. These wet "saltgrass meadows" are fed by streams slowly meandering to the Owens River. Saltgrass (*Distichlis spicata*) is dominant, while other important halophytic plants include clustered field sedge (*Carex praegracilis*), arrow grass (*Triglochin maritima*), and clustered goldenweed (*Pyrrocoma racemosa*). The closest published flora to Long Valley is northeast at Glass Mountain (el. 11,123 ft) towering 4,400 feet above the Owens River in the floor of Long Valley (Honer, 2001; Michael Honer of Rancho Santa Ana Botanic Garden) This study samples saltgrass meadows, sagebrush series, subalpine forests and meadows, and arid alpine peaks and ridges.

RETRACE via Antelope Springs Road (3S05) to Highway 395.

87.0 (.1.4) **TURN RIGHT** at Huntley Mine and 5-way intersection.

87.9 (0.9) Pass a left turn; continue on graded road.

90.4 (2.5) **TURN RIGHT** on Owens River Road.

90.9 (0.5) Cross Hot Creek.

91.0 (0.1) **TURN RIGHT** on Hot Creek Hatchery Road toward Whitmore Tubs Road.

91.6 (0.6) **BEAR LEFT** on 2507 toward Whitmore Hot Springs. Avoid Hot Creek Road to the right.

92.9 (1.3) **STOP 2-10. Regroup** at View Stop. From the rise, view 12:00 to 1:00 of moraines at Convict Creek; 11:00 moraines in McGee Creek; 10:00 moraines in Rock Creek.

93.1 (0.2) Continue past a left turn.

94.4 (1.3) Stop sign at junction of paved Benton Crossing Road (2584). Whitmore Hot Springs are

ahead on the left.

95.6 (1.2) **Stop** at northbound Hwy 395; watch for traffic, cross, and **TURN LEFT** onto southbound Hwy 395.

96.1 (0.5) Pass Mt. Morrison Road; view left of Crowley Lake.

96.9 (0.8) Glacial moraines in McGee Canyon are ahead at 1:00; moraines at Rock Creek are at 10:00.

97.9 (1.9) Continue past McGee Creek Road.

98.0 (0.1) **Stop** at Highway 395. **TURN RIGHT** (south).

99.8 (1.8) Scenic turnout; Crowley Lake is to the east.

100.4 (0.6) Continue past Crowley Lake exit.

100.8 (0.4) Look for unusual differential weathering in bedded volcanic tuffs.

104.4 (3.6) Tom's Place.

107.5 (3.1) View moraines at 3:00 in Rock Creek Canyon.

109.8 (2.3) Vista on the right. This trip has traversed the southern margins of the Long Valley caldera and we are heading south into Round Valley and Owens Valley.

111.2 (1.4) View at 12:00 of Owens River Gorge and LADWP pipeline.

111.7 (0.5) View at 2:00 of U-shaped Pine Creek and lateral moraines extending into the valley. Note brown debris cone at the east end of the moraines.

115.1 (3.4) Enter Inyo County.

116.5 (1.4) Pass turnout to lower gorge. Enter Round Valley.

118.5 (2.0) Continue past Pine Creek Road. The Tungsten Hills are ahead at 12:00.

126.9 (8.4) Traffic light at Barlow Lane. Bishop city limits.

128.3 (1.4) Junction with Highway 6. **BEAR RIGHT** (south) on Hwy 395. Check your gas tanks.

129.8 (1.5) Line Street in Bishop.

End Day 2

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Day 3

Theme: Records of Life in Owens Valley

with contributions from S. Baltzer, D. R. Berry, G. T. Jefferson, Tom Schweich, and Janet Westbrook

What We Will See: Our route runs south along the eastern margin of Owens Valley through the southwestern Basin and Range Province toward the Mojave Desert Province. We pass outcrops of Proterozoic, Paleozoic, and Mesozoic rocks that contain a fossil record of organisms spanning more than 500 million years. We will revisit basalt flows (Day 1) that range in age from the middle Miocene to the mid Holocene and help define the timing of past surfaces and landscapes. Vertebrate fossils from Pliocene and Pleistocene lake and fluvial deposits help define the time when basins first opened and the span of time during which sediments filled those basins. The history of base metal mining and “mining” for water resources—cold and hot—is relevant to the history and development of the eastern Sierra Nevada, Owens Valley, and southern California, including the thirsty metropolis of Los Angeles.

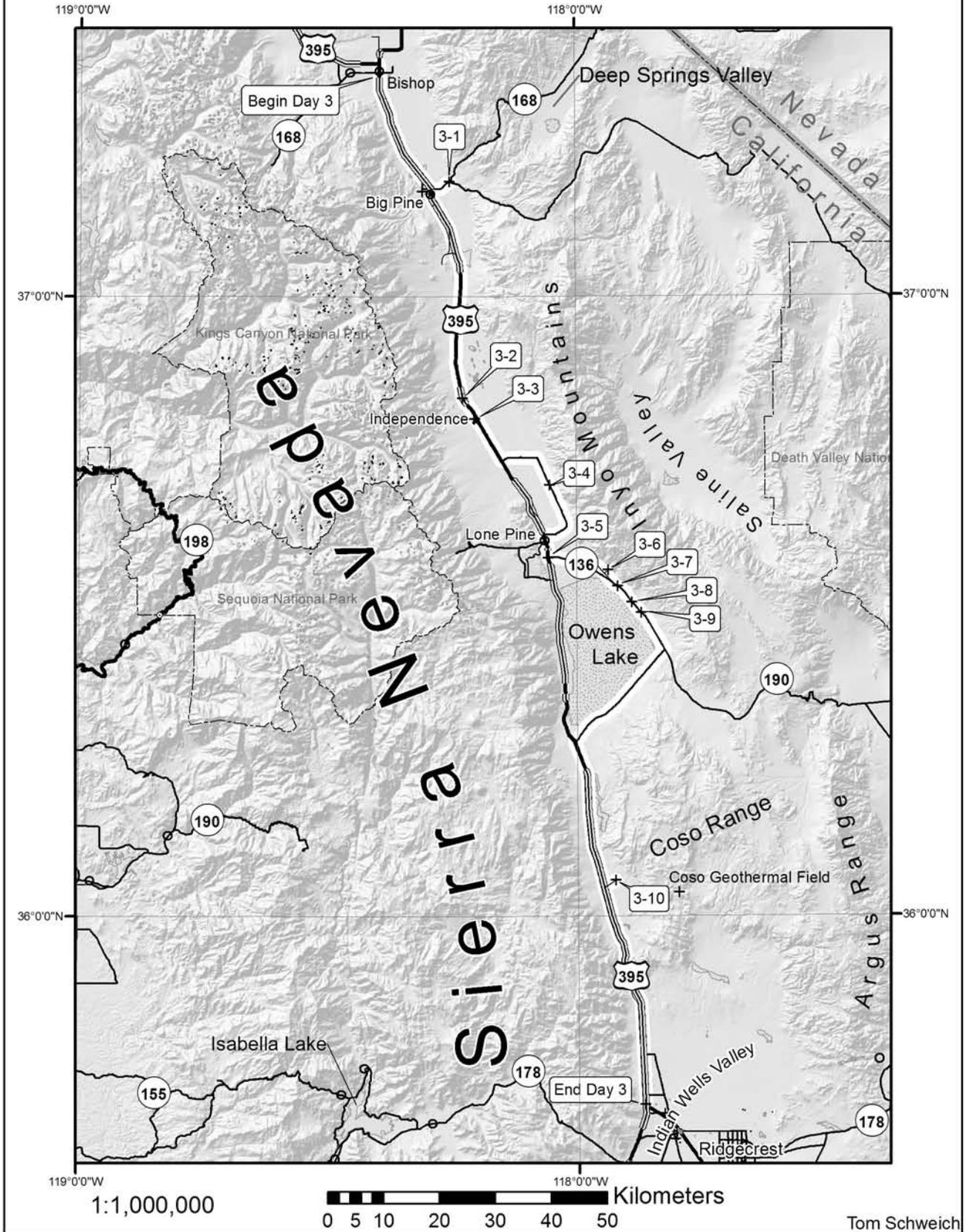
(0.0) (0.0) **Convene** in Bishop at the intersection of Hwy 395 and Hwy 6. The White Mountain Visitor Center is one-tenth mile south on Hwy 395. To the northeast, east of Hwy 6 on Silver Canyon Road, Laws is at the site of the 1863 town of Owensville (Hensher, this volume). The Laws Railroad Museum and Historical Village sits on the Carson and Colorado (C & C) Railroad (later Nevada & California (N & C) Railroad, which was in turn acquired in 1907–10 by Southern Pacific to connect Mojave with Owenyo). The N & C also connected to rails in the Tonopah–Goldfield–Rhyolite–Beatty area of western Nevada (Myrick, 1992). Proceed south on Hwy 395.

0.8 (0.8) Continue past Line Street in Bishop.

4.7 (3.9) Weigh station. View at 10:00 to Black and Marble canyons in the southern White Mountains. Notice the U-shaped valley with moraines.

8.1 (3.4) Continue past Keough Hot Springs Road. The Owens Valley Radio Observatory (OVRO) astronomy observatory to the southwest is operated by California Institute of Technology. Its original instrumentation included an array of six radio telescopes, a 40-m telescope and the Solar Array consisting of two 27-m telescopes. Only 3 dishes are here; the rest have been moved up to Cedar Flat at the head of Westgard Pass to form a Very Long Millimeter-Wavelength Ar-

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ray. The big 40m dish is now used by graduate students on projects, and the two 27m dishes usually track the sun. (CalTech Astronomy, n.d.). Look southeast at 10:00 past the OVRO antennae to two fault scarps that cut Quaternary gravels.

13.8 (5.7) Continue past Reynolds Road on the right. Slow entering Big Pine.

15.0 (1.2) **TURN LEFT** (east) onto Highway 168 and proceed across the Owens River toward Goldfield, Tonopah, and northern Death Valley. CA 168 heads east to Westgard Pass and the ancient bristlecone pine forest.



Figure 3-1. The Owens River. [J. Westbrook photo]

16.5 (1.5) Cross the Owens River.

16.9 (0.4) Continue past Leighton Lane on left.

17.3 (0.4) **PULL RIGHT** and **PARK** at the junction of Hwy 168 (to Tonopah) and the Wacouba Canyon–Death Valley Road. Westgard Pass is about nine miles ahead on Hwy 168 along Deep Spring Valley Road.

STOP 3-1. The biostratigraphic sequence to the east along Hwy 168 near Westgard Pass (Baltzer and Berry, this volume; Berry, this volume) contains some of the oldest rock units in western North America. These Precambrian and Cambrian rock units contain an abundant Lower to Middle Cambrian fauna, including Archeocyathids, trilobites, *Scolithus* (worm burrows) and *Girvanella* algal structures. The trilobite *Fallotaspis* in the shales of the Campito Formation marks the Cambrian-Precambrian boundary (Seiple, 1984). These trilobites are older than those from the Carrara Shale at Emigrant Pass (Palmer and Halley, 1979) east of Shoshone and those in the Latham Shale of the southern Marble Mountains (Mount, 1980). Because of the abundance of fossil fauna and tectonic history of the region, the Westgard Pass vicinity is a favorite location for geology summer field camps.

Look southeast at the gray, tuffaceous, well-bedded siltstone and sandstone that reach elevation 6,400 ft. These sediments from Lake Waucobi (Strand, 1967) have produced fossil horse that suggests a Pleistocene age (*Equus* sp.; Schultz, 1938; LACM (CIT) 265; Jefferson, 2008a). Later work suggests that a portion of the section is late Pliocene (2.3 Ma; Bachman, 1974,

1978; Sharp and Glazner, 1997) and documents the initial separation of the Inyo Mountains from the Sierra Nevada range. We are parked at el. 4,000 ft near the 100-year-old C&C/N&C railroad bed and former townsite of Zurich. **RETRACE** to Hwy 395 and Big Pine.

19.6 (2.3) **Stop** at Hwy 395. Watch for cross traffic. **TURN LEFT** (south) on Hwy 395 toward Independence. Continue past Baker Creek Campground Road.

20.1 (0.5) Continue past Crocker Road to the Big Pine Creek recreation area.

21.7 (1.6) Continue past Stewart Lane on the left (east).

24.7 (3.0) Continue past North Fish Springs Road on the right (west).

27.3 (2.6) Continue past South Fish Springs Road on the right (west).

28.5 (1.2) Continue past Elna Road. Watch for tule elk! Osprey (fishing hawks) are now common at the three CDFG fish hatcheries in the Owens Valley. They formerly dove into aqueduct raceways to fetch dinner. CDFG erected wires to foil the birds. Osprey now fish the outflow to catch escapees.

Tule elk are native to the tule marshes in the southern San Joaquin Valley. Hunting to feed miners reduced their numbers almost to zero by 1875. A small refuge was set up near Tipton, west of I-5, and the elk



Figure 3-2. Mobile home damaged by the Oak Creek debris flow. [Reynolds photo]

thrived without hunting. Herds introduced by Fish and Wildlife into Owens Valley did well, so more animals were sent. There are four major herds, the largest in the alfalfa fields west of Tinemaha Reservoir. When Hwy 395 was widened, the elk stopped migrating across the road. They spend summer and winter in the volcanic fields to the west, but gather in alfalfa fields in August when males gather harems. Their bugling can be heard from highway “wildlife viewing areas.”

32.0 (3.5) Continue past Aberdeen/Taboose Road. Mt. Whitney (el. 14,494 ft) and Mt. Williamson (el. 14,375 ft) are ahead. The intake to the LADWP aqueduct is to the east along Goodale Road. DWP was ordered by courts to “rewater” the Owens River, so they modified the aqueduct structure and the diversion dam and in December, 2007, allowed 40 cfs to flow down the natural Owens River channel. Recovery of the riparian system has been faster than imagined. The river is now navigable by canoe, has a good fishery, and supports bald eagles and lively crawdad ponds (Fig. 3-1).

37.6 (5.6) Continue past Saw Mill Creek/Black Rock Road.

42.8 (5.2) North Fort Road. We

are approaching the site of Camp Independence.

Prepare for a left turn: **move to the left lane.**

43.4 (0.6) Continue past Miller Street.

43.6 (0.3) **TURN LEFT** (east) on Schabbell Lane. Note fresh mud flow debris from Oak Creek crossing Schabbell Lane. Mt. Whitney Fish Hatchery, to the west along the north Fork of Oak Creek, was built in Tudor style in 1917 as the most lavish hatchery in the Owens Valley. Over 3,000 tons of granite collected from the site were intricately pieced together to construct the main building. After whirling disease was found on the premises in the late 1980s, the hatchery stopped raising fish and was nearly closed in 1996. Only local outcry kept it open. Until 2008, four million trout eggs were incubated annually and then shipped to other hatcheries that raised the fry for release into the lakes and streams of the Sierra Nevada. Schabbell Lane bears north.

43.8 (0.2) Historical marker for Fort Independence at the intersection with Oak Creek Road. Camp Independence was established on July 4, 1862 by Lt. Col. George S. Evans of the 2nd California Calvary Volunteers. The post, never an official fort, was to provide protection from Indian marauders for the area’s miners. It was temporarily abandoned in 1864, reoc-



Figure 3-3. The Mary Austin home in Independence, CA. [Reynolds photo]

cupied in 1865, and finally abandoned on July 5, 1877 (California State Military Museum, 2009). The town of Independence was named after the post. Note the mud flow-damaged mobile home at 10:00 (Fig. 3-2).

44.0 (0.2) **TURN LEFT** on Miller Street.

44.1 (0.1) Intersection with Dusty Lane. Proceed west on Miller Lane.

44.2 (0.1) Junction with Willow Road. Pull over and **PARK** before reaching Highway 395.

STOP 3-2. Look west-northwest at debris flow which came down north fork of Oak Creek (Wagner and others, this volume).

44.3 (0.1) Watch for cross traffic and **TURN LEFT** (south) on Highway 395 and continue past Schabbel Lane/Fish Hatchery Road.

45.3 (1.0) Continue past Fort Road South.

46.7 (1.4) Pass Dehy Park in Independence on the right (west). The locomotive is from the C & C narrow gauge railroad.

46.9 (0.2) The Inyo County Courthouse is on the left (east).

47.0 (0.1) **TURN RIGHT** (west) on Market Street.

47.1 (0.1) The Mary Austin house is on the corner of Market and Webster (Fig. 3-3). Mary Hunter Austin (1868–1934) was the author of “Land of Little Rain,” a 1903 classic in American nature writing that describes the lands and inhabitants of the Mojave Desert and Owens Valley (Austin, 1903; Crowley, 1937; Hoyer, 2003). 47.2 (0.1) **TURN RIGHT** on Grant and **LEFT** into the Eastern California Museum parking lot.

STOP 3-3. The Eastern California Museum has a collection of eastern California minerals and mining equipment. It also has an excellent exhibit of Native American baskets and artifacts and reproductions of rock art. The life of Japanese-Americans interned at Manzanar is also exhibited. **RETRACE** to Hwy 395.

47.3 (0.1) **TURN LEFT** onto Market Street.

47.5 (0.2) **Stop** at Highway 395 (Edwards Street). **TURN RIGHT** (south).

47.9 (0.4) Continue past a left turn to Mazourka Canyon Road and Kearsarge. Mazourka Canyon is a favorite spot for geology field camps because it provides excellent exposures of the representative stratigraphic section and structural relationships in

the eastern Sierra Nevada. Cretaceous–Jurassic granites and granodiorites intrude Paleozoic sedimentary rocks. Dikes of the Independence dike swarm cut the sedimentary units and older intrusives.

The Kearsarge, Silver Sprout, and Virginia silver claims were established in 1864. The mining camp of Kearsarge attracted investors who formed the Kearsarge Mining Company; the mine was worked between 1854 and 1883 and again in the 1920s. Kearsarge, with a population of 1500 people in 1865, competed unsuccessfully with Independence to become the Inyo County seat. The town was virtually destroyed by an avalanche on March 1, 1867, although the mines continued to be worked. Survivors moved to Onion Valley to found the second site of Kearsarge (Speck, 2007). The name Kearsarge refers to the Civil War Union battleship that defeated the Confederate privateer “Alabama” (for which the Alabama Hills were named) (Sowall, 1985).

The road also leads to the Snow Caps Mine (gold, silver; Clark, 1970) and the Green Monster (Aubury, 1908) and Copper Queen copper mines. The latter contains azurite, bornite, calcite, chrysocolla, malachite, and pyrite in diopside/andradite host rock.

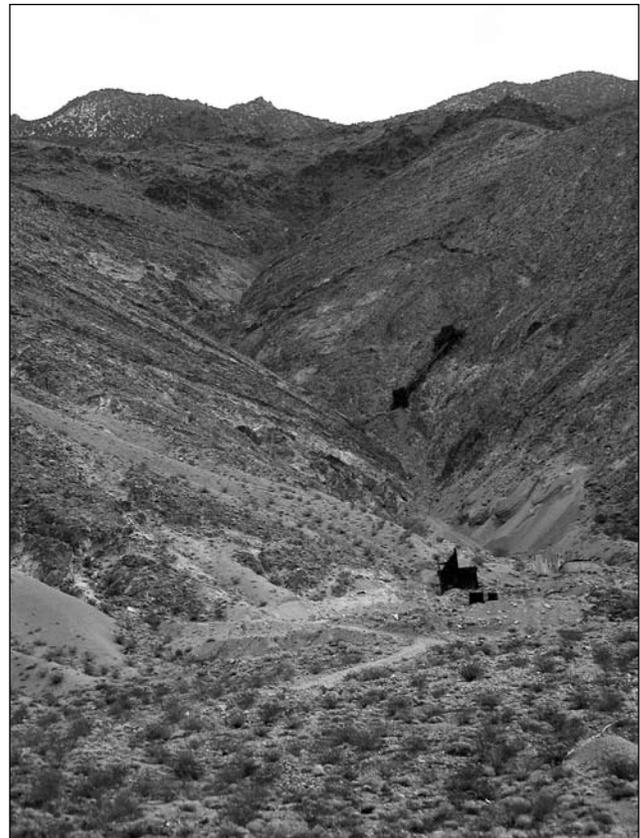


Figure 3-4. The Reward Mine. [R.E. Reynolds photo]

Table 3-1. Mineral assemblages at Reward and Brown Monster

Reward Mine		Brown Monster Mine	
Anglesite	Conicalcrite	Linarite	
Arsensumebite	Copper	Malachite	Calcite
Azurite	Corkite	Mimetite	Galena
Beudantite	Cuprite	Perite	Linarite
Brochantite	Dufite	Pyrrargyrite	Mimetite
Calcite	Fornacite	Pyrite	Quartz
Caledonite	Galena	Pyromorphite	
Cerussite	Goethite	Quartz	
'Chalcedony'	Gold	Schmiederite	
Chalcopyrite	Hedyphane	Tsumebite	
Chlorargyrite	Hemimorphite	Vanadinite	
Chrysocolla	Leadhillite	Wulfenite	

(Mindat.org, 2009; Pemberton, 1983)

52.6 (2.7) Prepare for a left turn. Watch for oncoming traffic. The Manzanar Relocation Center can be seen to the south.

52.8 (0.2) **TURN LEFT** (east) onto Manzanar/Reward Road. This road leads to the Union Mining District near the base of the Inyo Mountains where gold was discovered in the 1860s (Knopf, 1918). Chief producers were the Brown Monster and Reward mines (Fig. 3-4) with gold production estimated between 10,000 and 50,000 ounces. The country rock consists of Paleozoic shale, limestone, and conglomerate, as well as Jurassic-Triassic shale, tuff, and volcanic breccia. These rocks were intruded by Mesozoic granitoids generating lode vein deposits near the contact with the intrusions. Near-surface ore is highly oxidized consisting mainly of quartz and minor amounts of limonite, smithsonite, brochantite, linarite, chrysocolla, and wulfenite. Unoxidized ore consists of quartz with small amounts of pyrite, galena, sphalerite, and chalcopyrite.

53.1 (0.3) Cross Manzanar landing strip.



Figure 3-5. Stop 3-4: Deformed sediments of the Union Wash Formation. [S. Baltzer photo]

53.5 (0.4) Cross the LADWP aqueduct.

54.9 (1.4) Cross the Owens River.

56.5 (1.6) Continue past a power line road.

56.8 (0.3) **TURN RIGHT** (south) on the Lone Pine/Owenyo Road (former 1883 C&C/N&C railroad bed; Myrick, 1992). The site of Reward is two miles east-northeast. Pass the railroad siding of Manzanar on the west.

58.5 (1.7) The Eclipse irrigation ditch (dry) is on the west side of the road.

59.3 (0.8) Continue past a reverse turn to a mine.

60.2 (0.9) **PULL RIGHT** and **PARK** at Union Canyon Road.

STOP 3-4. The thick Jurassic marine section at Union Canyon (Fig. 3-5) contains ammonites (Berry, this volume; and see Batlzer, this volume: p. 68, fig. 5). The Eclipse ditch crosses the rail bed and joins the McIver Canal, and may have provided water for extensive agricultural fields east of Owenyo. Both flow south toward the mill at Swansea. **PROCEED SOUTH** on the Lone Pine/Owenyo Road.

61.4 (1.2) Pass the Owenyo townsite on the right (west); note abandoned fields to east. By 1883, the narrow gauge C&C/N&C Railroad reached Owenyo and ran south to Keeler. In 1907, the standard gauge Southern Pacific railroad extended the rail line from Mojave north along the west side of Owens Lake to Owenyo, and expanded service south of Keeler to soda plants on the southeast side of Owens Lake (Myrick, 1992).

63.3 (1.9) **Caution: bend in road.** Continue past the road leading east to French Spring.

64.6 (1.3) **BEAR RIGHT** (west) on Narrow Gauge Road toward Lone Pine. Long John Canyon and mines are to the east.

66.1 (1.5) Cross the Owens River.

66.5 (0.4) Slow as the road bends at ranch.

67.9 (1.4) **Stop** at Hwy 395. Watch for cross traffic and **TURN LEFT** (south). Enter Lone Pine.

68.6 (0.7) **Stoplight.** Continue past Whitney Portal Road.



Figure 3-6. Independence dikes. [S. Baltzer photo]

70.5 (1.9) **TURN LEFT** (east) on Highway 136 toward Keeler and Death Valley.

STOP 3-5. **PULL RIGHT** into the visitors center and regroup.

70.6 (0.1) **ENTER HWY 136** eastbound toward Death Valley.

72.8 (2.2) Pass through bedded Owens Lake sediments.

73.2 (0.4) Cross the Owens River.

73.6 (0.4) **Slow** for left turn ahead. Watch for on coming traffic.

73.8 (0.2) **TURN LEFT** (east) on Dolomite Loop Road.



Figure 3-7. The Dolomite quarry. [R. E. Reynolds photo]

75.9 (2.1) Bear southeast at the junction with Lone Pine/Owenyo Road (\former 1907 C&C/N&C railroad bed).

76.1 (0.2) Proceed 100 feet, **TURN LEFT** into turnout, and **PARK**.

STOP 3-6. Discuss the Paleozoic sedimentary section and Independence dikes (Fig. 3-6) that cut the face of the metamorphosed limestone in the dolomite quarry (Baltzer, this volume). The narrow gauge Dolomite Railroad hauled the white calcium/magnesium carbonate rock (Fig. 3-7) to the C&C/N&C (later S. P.) railroad. **PROCEED SOUTH** on Dolomite

Loop Road.

77.0 (0.9) Pass active mining at site of Dolomite. Proceed south on Dolomite Loop Road.

78.1 (1.1) **Stop** at HWY 136. Watch for cross traffic. **PROCEED SOUTH** on Hwy 136. View south-southeast (ahead) of Darwin volcanic field (Jessey, this volume).

79.0 (0.9) View west of the loading station for the Salt Tramway (Fig. 3-8). The salt ponds twelve miles north-northeast of Salt Lake (el. 1,060 ft) within the Saline Valley produced halite that was transported by tramway over the 8,800 foot crest of the Inyo Mountains to this loading station at el. 3,610 ft on the east shore of Owens Lake (Vargo, 2006; Mann, 2002).

80.1 (1.1) **PULL RIGHT** and **PARK** at Swansea historical site.

STOP 3-7. From 1870 to 1897, silver ore from the Cerro Gordo mines was shipped downhill to the Swansea smelter on the east side of Owens Lake. Timber was “roasted” with minimum oxygen at the Cottonwood charcoal kilns (Day 1, Stop 1-4) to produce charcoal, which was shipped on the Molly Stevens paddle-wheeled steam ship to Swansea for smelting lead-silver ore from the Cerro Gordo mines. The Molly Stevens picked up ingots, dropped them at Cartago, and then went north



Figure 3-8. Salt tram; foundations in foreground; pylon on horizon. [R. E. Reynolds photo]

to the Cottonwood kilns to get more charcoal. When the Molly Stevens burned, the Bessy Brady steam paddle-wheeler continued running silver ingots from Keeler and Swansea to Cartago (Day 1, MP-127.7). The smelter produced silver ingots faster than Remi Nedeau's nine mule teams could carry them on the route to Mojave. By 1873, over 30,000 ingots were stockpiled. Enterprising miners stacked the ingots, stretched canvas over them, and lived in shelters made of silver bars (Millspaugh, 1971). **PROCEED SOUTH** toward Keeler.

81.6 (1.5) **PULL RIGHT** and **PARK** at Owens Lake Desiccation Monument.

STOP 3-8. Owens Lake to the south has been the southern terminus for the Owens River for the last 800,000 years (Smith et al., 1993). Owens Lake surface water has been claimed and diverted for human use. Ground water is now tapped for the water supply. Arid-land surfaces that were previously stabilized by vegetation are increasingly susceptible to wind erosion, causing deflation and dust storms. Since desiccation, wind-blown dust and salt have created visibility problems for residents of Owens Valley and military research at China Lake in Indian Wells Valley (Reheis, in press).

During the early 1900s the lake fluctuated between about 20–45 feet deep and had an area of about 174 sq miles (280 km²). Water was first diverted from the Owens River to the City of Los Angeles in 1913, and by 1926 Owens Lake was dry.

The dry bed of Owens Lake has produced enormous amounts of windblown dust since the desiccation of the lake. The lake bed is probably the largest single source of dust (aerosol particles smaller than 10 microns) in the United States, by one estimate, 900,000–8,000,000 metric tons per year (Gill and Gillette, 1991). Assuming a lake-bed area of 280 km² and a density of 1.5–2.0 g/cm³ for lake sediment, erosion of this amount of dust would lower the lake bed by 0.2–1.5 centimeters per year. Micron-size dust is regulated by California

and United States agencies because dust particles are so small they can be inhaled into the human respiratory system to create health hazards. “One day [in 1995] in Keeler, particles surged to a nationwide record that was 23 times greater than a federal health standard allows” (Los Angeles Times, 12/17/96). Keeler residents are exposed to unhealthful levels 25 days a year. That situation occurs in Ridgecrest 10 days a year (Great Basin air agency). “When we see the white cloud headed through the pass, the ER and doctors’ offices fill up with people. It’s straightforward cause and effect” (Dr. Bruce Parker, physician at Ridgecrest Community Hospital). An additional health concern is inhalation of trace metals in dust from the lake bed

Table 3-2. Fossil faunas, Owens Lake playa.

Common Name	Generic Name
Bony fish	Osteichthyes
Minnow	<i>Siphateles bicolor snyderi</i> (Miller)
Owens sucker	<i>Catostomus fumeiventris</i> Miller*
Lesser scaup	<i>Athya affinis</i>
Meadowlark	<i>Sturnella neglecta</i>
Botta's gopher	<i>Thomomys bottae</i>
Wolf	<i>Canis lupus</i> **
Lion (medium-size)	Felidae
Antelope	<i>Antilocapra</i> sp. **
Horse	<i>Equus</i> sp.
Large camel	<i>Camelops</i> sp.
Bison	<i>Bison</i> sp.
Elephant	Proboscidea

(* Gerald R. Smith, p. c. to Reynolds, 3/09;

** R. Serrano, C. Shaw, p. c. to Reynolds, 3/09)

which has significant concentrations of arsenic (Reid et al., 1994).

In addition to health problems, Owens Lake dust affects visibility in nearby areas, including Sequoia, Kings Canyon, and Death Valley national parks. Dust plumes have been tracked to elevations of over 9,000 feet in the Ancient Bristlecone Pine Forest in the White Mountains, 62 miles (100 km) north of Owens Lake; dust-covered vegetation is common in the White-Inyo Range area (Reheis, and Kihl, 1995; Gill, 1996). Owens dust storms regularly cause suspension of military operations at China Lake Naval Weapons Center resulting in millions of dollars in losses (Reheis, in press).

82.9 (1.3) Keeler. Continue past Malone Street and “Cerro Gordo Road” on right. Keeler, a stop on the C & C Railroad in 1882, was named after J. M. Keeler, steamboat captain on Owens Lake.

Fossils in the collections of the University of California, Museum of Paleontology (UCMP), Los Angeles County Museum (Lone Pine Southeast-LACM 4691) and San Bernardino County Museum (Owens Lake East-SBCM 6.6.3-6.6.4) from the vicinity of Swansea and Keeler in Owens Lake playa are Rancholabrean and/or early Holocene in age. The fossil fish and scaup (duck) suggest fresh water and the meadowlark and mammals suggest grassland (Gust, 2003; Jefferson, 1989, 2008a).

83.3 (0.4) **PULL LEFT and PARK** at Cerro Gordo Road.

STOP 3-9. Cerro Gordo Mines are to the east of the Point of Historical Interest. The ore tram from the Cerro Gordo mines and townsite at el. 8,600 ft ran downslope to deposit ore at Keeler (el. 3,600 ft). This mining district is near the summit of the Inyo Range, 8 miles by mountain road from Keeler. Discovered early in the 1860s, production started in 1869, and reached between \$6.5 to \$20 million of silver and lead (Baltzer, this volume). Mississippian and Pennsylvanian limestones cut by intrusive diorite and monzonite dikes serve as host rocks for

the ore bodies. Initial sulfide ores were argentiferous galena and sphalerite. Tetrahedrite, chalcopyrite, and pyrite were prominent accessories in the primary sulfide vein. Oxidized lead ore in limestone consisted of lenses of cerussite to 6 feet diameter around a core of galena. Zinc carbonate ore (smithsonite) was encountered in 1911. Tenorite and the lead oxides bindheimite, caledonite, linarite, and leadhillite are also present. The Ignacio and Ventura mines in the district are southwest of the main mining camp at Cerro Gordo (Murdoch and Webb, 1966; Mindat.org, 2009; Baltzer, this volume). Return to vehicles, **PROCEED SOUTH.**

83.5 (0.2) A fault scarp to the left (east) cuts Pleistocene fanglomerates (Jennings, 1958).

84.9 (1.4) Sulfate Road to the right leads to soda evaporating ponds that produced sodium salts such as sodium carbonate (trona), sodium chloride (halite), and sodium sulfate (thenardite). Brine from wells near Cartago on the west side of Owens Lake was piped to the ponds for concentration by evaporation. The white surface of Owens Lake is a combination of salts: sodium, calcium and potassium chlorides, carbonates, and sulfates and borates. Hot summers allow blue-green algae to precipitate borax salts from the brine, giving the lake a red color. The lakebed is white during cool weather. The soda plant is now the LADWP headquarters for the dust mitigation project.

85.3 (0.4) The scarp to the left (10:00) appears to cut Pleistocene fanglomerates (Jennings, 1958). View southeast of Darwin volcanic field (Jessey, this vol-



Figure 3-9. Coso Formation and basement rocks. [S. Baltzer photo]

ume).

87.9 (2.6) **TURN RIGHT** (southwest) onto Highway 190 to Olancho at the Hwy 136/190 junction.

Hwy 190 leads southeast through the Darwin Volcanic field to the Darwin silver–lead–zinc mines, active since 1875 (Hall and Mackevett, 1958; Dunning and others, 2002), east across the Darwin Plateau and Panamint Valley before crossing Towne Pass and descending into Death Valley and Stovepipe Wells.

92.4 (4.5) Continue past a left turn to the Sierra Talc Mine. As we travel southwest, look south at the prominent fault-truncated fanglomerates on the north side of the Coso Mountains (Carver, 1975). Between here and Dirty Socks Spring, the elevation of Highway 190 is between 3,640 and 3,680 feet. Owens Lake's maximum highstand (Tahoe glaciation, 65 ka) lies at approximately 3,881 feet elevation south of the highway. Those shorelines are especially visible in the deeply incised pebble beaches and bars deposited along the fault scarp (Saint Amand et al., 1986). Tioga glacial shorelines (20–18 ka; Orme, 1995) lie at 3,753 feet elevation, also uphill, south of the highway. Holocene shorelines are north of the current position of the highway (Bates, this volume).

95.5 (3.1) View left (southeast) of Red Ridge, Vermillion Canyon, and the Coso Formation (Fig. 3-9).

97.5 (2.0) Continue past a right (north) turn to Dirty Socks Hot Springs, once a maintained Inyo County campground but abandoned in the 1970s. Many stories describe the origin of the name, but the consensus is that their pungent smell was the source (Sowall, 1985). The Coso Formation is exposed to the south from Vermillion Canyon and Red Ridge to farther south along Haiwee Reservoir. The Coso Formation contains fossils representing the late Blancan North American Land Mammal Age (NALMA) from 3.0–2.5 Ma. An absolute date of 3.0 Ma (K/Ar on volcanics) brackets the deposit. Fourteen species are known from sediments that apparently represent lacustrine, grassland and savannah communities (Schultz, 1937; Savage and Downs, 1954; Kurten and Anderson, 1980; Lundelius *et al.*, 1987; Jefferson, 2008a).

97.9 (0.4) The Bernard complex maintenance yard for LADWP equipment is at a former salt mining operation on the north side of the road.

100.3 (2.4) The Olancho Dunefield, south of Owens Lake, covers the overflow drainage of >2000 years.

Studies (Bates, this volume) indicate that the dune sands were derived from late Holocene deflation of local beach sediment. Shorelines of the final Holocene lake regression are located at the southern end of the dunefield below el. 3681 ft. The historic lake was 250 feet deep in 1872.

The plant community here is high desert Great Basin scrub, dominated by sagebrush (*Artemisia tridentata*). Creosote bush and Joshua trees are absent. Winters are cold and snowy and summers are mellow compared to the Mojave Desert.

101.8 (1.5) **Slow** for a curve and enter Olancho.

102.1 (0.3) **Stop** at Highway 395. Watch for traffic and **TURN LEFT**.

103.7 (1.6) Pass Walker Creek Road (west).

104.1 (0.4) Continue past Cactus Flat Road (east) and pass through Joshua trees. Cross the Los Angeles aqueduct one mile ahead where it enters Haiwee Reservoir.

105.2 (1.1) Cross the Southern Pacific railroad bed here and again at Rose Spring as it climbs from Indian Wells Valley through Rose Valley to the west side of Owens Lake (Myrick, 1992).

107.4 (2.2) Continue past Sage Flat Road (west).

107.9 (0.5) Continue past Lake Village Road; view east of the Coso Formation.

109.1 (1.2) Continue past Lakeview Road (east).

111.2 (2.1) Continue past Haiwee Canyon Road to Haiwee Dam (east). Pleistocene sediments referred to as the Haiwee Reservoir locality (LACM 3514, 4538) are Irvingtonian or Rancholabrean NALMA and contain mammoth (*Mammuthus* sp.; Gust, 2003).

Table 3-3. Fossils in the Blancan NALMA Coso Formation

Common Name	Generic Name
Bony fish	Osteichthyes
Cormorant	<i>Phalacrocorax</i> sp
Rabbit	<i>Hypolagus</i> sp. nr. <i>H. limnetus</i>
Rabbit	<i>Hypolagus</i> sp. (small)
Rabbit	<i>Pewelagus dawsoni</i>
Wood rat	<i>Neotoma (Paraneotoma) fossilis</i>
Meadow mouse	<i>Mimomys (Cosomys) primus</i> (type)
Bear	<i>Agriotherium</i> sp. cf. <i>A. sivalensis</i>
Bone-crushing dog	<i>Borophagus direptor (Hyaenognathus solus)</i> (type)
Mastodon	<i>Pliomastodon cosoensis</i> (type)
Horse	<i>Equus. (Dolichohippus) simplicidentis</i>
Peccary	<i>Platygonus</i> sp.
Camel (large)	Camelidae (large)
Llama	<i>Hemiauchenia</i> sp.

The Pliocene (Blancan NALMA) Coso Fm. is exposed high on the west-facing slope of the canyon (Duffield, and Bacon, 1981).

114.3 (3.1) Cross the Southern Pacific railroad bed west of Rose Spring and the Haiwee powerhouse.

114.7 (0.4) Dunmovin. Pliocene volcanic rocks to the east have been excavated for pumice. After 1910, places like Dunmovin and Gill's Oasis developed to serve increased traffic from improved roads and aqueduct construction.

117.4 (2.7) Prepare for a left turn into Coso Junction.

117.6 (0.2) **TURN LEFT** (east) into Coso Junction. Services include gas and market. Continue east on Gill Station/Coso Road.

117.9 (0.3) Continue past the Coso Geothermal Plant operation office on the left (north). Continue east.

119.6 (1.7) **PULL RIGHT** and **STOP** at Power Line Road.

STOP 3-10. View left (northeast) of white pumice mine cuts. The Sugarloaf perlite dome is to the southeast, along with cinder cones of the Coso Volcanic Field (Bruns, et al., this volume). The Coso Geothermal Field hosts a power-generating project in continuous operation for two decades. In 2002, four geothermal power plants were located in the main production area, constructed from 1987 through 1990. Turbine-generator sets are operated by steam; the geothermal system is heated by magma that has been migrating upward from a depth of 10 km to as shallow as 5 km. This near-surface system is evidenced by hot springs, mud pots, mud volcanoes, and fumaroles (Monastero, 2002; Unruh and others, 2002).

Coso Hot Springs is located in an area of rhyolitic volcanism with glassy perlite domes such as Sugarloaf Mountain, a well-known source of obsidian for Native American tool manufacture and trade. Sugarloaf Mountain has spherulites and lithophysae that contain high temperature silica (christobalite and tridymite) with opal, associated with orthoclase and iron silicates (fayalite, clinoferrosilite) and magnetite. Fumaroles at Coso Hot Spring are actively depositing cinnabar,

metacinnabar, sulphur, and opal, associated with the hydrous iron sulfates halotrichite, rhomboclase, and voltaite (Pemberton, 1983; Mindat.org, 2009).

“Boiling hot springs” first noted by a miner in 1860 developed into the Coso Hot Springs Resort by 1909. Coso waters, mud, and steam were claimed to cure complaints ranging from venereal disease to constipation. Bottled water promised “Volcanic Health and Beauty from Nature’s Great Laboratory ... a vitalizing blood builder which aids digestion, destroys invading bacteria and is especially recommended in cases of gastritis, stomach and intestinal catarrh ...” The resort operated until 1943 when the U.S. Navy began purchasing land for the China Lake Naval Ordnance Test Station. By 1947 Coso Hot Springs was permanently closed (Monastero, 2002; Unruh and others, 2002; Rogers, 2008). Return to Hwy 395.

125.3 (5.7) **Stop** at Hwy 395, watch for cross traffic. **TURN LEFT** (south) toward Little Lake. Red Hill cinder cone is visible to the south on the east side of Hwy 395.

130.1 (4.8) Continue past a left turn to Fossil Falls and campground located in basalt flows that have been eroded into unusual-shaped formations by last flows of the Owens River into Indian Wells Valley (Duffield and Smith, 1978).

131.8 (1.7) Continue past Little Lake Road to the right. The Southern Pacific railroad bed on the west side of Hwy 395 drops from Rose Valley into Indian Wells Valley.

134.6 (2.8) Continue past a right (west) turn to a power line road that provides access to exposures along the SNFF. (Reynolds, 2002, Stop 2-6, MP 134).

138.8 (4.2) Continue past a right turn to Nine Mile

Table 3-4. Rancholabrean fauna at China Lake

Lower vertebrates		Mammals	
Mojave chub	<i>Gila bicolor mohavensis</i>	Sloth	Edentata
Amphibian	Anura	Meadow mouse	<i>Microtus californicus</i>
Turtle/tortoise	Testudines	Coyote	<i>Canis</i> sp.
Grebe	<i>Aechmophorus</i> sp.	Saber Cat	<i>Smilodon</i> sp.
Cormorant	<i>Phalacrocorax</i> sp.	Lion	Felidae (large)
Mallard duck	<i>Anas</i> sp.	Mammoth	<i>Mammuthus</i> sp. cf. <i>M. columbi</i>
Scaup	<i>Aythya</i> sp.	Horse - small	<i>Equus</i> sp. cf. <i>E. conversidens</i>
Goose	<i>Branta</i> sp.	Horse - large	<i>Equus</i> sp. cf. <i>E. occidentalis</i>
Swan	<i>Cygnus</i> sp.	Camel	<i>Camelops</i> sp. cf. <i>C. hesternus</i>
Ruddy duck	<i>Oxyura</i> sp.	Llama	<i>Hemiauchenia</i> sp.
Golden eagle	<i>Aquila</i> sp. cf. <i>A. chrysaetos</i>	Deer	<i>Odocoileus</i> sp.
Bald eagle	<i>Haliaeetus</i> sp.	Ancient Bison	<i>Bison antiquus</i>
Coot	<i>Fulica</i> sp.	Bison - large	<i>Bison</i> sp. (large)
Crane	<i>Grus</i> sp.		

Canyon Road leading to Kennedy Meadows. China Lake is to the east. White sediments have produced the China Lake Local Faunule north of Ridgecrest (localities: LACM (CIT) 266, LACM 1543, 3569, 7013, 7262) dating to the middle to late Wisconsinan of the late Pleistocene Rancholabrean NALMA (Jefferson, 2008a) based on fauna and radiometric dates (U/Th $42,350 \pm 3,300$, ^{14}C $18,600 \pm 450$, $11,800 \pm 800$ yr BP). Fish, many birds, and meadow mice suggest a fresh lacustrine environment. Surrounding slopes were covered with grassland and copses of trees (VonHuene, 1971; Fortsch, 1978; Kurten and Anderson, 1980; McDonald, 1981; Davis and others, 1980, 1981a, 1981b; Davis, 1982; Agenbroad, 1984; Jefferson, 1989; Basgall, 2003; Moore and others, 2003; Jefferson, 2008a).

141.5 (2.7) Pearsonville. Founded in 1959 as a steakhouse, the Pearsons were apparently better mechanics than cooks and the steakhouse became a garage. Last services at the Kern County line.

147.6 (6.1) Leliter Road. Prepare for the junction of Highway 14 and 395.

148.6 (1.0) Junction of Highway 395 and Highway 14.

Visit the Maturango Museum Ridgecrest by taking Highway 178 east, then south on China Lake Boulevard (Hwy 178). The museum (founded in 1962) is east of Home Depot and north of Ridgecrest Boulevard. The museum emphasizes the cultural and natural history of the Indian Wells Valley and surrounding northern Mojave Desert. One sequence is a "Time Line" from the Pleistocene to the Navy installation; the other contains a rotating art/sculpture/photography exhibits. An extensive gift shop has books, petroglyph motif items, and jewelry.

Highway 395 leads southeast to Kramer junction on Hwy 58 (services) and is the best route to reach San Bernardino, Barstow, Las Vegas, and Kingman.

Highway 14 runs southwest to Mojave (services) and is preferable for reaching Los Angeles and Bakersfield.

End of Day 3

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Stolzite and jixianite from Darwin, Inyo County, California

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Introduction

Stolzite (PbWO_4) is not listed in Pemberton's 1983 Minerals of California. It is listed as occurring at Darwin, Inyo County in Murdoch and Webb's 1966 version where Tucker and Sampson (1941) were referenced. However, the validity was questioned based on studies by Hall and MacKevett (1963) that only showed the presence of scheelite (CaWO_4) at Darwin. Stolzite and scheelite are isostructural tetragonal minerals. Tucker and Sampson (1941) described the stolzite as occurring as bunches of crystals in the Thompson ore bodies. Hall and MacKevett (1963) also mentioned the occurrence of stolzite and referenced Tucker and Sampson (1941) along with personal communication with Dudley L. Davis (1955). Mr. Davis was a former resident geologist for the Anaconda Company, which controlled and worked the Darwin mines for a number of years. Murdoch and Webb's (1966) description of the stolzite crystals, attributed to Tucker and Sampson (1941), described it as being coated with tungstite, but there is no mention of tungstite in Tucker and Sampson (1941).



Figure 1. Gray stolzite after scheelite crystal (2 cm).



Figure 2. Brown jixianite after scheelite (2 cm).

Analyses

Recently an Internet mineral dealer (www.miner-slunchbox.com) offered stolzite crystals and powellite after scheelite from Darwin. These were light gray sharp octahedral single crystals and groups of crystals to 2 cm (Figure 1). Reddish-brown, more crudely formed crystals described as powellite after scheelite were also offered (Figure 2). Four gray crystals and two brown crystals were sent to us for verification by Robert E. Reynolds. These were examined by X-ray diffraction (XRD) and energy dispersive X-ray spectroscopy (EDXS) in a scanning electron microscope (SEM). An octahedral face of each specimen was examined directly by XRD using copper $k\alpha$ radiation. All patterns of the gray crystals were good matches to

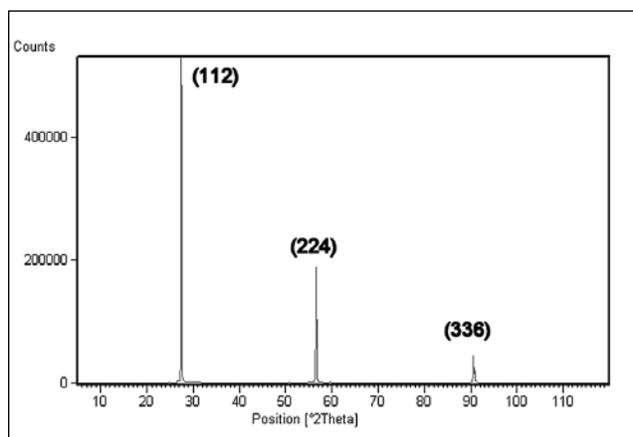


Figure 3. XRD pattern of octahedral face of gray stolzite crystal.

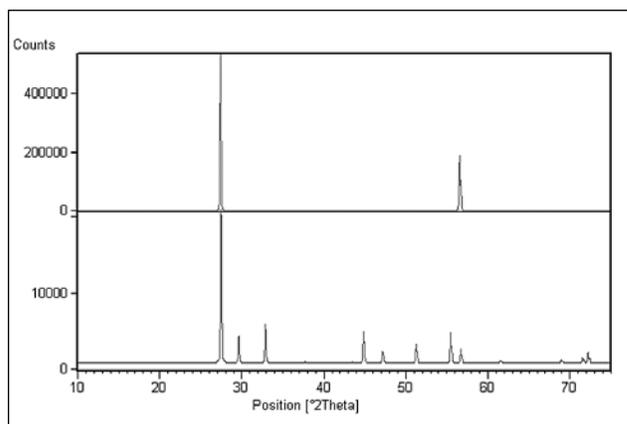


Figure 4. Comparison of observed XRD pattern of Darwin stolzite (top) with calculated XRD pattern for random stolzite (bottom).

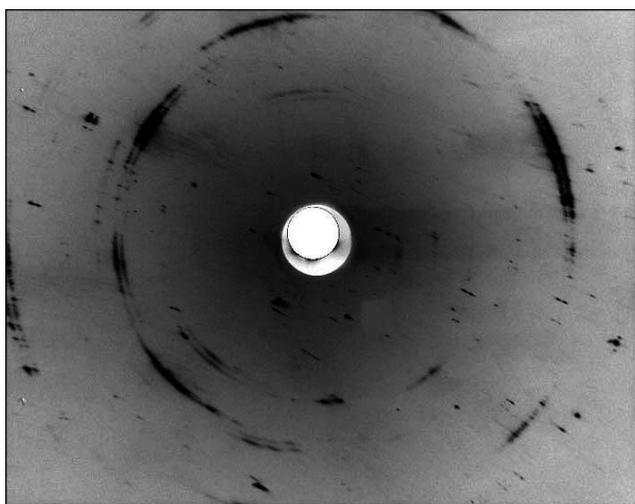


Figure 5. Back-reflection Laue photograph from octahedral face of a gray stolzite crystal (5 cm film to sample distance).

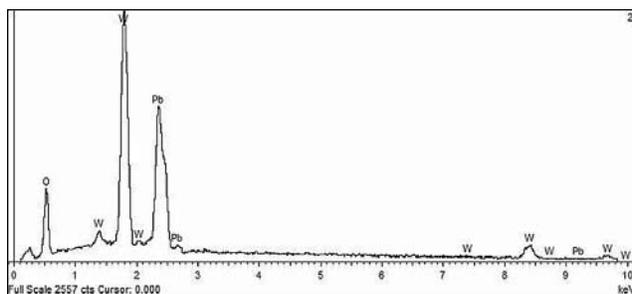


Figure 6. EDX spectrum from gray stolzite crystal.

stolzite, but with extreme (112) preferred orientation (Figures 3 and 4). The dominant reflections are (112), (224) and (336). Other reflections are extremely weak to absent. This suggested that the stolzite crystals may be primary crystals, although they had been described as pseudomorphs after scheelite (Moore, 2006).

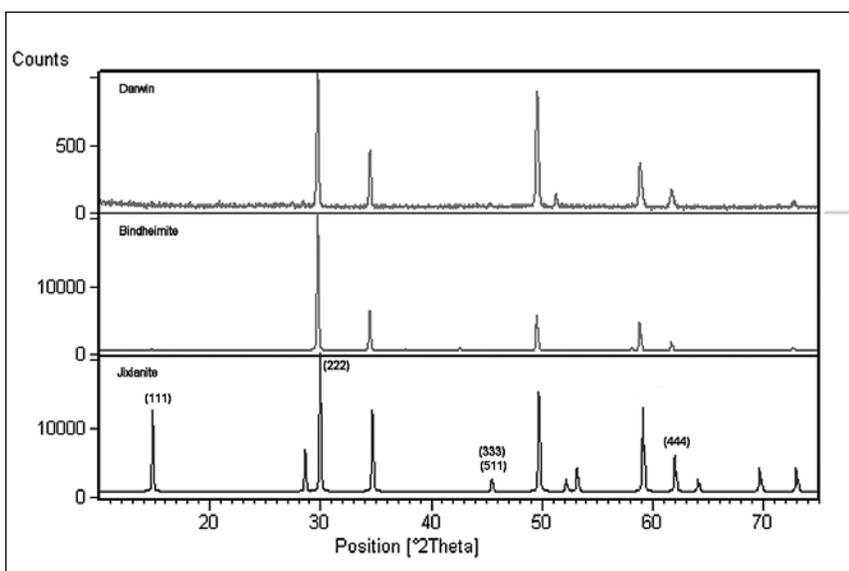


Figure 7. XRD pattern of brown jixianite after scheelite crystal (top) compared with calculated XRD pattern of jixianite reference (bottom, ICDD 33-760) and calculated pattern for bindheimite.

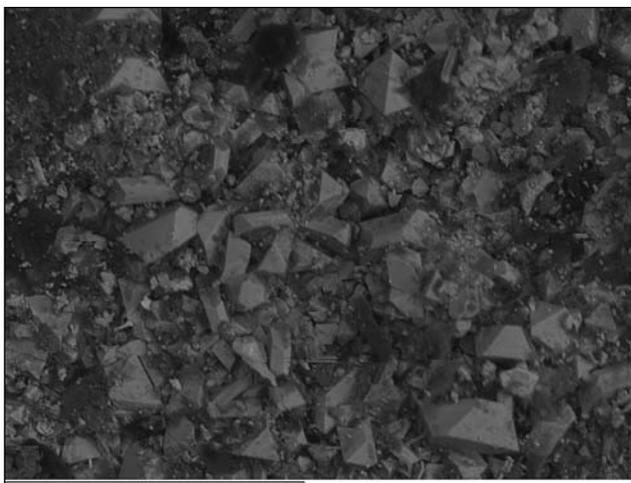


Figure 8. SEM image of jixianite crystals (to 30 μm) on surface of brown crystal.

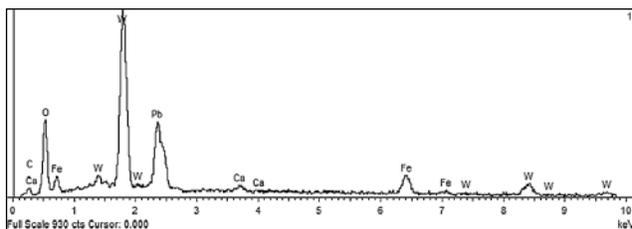


Figure 9. EDX spectrum of brown jixianite after scheelite crystal.

In order to determine the primary or secondary nature of the stolzite crystals, the octahedral faces of two smaller, sharper crystals were examined by the back-reflection Laue technique. With this technique single crystals produce geometric arrays of intense spots which display the symmetry associated with a particular face. In comparison, polycrystalline materials produce a series of rings. The two stolzite faces that were examined produced an array of discontinuous spotty rings (Figure 5). This indicates they are not primary crystals but pseudomorphs, but based on the very high degree of (112) preferred orientation in the powder XRD scan it is suggested that they are semi-epitactic replacements of stolzite after scheelite. It is noted that the XRD pattern only originates from approximately the outer 5 microns of the crystal and does not reflect the state of replacement for the interior of the crystal. The octahedral faces of scheelite are typically (011) or

(112) (Palache, et al., 1951). The EDX spectrum (Figure 6) of a gray crystal showed only lead (Pb), tungsten (W) and oxygen (O), consistent with stolzite. The crystals fluoresce a light canary yellow under SW ultraviolet light with little response from LW UV.

The XRD patterns of the two brown crystals were reasonable matches with jixianite (Figure 7). It is noted that several jixianite reflections (in particular (111)) were very weak and it was assumed that this was a result of preferred orientation from analyzing the crystal face directly, even though it is a better match to bindheimite. The crystal structure of jixianite is related to the pyrochlore and stibiconite groups of oxides. Bindheimite [Pb₂Sb₂O₆(O,OH)] belongs to the stibiconite group. It is noted that when a small amount of material was removed from one of the crystals and randomized and analyzed by XRD in transmission mode, the pattern was a good match with a mixture of jixianite and scheelite (A. Kampf, pers. comm.). In this case the jixianite reflections had their normal intensities and the (111) reflection was prominent. In the SEM the surface of one of the brown crystals was coated with small (to 30 μm) distorted octahedral crystals (Figure 8). The EDX spectra of these and the other brown crystal contained Pb, W, Fe, O and minor Ca, consistent with jixianite (Figure 9). Locally there were small (to 75 μm) dipyrarnidal crystals that contained significant calcium, tungsten and oxygen. This is consistent with the identification of scheelite in one of the XRD patterns.

Jixianite, Pb(W,Fe³⁺)₂(O,OH)₇, was first described as a new mineral in 1979 from Jixian, Hebei Province, China (Jianchang, 1979). It occurs as red or brownish red octahedral crystals to 160 μm but is more commonly microcrystalline or granular (Anthony, et al., 1997) It occurs in the oxidized zone of tungsten-bearing quartz veins adjacent to a quartz monzonite stock. Associated tungsten bearing minerals include wolframite, scheelite, wulfenite, and stolzite (Anthony, et al., 1997). The only other reported world localities for jixianite are the Clara mine, Baden-Wurtemberg, Germany, and the Apache mine, Hidalgo County, New Mexico, where it is also associated with scheelite (DeMark, 2004).

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Geology and mining history of Cerro Gordo ghost town

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Introduction

The ghost town of Cerro Gordo is located at an elevation of 8300 feet in the southern Inyo Mountains 7.5 miles east of the town of Keeler, California (Figure 1).

Cerro Gordo is a Spanish term meaning “fat hill”. The mining district was so named for the abundant silver ore discovered there. The road from Keeler is steep, narrow, and winding, traversing numerous Paleozoic to Triassic sedimentary units which have been highly faulted and folded. The road bears the nickname “Yellow Grade” and climbs 5000 feet in elevation over the seven and a half miles. The mines of Cerro Gordo were the largest producer of silver ore in California and played an important role in the development of the City of Los Angeles. Today the town site (Figure 2) is privately owned by Michael Patterson who continues the restoration work he and his late wife Jody began in 1985. The town is open to visitors year round (weather permitting). The restored bunkhouse and Belshaw houses are available for overnight stays via reservation during the months of June thru October (approximate dates dependent on thawing and freezing periods).

The climate is typical of any desert environment with an average rainfall of less than six inches per year. Due to the high elevation, temperature changes from hot to near freezing can occur in a matter of hours. Flash floods are not uncommon in Keeler Canyon, making traveling during storms potentially hazardous. In the winter, occasional snow storms blanket the town, so come prepared when visiting Cerro Gordo.

Mining History

Cerro Gordo was discovered in 1865 by a Mexican prospector, Pablo Flores, and several unnamed companions. The early workings were the San Lucas, San Ignacio, San Francisco, and San Felipe mines. Due to the rugged terrain, lack of water, and remoteness from any significant population center,

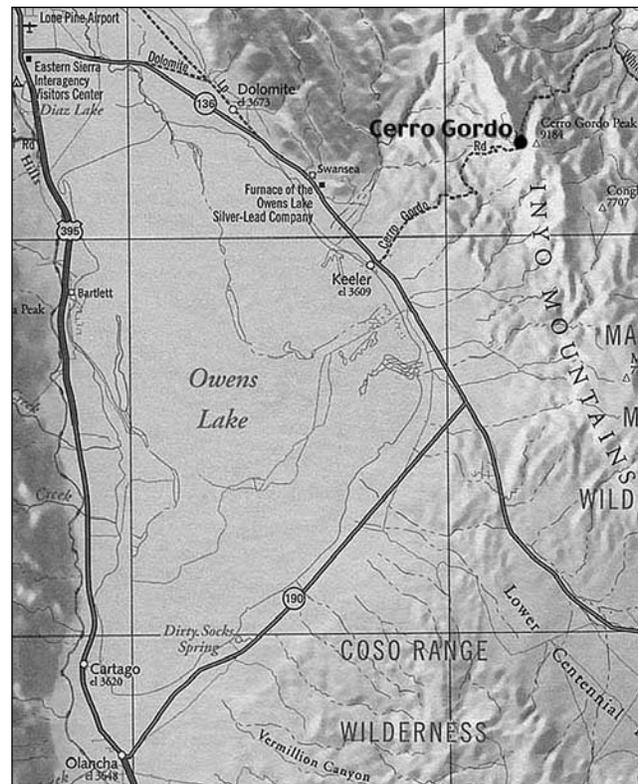


Figure 1: Index Map Showing Location of Cerro Gordo Ghost Town (Used with verbal permission of Michael Patterson)

Cerro Gordo was not an overnight success. However, the high grade of the ore (\$300 per ton) attracted Victor Beaudry, a merchant in Fort Independence, CA. Prior experience during the California gold rush (1848–1853) made Beaudry very familiar with min-



Figure 2: Cerro Gordo Ghost Town with snow. Photo by author.

ing gold and silver deposits. From 1866 to 1868 he managed to acquire several of the claims, including part of the Union Mine, which turned out to be Cerro Gordo's largest ore producer.

In 1868 Mortimer Belshaw arrived in Cerro Gordo. He realized the importance of the galena deposits (silver-lead) in the smelting process and that whoever controlled the galena deposits would control the mines. He acquired partial ownership of the Union mine, placing him in partnership with Victor Beaudry. Belshaw also realized that the transportation of ore, equipment, and materials was a major limiting factor due to the remoteness of the mine. After obtaining funds from investors in San Francisco, Belshaw developed the first "Yellow Grade" road, named for the yellow-colored rock formations that the road passed through. At the entrance he charged a toll for anything coming in and out of Cerro Gordo. The year 1868 was the transition year for Cerro Gordo, turning it from a small mining operation high in the Inyo Mountains to the largest silver producer in California history.

In 1869 Belshaw bought out several partners, leaving ownership of the Union mine to himself, Victor Beaudry, and Egbert Judson (the financier from San Francisco). These men then formed the Union Mine Company, which held the controlling interests of several mines on the mountain, including the Union. The second largest producer, the San Maria mine, was owned by the Owens Lake Silver Mining and Smelting Company. The third largest producer, the San Felipe mine, was acquired by the owners of the Union mine for past due debt. Some time later, the San Felipe Company, whose biggest stockholder was the Owens Lake Silver Mining and Smelting Company, declared ownership of the Union mine, stating that the San Felipe had been taken over illegally and that the underground workings of the Union mine were actually a part of the San Felipe. A protracted legal battle ensued between the Union owners and the San Felipe Company, which was financially backed by the Owens Lake Silver Mining and Smelting Company. Eventually, a judge ruled that the Union title belonged to the original owners of the San Felipe mine, giving controlling interests of the Union mine to the Owens Lake Silver Mining and Smelting Company. However, a "stay of proceedings" was declared, allowing the Union mine owners to continue operating both the Union and the San Felipe mines (Likes, 1975).

Improvements in the smelting process allowed the mines to produce ore at a faster rate. Belshaw developed a larger, more efficient blast furnace for the



Figure 3: The bar inside the American Hotel, built in 1871. Photo by author.

smelting process and incorporated this with a "water jacket" which equalized the heat in the furnace, allowing the mines to produce five tons of silver bullion per day, unheard of at the time. However, this posed a problem in that the mines were producing bullion at a faster rate than it could be shipped to Los Angeles. In 1869 Mortimer Belshaw hired Remi Nadeau, a highly respected teamster, to run the freighting operation that delivered the ore to Los Angeles. This business arrangement is considered directly responsible for the early growth of the city of L.A.

The Owens Lake Silver Mining and Smelting Company owned the smelter on the shores of Owens Lake at Swansea. An abundant source of water for operating the smelter was not an issue for them; however, the toll that Belshaw charged for use of his road was. As a result, the Owens Lake Silver Mining and Smelting Company sued Belshaw, charging that the tolls for use of the Yellow Grade road were excessive. The judge ruled that the toll be cut in half. This constant conflict between the two largest mine operators continued throughout the life of Cerro Gordo.

Water supply to the mines was also a serious problem. Belshaw resolved this by installing a pipeline from several springs north of the mining camp. Prior to this, all of the water was delivered by burros, a costly venture. The pipeline lasted three years; however, constant problems with the water supply meant that something else needed to be done. One of the other Cerro Gordo mine owners, Stephen Boushey, organized the Cerro Gordo Water and Mining Company. This new system pumped water from Miller Springs, ten miles north of Cerro Gordo and 1,860 feet below the crest of the mountains on the Saline Valley side (Likes, 1975). As a result of this venture,

Cerro Gordo now had an abundant water supply.

In 1871 the American Hotel was completed, and is currently the oldest standing hotel in California (Figure 3). The Owens Valley earthquake took place on March 26, 1872 and even though the towns in Owens Valley suffered extensive damage, the town of Cerro Gordo was largely unaffected. Mine production peaked in 1874. Afterwards, it steadily declined until the final wagon left Cerro Gordo in October 1878 when Victor Beaudry shut his furnace down for the last time. Production records of silver-lead ore from Cerro Gordo during the period of 1869 to 1876 are sketchy, but existing records place the value of production somewhere in the neighborhood of \$7,000,000 (Knopf, 1918), based on 1918 dollars.

A brief period of renewed interest in Cerro Gordo occurred in 1906 when it was discovered that zinc ore was present at the Union mine. This ore was mined by L.D. Gordon between the years 1910 and 1919. As a part of the mining operation, an

Table 1: Generalized Stratigraphic Column of the Cerro Gordo Mining District (Modified after Merriam, 1963 and Stone, 2004)

AGE	FORMATION	THICKNESS (FEET)	LITHOLOGY
Triassic	Union Wash Formation	???	Fine grained marine shales, siltstones, sandstones and limestones. Ammonite fossils common in the limestone units.
	Unnamed Units	4000 +	Marine shales and limestones.
	Unconformity		
Permian	Owens Valley Group	1800+	Silty and sandy limestone, fusulinid limestone, siliceous conglomerate, limestone conglomerate, shale, siltstone, sandstone and hornfels.
	Local Unconformity		
Pennsylvanian	Keeler Canyon Formation	2200+	Sandy and pebbly fusulinid limestone, shale, siltstone and marble.
Mississippian	Rest Spring Shale (Formerly mapped as Chainman Shale)	1000+	Dark grey silty shale and phyllite. Limestone interbeds.
	Mexican Spring Formation (Formerly mapped as Perdido Limestone)	0-200+	Limestone, chert, siltstone and quartzite.
	Tin Mountain Limestone	350	Dark grey limestone, chert nodules
Devonian	Lost Burro Formation	1600+	Light and dark grey marble, dolomite, quartzite
Early Devonian and Silurian	Hidden Valley Dolomite (Lower boundary difficult to establish in this area)	1700+	Massive light and dark grey dolomite, quartzite.
Ordovician	Ely Springs Dolomite	240-550+	Light and dark grey cherty dolomite
	Eureka Quartzite	400+	Light grey vitreous quartzite
	Badger Flat Limestone (basal part not exposed in this area) Formerly Mapped as Pogonip Group	1350+	Sacchoroidal dolomite and limestone

aerial tramway was built from just east of Keeler to Cerro Gordo. Remnants of this tramway can still be seen today. During this period, Cerro Gordo became one of the major sources for high-grade zinc ore in

the United States.

Several small deposits were found and mining operations continued until 1933, but none of the operations ever matched those of the peak years. From 1933 to 1985 several companies investigated the deposits at Cerro Gordo, but though potential economic ore bodies were discovered, political and environmental difficulties did not make re-opening the mines feasible. The current owner is currently disputing a bill that is moving through the U.S. Senate which will make all land surrounding Cerro Gordo part of a Federal Wilderness area.

In 1973, a share of Cerro Gordo was purchased by Jody Stewart, a retired game show host. Jody became sole owner in 1985 and began to restore some of the towns' buildings. She married Mike Patterson and together they worked on restoring Cerro Gordo. Jody passed away from cancer in 2001. Mike continues the restoration project along with the help of volunteers and the Cerro Gordo Historical Society. Cerro Gordo is a great place to get away from it all, and if you are lucky you may see the *ghost* in the Belshaw house. The restored bunkhouse is a great place to stay for group field trips. Reservations are a must for overnight stays but day visitors are always welcome. The road is hard on vehicles and four wheel drive is recommended, but not necessary. Just remember to drive with caution.

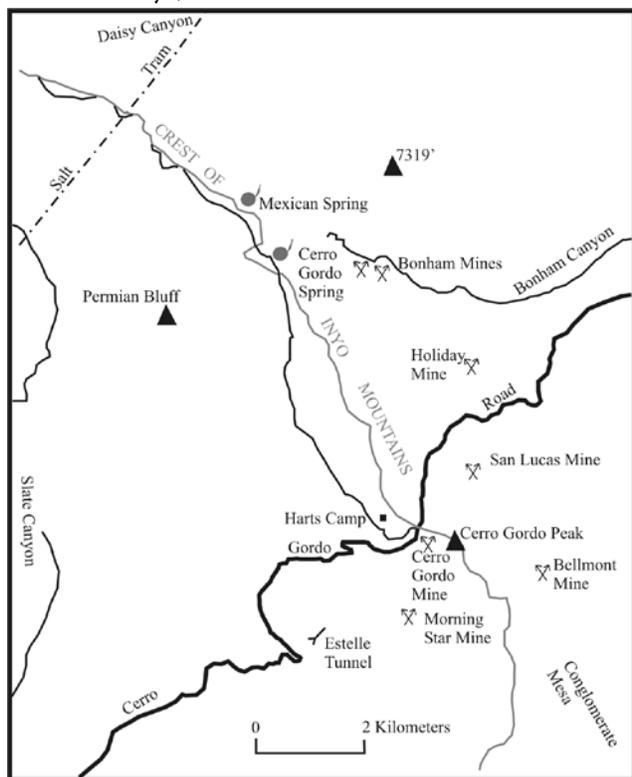


Figure 4: Map of Cerro Gordo Peak 7.5' Quadrangle showing selected geographic features and geochronologic sample locations of Stone and others. (From Stone, U.S.G.S. Map 2851)

Geology of Cerro Gordo

Regional Geologic Setting

In Paleozoic time the southern Inyo Mountains were part of the western continental shelf of North America. The carbonates and quartzites of the southern Inyo Mountains in the vicinity of Cerro Gordo contain sedimentary structures and fossils associated with shallow water deposition. Field work by Stevens and others (1991, 1997) indicate that these sediments accumulated along a northeast-trending continental margin. At Cerro Gordo these Paleozoic units are predominantly dolomites and limestones. Farther north, the strata exhibit characteristics associated with deep-water deposition (Stevens 1991).

During early Permian time, the area surrounding Cerro Gordo underwent contractional deformation and uplift. A second contractional event took place during late Permian time, causing further tilting and uplift. This subsequent uplift resulted in formation of an erosional unconformity between Permian and Triassic rock units. Later subsidence resulted in marine transgression during the Triassic. It was during this time period that deposition of the Union Wash Formation occurred.

Later withdrawal of marine waters from the region was followed by a period of volcanism. Deposition of volcanogenic strata during the early Jurassic (Stone, 2004) occurred simultaneously with early uplift of the Sierran magmatic arc. Sporadic intrusions in the form of dikes, sills and plutons occurred into the late Jurassic and Cretaceous. Radiometric age dates indicate that during this time emplacement of the Independence dike swarm occurred (Glazner and Carl, 2002). Extensional events during late Cenozoic time created the Inyo Mountains and adjacent valleys (Snow and Wernicke, 2000). This extensional faulting continues today.

Stratigraphy

The stratigraphic units of the Cerro Gordo mining district range in age from Ordovician to Triassic (Table 1). The Paleozoic sections at Cerro Gordo have an approximate total thickness of 11,100 feet (Merriam, 1963).

The Ordovician units are made up of the Badger Flat Limestone, the Eureka Quartzite and the Ely Springs Dolomite. The Badger Flat Limestone has been dated as middle Ordovician based on fossil assemblages. In earlier papers (Merriam, 1963) it was mapped as part of the Pogonip Group. The Badger

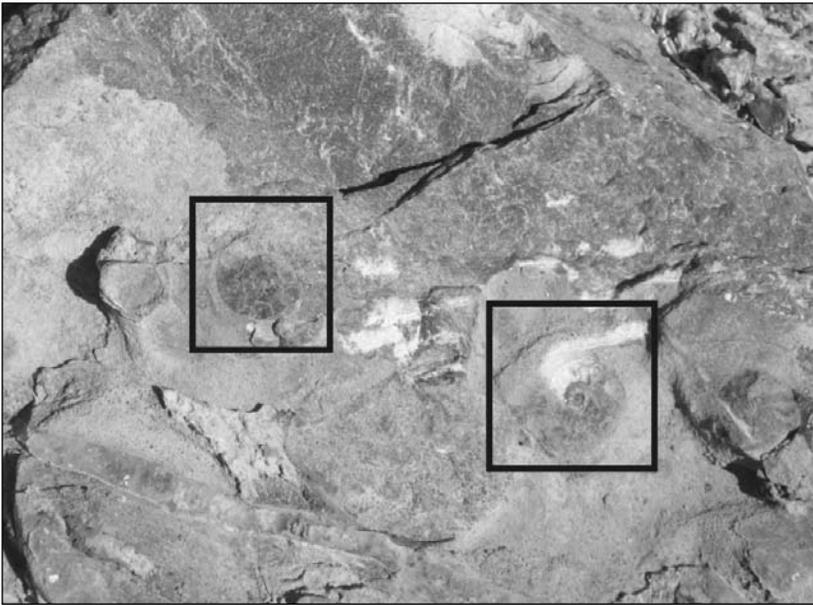


Figure 5: Ammonite in Union Wash Limestone (~2.5cm)/ Photo by author.

Flat Limestone is an iron-stained, medium to light grey, sugary dolomite. The main exposure is on the south side of Bonham Canyon (Figure 4), where it is overlain by the Eureka Quartzite on the east and is in fault contact with the Hidden Valley Dolomite to the west. The middle Ordovician Eureka Quartzite is a light-colored (nearly white), vitreous, medium- to fine-grained quartzite. It is best exposed along San Lucas Canyon, two miles northeast of Cerro Gordo. The Ely Springs Dolomite is a dark grey cherty dolomite. In some places the dark grey appears almost black. *Streptelasmid* horn corals of late Ordovician age are common in this unit between Bonham Canyon and San Lucas Canyon. The Hidden Valley Dolomite is late Silurian to early Devonian in age and can be described as a very light grey massive dolomite. This particular unit tends to form irregular ledgy slopes and differs from the type section in that it contains a higher percentage of quartzite and chert, most likely the product of hydrothermal alteration in the area.

The Devonian Lost Burro Formation was the most important ore-bearing unit in the district. Known to those in the mining profession as the “Cerro Gordo marble” (Merriam, 1963), all but one of the significant silver-lead ore bodies were found in it. The Lost Burro Formation is a massive, craggy, cliff-forming marble and limestone unit within the district. Its distinctive white to bluish grey and dark grey layers form alternating bands of white and grey. This distinct banding can be seen looking southeast from the Cerro Gordo town site. In the lower part of the formation, fossils are abundant and consist of stromatopoids, spaghetti corals and the late middle Devonian brachiopod *Stringocephalus*.

Fractures within the Lost Burro marbles acted as favorable ore hosts for sulfide replacement. Stratification in the Lost Burro exerted local influence upon ore deposition. Bedded ore was mined within the stratifications. Underground, the *Stringocephalus* beds

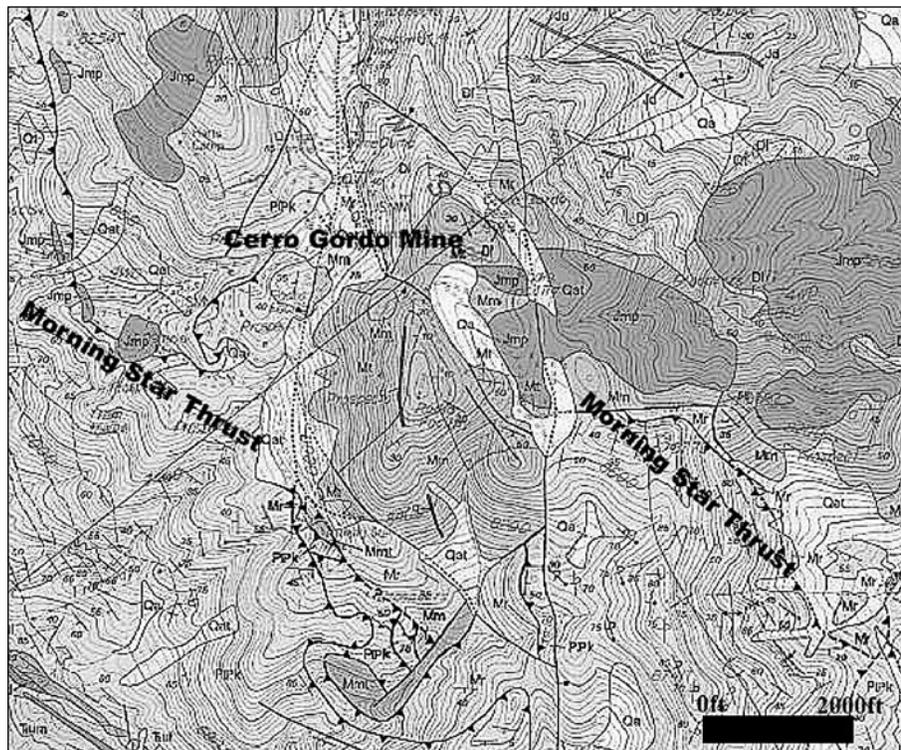


Figure 6: Portion of Geologic Map Cerro Gordo Peak Quadrangle USGS Map 2851 by Paul Stone & Others. Dsh—Hidden Valley, Dl—Lost Burro, Mt—Tin Mountain, Mm—Mexican Spring, Mr—Rest Spring Shale, Mmt—Undifferentiated Mexican Spring and Tin Mountain, PIPk—Keeler Canyon, Trum—Unnamed shales & limestones, Imp—monzonite porphyry, J—Jurassic dikes, q—quartz veins, Qa, Qat, Qt—Quaternary sediments

at the base of the Lost Burro were not evident. However, “coralline” beds were especially numerous in the mine workings (Merriam 1963). These “coralline beds” contained abundant spaghetti corals with some stromatoporoids. These distinct marker beds helped miners locate themselves within the section and were referred to as zone A and zone B.

The Tin Mountain Limestone, Mexican Spring Formation (previously mapped as Perdido Formation by Merriam, 1963) and the Rest Spring Shale are all Mississippian in age. At Cerro Gordo, the Tin Mountain Limestone is the dark grey band at the top of the cliff directly above the lighter colored Lost Burro formation. It is a bluish grey to dark grey, fine-grained limestone. The limestone has a platy weathering pattern, and crinoid and coral debris is common. It is overlain with sharp contact by the Mexican Spring Formation (Merriam 1963, Stone 2004). This formation is comprised of various lithologies ranging from siltstone through shale, sandstone, chert, limestone and conglomerate. It has very limited exposure within the district, having a measured thickness of less than 50 feet.

The Rest Spring Shale (formerly mapped as Chainman Shale by Merriam, 1963) is described as a dark grey to black carbonaceous clay shale, silty shale, sandstone and limestone. In the Cerro Gordo area the shale has been widely affected by low-grade metamorphism and alteration due to hydrothermal activity from the numerous igneous intrusions. Excellent exposures of this unit can be seen just northwest of the ghost town, an area popular among geology classes for fossil collecting.

The Keeler Canyon Formation overlies the Rest Spring Shale and is Pennsylvanian to early Permian age. The Keeler Canyon Formation is a deep-water marine turbidite sequence consisting of limestone, siltstone, and mudstones. The type section for the Keeler Canyon Formation is upper Keeler Canyon near the Estelle tunnel. Near Mexican Spring, chert nodules within the limestone beds have given the Keeler Canyon Formation the nickname of the “golf-ball horizon”. This horizon is a key marker bed within the district.

The Keeler Canyon Formation is unconformably overlain by the Permian Owens Valley Group, consisting of the Lone Pine Formation, several unnamed sedimentary members, and the Conglomerate Mesa Formation. The group consists of both marine and nonmarine lithologies that include silty to sandy limestone, biogenic limestone, argillaceous shale, siltstone, sandstone and conglomerate. Within the

district, it outcrops two miles south of Cerro Gordo road on the west side of the Inyo Mountains (Merriam 1963).

Formations of Triassic age unconformably overlie the Owens Valley Group. These as yet unnamed units consist of a basal sandstone unit and a series of interbedded limestones, siltstones and sandstones. Directly above the unnamed units lies the Union Wash Formation, which consists of interbedded shales, siltstone, sandstone and limestone. These lithologies are further subdivided into four members, all of which are found within the district. Several of these units are within imbricate sheets, the result of the numerous faults in the area. Measured sections of the Union Wash Formation along Cerro Gordo Road are fault-bounded (Stone and others, 1991) and exhibit severe deformation. Ammonites within the Triassic limestone and shale units are not uncommon (Fig. 5).

Igneous intrusives of the Cerro Gordo district consist of Jurassic-Cretaceous age granitoids and abundant andesitic and dacitic dikes. The granitic rocks were identified by Knopf in 1918 as quartz monzonite porphyry. As a result of the alteration and leaching by ore fluids, determination of the original rock type is difficult (Knopf, 1918). Of particular importance to Cerro Gordo was a narrow granitoid intrusive known as the Union dike. This dike was intruded along a north-trending shear zone that played an important role in subsequent ore deposition. The dike is 45 feet wide in the Union tunnel and closely followed the trend of the Cerro Gordo fault. This dike separates the Lost Burro from the Rest Spring Shale.

The andesite and dacitic dikes are thought to be a part of a northwest-trending series of dikes collectively known as the Independence dike swarm, as named by Moore and Hopson (1961). These dikes are found throughout the Inyo Mountains, the Sierra Nevada, and into the southern Mojave Desert. Late Jurassic dikes of the Independence dike swarm are extensively exposed in the Cerro Gordo Peak area (Stone and others, 2004). In the Cerro Gordo mine, these dikes cut the quartz monzonite Union dike and have frequently undergone hydrothermal alteration.

Geologic Structure

In the vicinity of Cerro Gordo the stratigraphic units have been severely deformed and truncated by both thrust and normal faults. Folding, overturned beds, and imbricate sheets of strata are not uncommon. The major thrust fault directly responsible for the deformation affecting the Cerro Gordo mining district is

the Morning Star thrust, as named by Elayer (1974) for its location adjacent to the Morning Star Mine (Figure 6). Structural relations suggest that the major zone of movement is localized along the Rest Spring Shale. The trace of the fault separates Rest Spring Shale from the overlying Keeler Canyon Formation. Key field observations within the thrust zone include variation of thickness of the Rest Spring Shale along strike, shearing along the contact between the Rest Spring Shale and the Keeler Canyon Formation, and truncation of folds within the Keeler Canyon Formation at the contact with the Rest Spring Shale. Near the Morning Star Mine, isolated blocks of Tim Mountain Limestone, Mexican Spring Formation and the Keeler Canyon Formation are enclosed by Rest Spring Shale along low-angle faults. These faults are believed to be syntectonic with the Morning Star thrust. The Morning Star thrust is believed to be part of an early Permian thrust system that includes the Last Chance thrust of Stewart and others (1966). Stevens and others (1997) proposed that the Morning Star thrust may actually have developed as a distal younger part of the Last Chance thrust.

The major structure within the area is the Cerro Gordo anticline (Merriam 1963). This anticline involves strata from Middle Ordovician to Middle Triassic and plunges to the south. It has been mapped over a distance of 15 miles. The important ore deposits of the Lost Burro Formation occur in the west limb of the anticline. Ore bodies and ore shoots within the Cerro Gordo mine rake southward with the plunge of the anticline (Merriam, 1963). Mississippian and Pennsylvanian strata within the west limb are more steeply dipping than the Silurian and Devonian strata of the east limb. Along the limbs of the anticline are numerous disharmonic and drag folds. Folding and deformation in the region is assumed to be either late Jurassic or early Cretaceous.

The northwest-trending Bonham Canyon fault structurally underlies the Morning Star thrust (Elayer, 1974). The Bonham Canyon normal fault's structural relations indicate that this fault pre-dates intrusion of the quartz monzonite porphyry (Merriam, 1963; Elayer, 1974). In Bonham Canyon, Merriam (1963) described a mile-wide zone of severe deformation and shearing along the Bonham Canyon fault. Exposures in the Estelle tunnel juxtapose eastward-dipping strata of the Hidden Valley Dolomite against westward-dipping strata of the Lost Burro Formation (Merriam, 1963). This relationship indicates that the Cerro Gordo Anticline is displaced along the Bonham fault by at least 5,000 feet (Elayer, 1974).

The Cerro Gordo fault is considered the district's master fault. This fault has a northerly trend and creates a zone of shearing for six miles from Soda Canyon to Bonham Canyon. Dip-slip movement along the Cerro Gordo fault has juxtaposed Rest Spring Shale and limestones of the Keeler Canyon Formation and marbles of the Lost Burro Formation and Tin Mountain Limestone in the footwall block. Exposures of fusulinid limestones in Keeler Canyon indicate a displacement magnitude of several hundred feet. Smaller en-echelon faults of the Cerro Gordo master fault intersect the Bonham Canyon fault in several places.

Both the Bonham Canyon and Cerro Gordo faults are believed to be part of an older fault system, that antedates the thrust and reverse faults of the area. Numerous high-angle normal faults cut the Cerro Gordo fault. These younger faults have separated the strata into isolated sheets. This is particularly evident in the underground workings (Merriam, 1963). The general strike of these younger faults is northwesterly.

The southern Inyo Mountains lie within the present day Western Basin and Range province. As such, they are bounded by an active Cenozoic fault system on the east side of the Cerro Gordo Peak quadrangle. This fault system is believed to have evolved during the last 3m.y. (Zellmer, 1980, Burchfiel and others, 1987, Snow and Wernicke, 2000). In the northeast portion of the quadrangle, evidence exists for an older episode of uplift and extensional faulting. No major zone of active faulting has been mapped along the western side of the southern Inyo Range. However, a northwest-trending gravity anomaly along the base of the range has been inferred to delineate a major normal fault system (Pakisier and others, 1964).

Mine Geology

Three types of igneous rock occur in the mine workings, these are quartz monzonite porphyry, quartz diorite porphyry, and diabase. The quartz monzonite porphyry is the oldest unit, followed by the diabase, and the quartz diorite porphyry. All of these rocks have been altered either by shearing or hydrothermal solutions. The porphyry is characterized by abundant feldspars and hornblende prisms. Accessory minerals such as titanite, apatite, and magnetite are found in thin section (Knopf, 1918). The diabase has been so extensively sheared that initial recognition was difficult, except for the fact that a less sheared exposure of it was found on the surface. It contains abundant

silver-lead ore and is the one exception to ore bodies that normally lie within the Lost Burro formation.

The predominant ore host is the marbles of the Lost Burro Formation. The primary ore bearing zone has a north-northwest trend with an average dip of 70 degrees SW (Knopf, 1918). This zone follows the Cerro Gordo fault. Some of the larger ore bodies are at least 40 feet thick. The lead ore bodies generally occur as lenticular masses distributed throughout a zone that is 1500 foot long, by several hundred foot wide zone within the Lost Burro Formation. The primary silver-lead ore is argentiferous galena. The zinc ore mineral is smithsonite. However, lesser amounts of sphalerite, tetrahedrite and pyrite are also present. Common oxidation products associated with the main ore minerals are cerussite, anglesite, and bindheimite. Linarite (a rare mineral for which Cerro Gordo is well known), caledonite and chrysocolla are also present.

Special significance is given to the many fissures and fractures in the footwall of the Cerro Gordo fault in terms of structural control. These fissures and fractures contained the largest ore bodies. The primary ore bodies follow the steeply inclined pipes which raked south in the direction of anticlinal plunge. These channels narrowed and widened, sometimes widening out to as much as 100 feet. It was in some of these "open spaces" that rich ore bodies were recovered.

There are four different types of ore at Cerro Gordo. These consist of the massive silver-lead ores, the diabase dike silver-lead ores, the siliceous vein ores and the carbonate zinc ores. Deposition of the silver-lead ore followed the intrusions of the quartz monzonite porphyry and diabase dike. The galena occurs in circular masses intergrown with sphalerite and pyrite. The siliceous veins typically have associated oxidation minerals such as malachite and azurite. The diorite porphyry is the primary unit associated with deposition of the zinc-bearing ore fluids. The lead ore typically contained from 50 to 80 ounces of silver per ton, and the zinc ore about 25 ounces of silver per ton (Merriam, 1963).

The Story Continues

The story of Cerro Gordo Ghost Town is a rich and colorful one. In terms of the geologic history, there are still some missing pieces which need to be filled in. For geologists, this is what keeps life interesting. For geology students it presents a challenging area to study and learn from. Some of the old buildings still

need to be restored and the history of Cerro Gordo continues to unfold. For some, it is just a place to visit. For many, the town holds an attraction, stepping back in time and feeling the town's pulse as the ghosts of Cerro Gordo rattle around the buildings and the old mine workings.

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Significance of the Independence Dike Swarm in understanding the tectonic evolution of Owens Valley and the western Basin and Range—a summary

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Introduction

The Present is the key to the Past—so stated Charles Lyell in his book *Principles of Geology* in 1830. This particularly holds true in the case of the Independence dike swarm discovered during the late 1950s. The discovery took place just prior to the emergence of the theory of plate tectonics during the 1960s. Over time, their recognition as an important structural and chronological marker unit has led to a better understanding of the development and timing of tectonic events within California. The following is a summary of past and present research on the Independence dike swarm and the evolution of ideas regarding it.

Geologic History and Tectonic Setting

The tectonic puzzles of the Mojave Desert, southwestern Basin and Range, and Sierra Nevada are very complex. Within the Mojave Desert region, the major developmental phases of the Cordilleran orogen have been recorded. During the Neoproterozoic, the western edge of continental North America, which was part of the supercontinent of *Rodinia*, underwent continental rifting (Karlstrom, 1999). Remnant basement rocks from this rifting can be found throughout the Mojave Province. These 1.8 Ga-old remnants are referred to by several names, including Fenner Gneiss, Kilbeck Gneiss, and Dog Wash Gneiss. During the Paleozoic, the continental margin was fairly quiet, undergoing a long period of sedimentation. A thick sequence of miogeoclinal sediments (~10,000 m) (Stevens, 1991) to the east of the margin and eugeoclinal strata to the west of the margin were deposited during this time. Strike-slip truncation of the continental margin occurred late in the Paleozoic,

causing southward displacement of eugeoclinal rocks. The boundary created by this truncation placed miogeoclinal–cratonal facies against eugeoclinal rocks (Walker et al., 2002). Deformation of the Paleozoic strata within this accretionary wedge subsequently took place.

The change in the position of the continental margin from a passive margin setting to an active convergent plate margin setting is reflected in the character of the rock units. Evidence consists of changes in stratigraphic sequences, variations in stratal deformation, and in local magmatic activity (Walker et al., 2002). In the northern Mojave Province, eugeoclinal rocks and volcanoclastic strata within the El Paso Mountains provide evidence for the timing of this change. Deformation within the sedimentary units and late Permian volcanoclastic units consists of west-vergent folds and west-directed faults. Additional evidence is provided by a gneissic pluton, radiometrically dated at 260 Ma (Miller and Glazner, 1995), that intruded these deformed units and is interpreted to be synkinematic with changes along the margin (Carr et al., 1984). Subduction along the margin is believed to have begun during the late Permian and continued into the early Triassic. Further evidence supporting the beginning of subduction is shown by stratigraphic relationships, associated deformation, magmatic activity, and crustal shortening events. However, the mechanism for late coeval Paleozoic strike-slip truncation of the margin still remains controversial.

Several orogenic events in central and northwestern Nevada were synchronous with margin subduction, beginning during the Permian and continuing into the Mesozoic. The oldest of these events was the

Devonian–Mississippian Antler Orogeny whose timing coincided with the breakup of *Rodinia*. Although Proterozoic remnants (1.8 Ga basement rocks) of *Rodinia* are present, deposits of the Antler orogenic event are not found (Walker, 1988). The Antler Orogeny was followed by the Sonoma Orogeny during the late Permian into the Middle Triassic. Sedimentary strata from this event are consistent with those associated with back arc basin deposition.

The Nevada Orogeny began during the Jurassic, about 155 to 150 Ma. Emplacement of the Independence dike swarm (~166–148Ma) occurred late in the development of the Jurassic arc. A magmatic lull occurred in the period shortly after emplacement of the dikes. Then, during the Early Cretaceous (~120Ma), a renewal in tectonic activity occurred and lasted until approximately 75Ma. During this time period (96–88Ma) (Coleman et al., 2000), a second episode of emplacement of the Independence dike swarm intruded older rock units throughout the Sierra Nevada and Mojave provinces.

Coincident with orogenic events, the continental crust experienced both compression and extension. During the Sevier Orogeny of the Cretaceous, thin skinned thrusting and uplifts of cratonal rocks took place. About 75 Ma, the Laramide Orogeny began. At this time subduction of the margin became shallower and “underplating” of accretionary wedge material to continental crust occurred under California. As a result of this shallow subduction, arc magmatism migrated eastward, causing the Sierran arc to become inactive.

Widespread detachment faulting took place during the Oligocene when the East Pacific Rise intersected the subduction zone of the Fallarón Plate marking the early phase of Basin and Range extension. Later, high-angle extensional block faulting occurred during the Miocene, about 17 Ma. Shortly after this time (~16Ma), the transition from a converging plate boundary to a strike-slip boundary began. Rap-

id changes in the direction of plate motion during this time as well as plate interactions may have influenced rotation of the dike swarm. Strike-slip motion trends in a northwesterly direction and continues today. As a result of this northwest-directed strike-slip motion, rotation of the Mojave Block occurred. This rotation continues south of the Mojave Block with the rotation of Baja California. The importance of the Independence dike swarm becomes evident at this point. Using the rotation data from the dike swarm, geologists are attempting to reconstruct the paleogeographical position of the Mojave and Sierra Nevada blocks, which leads us to the next part of the story.

The Independence Dike Swarm

The Independence dike swarm, first recognized during the late 1950s, was named for the town of Independence, CA, which is near the type locality (Moore and Hopson, 1961). Smith (1962), working south of the Garlock fault, recognized similar dikes

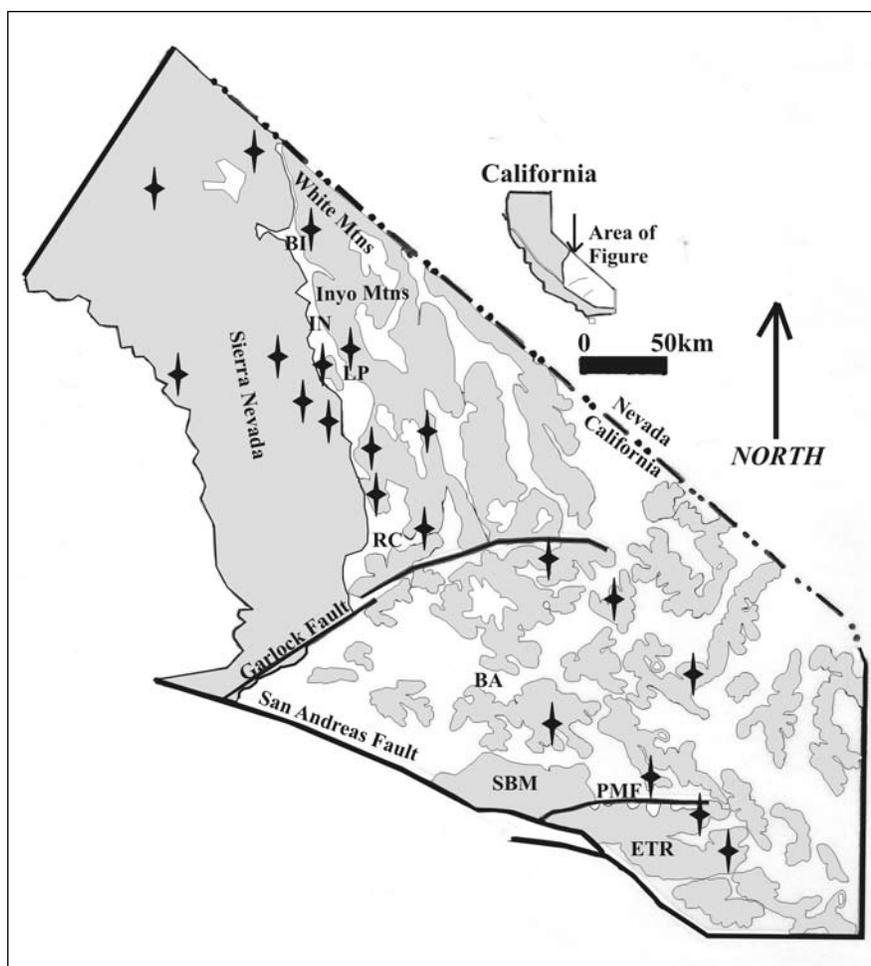


Figure 1: Regional extent of the Independence dike swarm. Star crosses indicate dike localities. BA, Barstow; BI, Bishop; ETR, Eastern Transverse Ranges; IN, Independence; LP, Lone Pine; PMF, Pinto Mountain Fault; RC, Ridgecrest; SBM, San Bernardino Mountains. Modified from Hopson et al., 2008



Figure 2: Coarse-grained Independence dike off of Dolomite Loop Road, Hwy 190. These dikes cut Ordovician Ely Springs dolomite and Devonian Hidden Valley Dolomite. Geochemical analysis indicates that the dike at this location is a gabbro. Photo by author.

and proposed that they were part of the same swarm. Additional field studies in both the Sierra Nevada and Mojave provinces showed the dikes to be continuous (Moore and Hopson, 1961; Smith, 1962). Later studies and geochronology showed that the swarm extended at least 600 km from the central Sierra Nevada to the southern Mojave Desert. The Independence dike swarm is an important geologic marker because of its regional extent and short period of intrusion (Figure 1). Radiometric dates place emplacement of the dikes between 166 Ma to 148 Ma and 96 Ma to 88 Ma.

Petrographic composition is highly variable, in part because the dike swarm has such an extensive range and the dikes crosscut multiple rock types. The dikes are dominantly bimodal, being either silicic or basaltic (Moore and Hopson, 1961). However, many intermediate types are also present (McManus and Clemens-Knott, 1997) (Figure 2).

Dike textures are also variable. They can be fine- or coarse-grained, porphyritic, or glassy. The porphyritic dikes range in composition from diorite porphyry to granodiorite porphyry. The fine-grained dikes are microcrystalline diorite or microcrystalline granite. Most of the dikes have been altered or partially recrystallized. In the Sierra Nevada, the dikes are pervasively metamorphosed (Chen and Moore, 1979). Typically, the northern Independence dike swarms found north of the Garlock fault, in the Sierra Nevada and White-Inyo Mountains, are predominantly mafic

(e.g., Moore and Hopson, 1961; Moore, 1963; Ross, 1965; Nelson, 1966a, 1966b; Ernst, 1997; Carl et al., 1998). In the southern part of the swarm south of the Garlock fault, felsic and intermediate compositions are more common (Carl et al., 1997; Carl, 2000) (Table 1).

The dikes range in size from a few centimeters to more than 10m; the average is about one meter. Most of the Independence dikes sharply transect the country rock and have chilled margins (Chen and Moore, 1979). Generally, most dikes have a dip between 60–90 degrees and in the Sierra Nevada most strike N30°W (Figure 3). However, southeastward of the Inyo Mountains the strike is variable ranging from N10°W to N70°W (Chen and Moore, 1979). The consistent northwest strike



Figure 3: Quartz dike cutting coarse-grained Independence Dike, Little Lake Road West, Southern Sierra Nevada Range. The dikes at this location exhibit characteristics consistent with Cretaceous dikes (Glazner et al., 2005). Photo by author.

Table 1: Characteristics Summary of Independence Dike Swarm (modified from Carl and Glazner, 2002; Hopson et al., 2008)

Age (Ma)	Composition	General Location	Dating Method	Reference
140	Metadiabase	White Mountains	⁴⁰ Ar/ ³⁹ Ar	Hourigan et. al (2003)
140	Diabase	Inyo Mountains	U-Pb	Dunne & Walker (2004)
140	Felsic	Inyo Mountains	U-Pb	Dunne & Walker (1993)
144	Metadiabase	White Mountains	⁴⁰ Ar/ ³⁹ Ar	Hourigan et. al (2003)
145	Felsic	Old Woman Mountains	U-Pb	Gerber et. Al (1995)
145	Rhyolite	Eagle Mountains	U-Pb	James (1989)
146	Dacite	Colton Hills	⁴⁰ K/ ¹⁰ Ar	Fox & Miller (1990)
147	Rhyolite	Stoddard Well	U-Pb	James (1989)
148	Silicic	Spangler Hills	U-Pb	Chen and Moore (1979)
148	Rhyolite	Woods Lake	U-Pb	Coleman et al.(2000) <i>in</i> Carl and Glazner (2002)
148	Diorite	Woods Lake	U-Pb	Coleman et al.(2000) <i>in</i> Carl and Glazner (2002)
148	Silicic	Alabama Hills	U-Pb	Chen and Moore (1979)
148	Silicic	Argus Mountains	U-Pb	Chen and Moore (1979)
148	Microdiorite	Coso Range	N.D.	Whitmarsh ¹ <i>in</i> Carl and Glazner (2002)
148	Silicic	Coso Range	N.D.	Whitmarsh ¹ <i>in</i> Carl and Glazner (2002)
148	Silicic	Cowhole Mountains	N.D.	Carl (2000) <i>in</i> Carl and Glazner (2002)
149	Granite	Sierra Nevada	U-Pb	Coleman et. al. (2000)
151	Diorite	Sierra Nevada	U-Pb	Coleman et. al. (2000)
151	Mafic	Coso Range	N.D.	Whitmarsh ¹ <i>in</i> Carl and Glazner (2002)
152	Felsic	Cronese Hills (NE Mojave Desert)	N.D.	Walker <i>in</i> Carl and Glazner (2002)
153	Granite	Sierra Nevada	U-Pb	Coleman et. al. (2000)
157	Diorite	Sierra Nevada	U-Pb	Coleman et. al. (2000)

Note: N.D. -No Data.

¹R. Whitmarsh (2000, personal commun.)

has proven invaluable in reconstructing the deformational history of the regions which have undergone vertical axis rotation and lateral displacement (Smith, 1962; Schermer et al, 1996; Glazner et al., 1989; Dunne and Walker, 1993; Ron and Nur, 1996; Hopson et al., 2000; Hopson et al., 2008).

The Independence dikes have a consistent northwest strike. The Garlock fault separates the northern Independence dikes from the southern Independence dikes. However, dikes north of the Garlock fault, including those in the Alabama Hills, Coso Range, and Spangler Hills, consistently strike west-northwest (~315°), counterclockwise to the overall ~330° trend of the swarm (Glazner et. al., 1999). Locally in the Sierra Nevada, strike of the dike swarm deviates almost 90° from a northwest trend. These deviations may be a result of the dikes intruding fractures that were oriented obliquely to the swarm's overall trend, or they could be a result of rotation due to intrusion of plutons during the Cretaceous (Moore and Sisson, 1987).

South of the Garlock fault, dikes have been variably rotated due to Cenozoic faulting and crustal extension (Smith, 1962; Schermer et al., 1996). Independence dikes in the Granite Mountains have been

rotated about 50° clockwise relative to dikes in the Sierra Nevada (Smith, 1962; Ron and Nur, 1996). Structural data reported by Carl and Glazner (2002) support clockwise rotation in the north-eastern Mojave Desert. Mafic dikes in the western Avawatz Mountains, in the northeastern Mojave, strike predominantly ~340°. Clockwise rotation associated with movement across the Garlock fault likely caused rotation of dikes in the Granite and Avawatz Mountains (e.g., Smith, 1962; Luyendyk et. al., 1985; Schermer et al., 1996).

A prominent gap within the swarm occurs within the region between the northeastern Mojave Desert and the area around

Barstow, CA, within the central Mojave Desert. This gap most likely exists due to structural deformation by low-angle detachment faults in the area. North-east-directed extension possibly exceeded 40km (e.g., Glazner et al., 1989; Walker et al., 1990a., Martin et al., 1993; Glazner et al.,1994). The swarm continues south in the Stoddard Ridge area southwest of Barstow in the west-central Mojave Desert. Dikes in the Stoddard Ridge area have a mean strike of ~312°, subparallel to unrotated Independence dikes north of the Garlock fault (Carl and Glazner, 2002). Paleomagnetic data from several mountain ranges within the central Mojave Desert indicate significant clockwise rotation (Ross et. al. 1989). The difference in rotation is probably due to Miocene extension in the central Mojave Desert.

Investigations

Early investigations by Smith (1962) used Independence dike swarms in the Granite Mountains south of the Garlock fault and dike swarms in the Spangler Hills north of the fault to determine the amount of left-lateral displacement along the Garlock fault. Smith concluded that the previous interpretation of

40 km of displacement was too low. The evidence provided by the dike swarms in the Spangler Hills, Argus Range, and Granite Mountains at Fort Irwin showed that the horizontal displacement was closer to 65 km. Smith also recognized that dikes found farther south in the Soda Mountains, Tiefert Mountains, Alvord Mountains, and Colton Hills had similar characteristics to those in the Granite Mountains. Based on this evidence, he believed that the dike swarm extended for a total distance of 250 km. Because of this, he also realized that the dike swarm could be used in future investigations to reconstruct the geologic history of the region. This conclusion was supported by Moore and Hopson's 1961 work, which involved an investigation of a 135 km-long Independence dike swarm that they believed was Cretaceous in age. Later studies showed that most of the dikes were Jurassic in age (Chen and Moore, 1979) but that some of the swarm was indeed Cretaceous (Glazner et al., 2000). The true importance of the Independence dike swarm to the reconstruction of tectonic settings and timing of events has become clearer as new pieces of the puzzle in the tectonic evolution of California have been identified.

The significance and extent of the Independence dikes was the subject of a paper by James in 1989. James believed that defining the regional extent of the dike swarm might help provide critical evidence for reconstructing Jurassic tectonic events. While examining Jurassic dikes throughout the western Mojave Desert and eastern Transverse Ranges, James realized that these dikes had similar characteristics to the already-known Independence dike swarm. Based on this new evidence, James concluded that the dike swarm actually extended regionally for a total distance of more than 500 km and may even extend as far south as Sonora, Mexico, based on evidence provided by Silver and Anderson (1983). He also concluded that the Independence dikes, because of their restricted age range, were useful for constraining tectonic and age relationships. Based on the dikes' cross cutting relationships, they could be used to establish relative ages of Upper Jurassic and Cretaceous rocks. They could possibly be used to constrain terrane accretion ages, in that they represented magmatic processes across a wide range of wall-rock settings. The dike swarm also appeared to record translations and rotations of crustal flakes, and field relations involving the dike swarm indicated that it was part of the last episode of extension in a series of alternating extensional and compressional events during the Jurassic. James also concluded that the tectonic regime re-

sponsible for Independence dike emplacement could be related to arc-normal extension, changes in plate motion, or oblique subduction with left-lateral shear.

Reconstruction of the Paleogeographic setting of the Sierra Nevada and Mojave Province was the subject of several papers throughout the 1990s. Studies by Walker et al. (1990) of deformation within the Cronese Hills used the continuity of the Independence dikes within the study area as a means to restore Tertiary offsets and confirm evidence for stacking of miogeoclinal rocks against eugeoclinal rocks as a result of Middle Triassic and Jurassic age thrust events. It is this author's belief that the Cronese shear zone was a part of the southern continuation of the east Sierran thrust system during the Jurassic. Preservation of Jurassic arc-related rocks by Mesozoic thrust faults may be responsible for the structural stacking of Paleozoic eugeoclinal and miogeoclinal strata in the central Mojave desert.

In 1994, Schermer and Busby presented a paper that examined the changing character of volcanism and the paleogeography of the continental arc during the Jurassic. Examination of the Lower Sidewinder volcanic series indicated collapse of a nested caldera complex followed by north-south extension that coincided with batholith emplacement. The Upper Sidewinder volcanic series provides evidence for a larger-scale transtensional event during intrusion of the Independence dike swarm. The explosive volcanism that caused the collapse of the caldera complex was followed by normal faulting, tilting, and erosion. Intrusion of Middle Jurassic plutons was contemporaneous with these tectonic events. The 148Ma Independence dikes intrude the Upper Sidewinder Volcanics. Regional northeast-southwest extension is supported by the presence of the northwest-striking dike swarm. The lack of debris flow deposits and epiclastic rocks suggests an intra-arc region of low relief. Based on this evidence, the authors suggested a left-oblique subduction regime for the Jurassic arc of the southern cordillera (Schermer and Busby, 1994).

Examination of rock units within the Chuckwalla Mountains of the eastern Transverse Ranges, California, by Davis et al. in 1994 attempted to distinguish timing and kinematics of tectonic events along the continental margin during the Jurassic. Widespread occurrences of Late Jurassic deformation and plutonism appeared to reflect regional responses to complex Jurassic plate interactions. Emplacement and deformation of Jurassic dike swarms of the eastern Transverse Ranges seemed to indicate a series of alternating extensional and compressional events

(Wolf and Saleeby, 1992). The presence of Independence dikes in the Eagle and Chuckwalla mountains provided direct evidence of a Late Jurassic zone of extension from the eastern Transverse Ranges through the Mojave Desert to the Sierra Nevada. During this extension, a suite of bimodal, alkalic element enriched plutons were emplaced throughout the central and southern Mojave region. Glazner (1991) suggested that emplacement of these plutons during transtension occurred in response to left-oblique convergence. The question still remains, however, whether or not the opening of the Gulf of Mexico was responsible for generation of the alkalic magmas in the Mojave Desert region. After reviewing geochemical data, emplacement, and deformation of the dike swarm and alkalic plutonism, Davis et al. (1994) concluded that the nearly contemporaneous expansion and contraction of continental crust resulted from the combination of oblique convergence and the opening of the Gulf of Mexico.

Rotation of the Mojave block around a vertical axis was the topic of a paper presented by Ron and Nur in 1996. They used data from the Independence dike swarm to determine the best of four rotation models previously proposed by several authors. The first model, proposed by Garfunkel (1974), stated that the Garlock fault direction had remained fixed and postulated that the northwest-trending faults had rotated 40° to 50° counterclockwise, whereas the east-west trending faults of the northeastern Mojave Desert and eastern Transverse Ranges were unrotated. The second model, put forward by Luyendyk et al. (1985), proposed a 30° to 50° clockwise rotation of the east-west trending faults in the northeastern Mojave and eastern Transverse Ranges and no rotation of the northwest-trending faults in the central Mojave. The third model, proposed by Dokka (1983) and Dokka and Travis (1990), concluded that no rotations of blocks or faults around a vertical axis had taken place. The final model, as proposed by Nur et al. (1989), stated that a symmetrical rotation around the principal stress direction caused a 25° to 30° counterclockwise rotation of the northwest faults and a 25° to 30° clockwise rotation of the east-west faults. Using the presumed unrotated Sierra Nevada as a reference direction, Ron and Nur (1996) determined that the model proposed by Luyendyk et al. (1985) was the most accurate model. Dike orientations indicated a 40° to 50° vertical clockwise rotation of the eastern Mojave and Eastern Transverse Ranges domain and no rotation for the central Mojave domain. The clockwise rotation direction of the eastern Mojave

and eastern Transverse Ranges is consistent with the dominant right-slip deformation mechanism of the Eastern California shear zone and San Andreas fault.

The existence of Cretaceous age Independence dikes was the subject of a paper presented by Coleman et al. (2000). Several dikes were sampled within the type locality at Woods Lake in Onion Valley in the Sierra Nevada and at Pine Creek near Bishop, CA (Table 2). Dike orientations and petrographic and chemical analysis indicated that the Cretaceous dikes were remarkably similar to the Jurassic age dikes. The Cretaceous Independence dikes are believed to be somehow related to coeval mafic and felsic plutons in the area. However, Coleman et al. also stated that the presence of Cretaceous dikes is not a local phenomenon but that they are found throughout the swarm. Comparison of petrographic and chemical characteristics showed no distinguishing criteria between the Jurassic and Cretaceous age dikes. The paper did mention that the Jurassic dikes contained a left-shear fabric that was not present in the Cretaceous dikes. The authors concluded that the orientation of the dikes may have been controlled by preexisting faults or by the orientation of the continental margin, rather than by stresses related to the subducting plate.

A recent paper by Hopson et al. (2008), addressed the topic of dike orientations and tectonic rotation in eastern California. Fault block rotations are an important consequence of movement along the San Andreas fault. These rotations are believed to have occurred coincidentally with Miocene extensional events. Analysis of the strikes of 3841 dikes across 47 domains indicated a clockwise rotation from the

Table 2: U-Pb data for Independence dikes and Cretaceous dikes (modified from Coleman et al., 2000)

Location	Rock Type	Ages (Ma)		
		²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb
Woods Lake	Diorite	157	158	167
Woods Lake	Diorite	151	151	154
Woods Lake	Diorite	151	151	154
Woods Lake	Diorite	94.6	95.1	108
Woods Lake	Diorite	94.2	94.5	101
Woods Lake	Diorite	94.3	95.8	132
Woods Lake	Diorite	94.4	94.7	102
Onion Valley	Granite	236	252	398
Onion Valley	Granite	149	151	178
Onion Valley	Granite	153	156	205
Pine Creek	Diorite	90.0	93.7	189
Pine Creek	Diorite	89.3	93.6	203
Pine Creek	Diorite	88.8	91.8	170

Data Source: Coleman et al. 2000

Table 3: Rotation Data Summary of Independence Dike Swarm (modified from Carl and Glazner, 2002; Hopson et al., 2008).

Domain	Dominant Composition	Number of Dikes	Mean azimuth as measured today (°)	Standard Deviation (°)	0.05	Mean when restored for paleomagnetic rotation (°)	Paleomagnetic Rotation	Reference	
Sierra Nevada									
Alabama Hills	Felsic, Mafic	37	311.1	19.2	6.4	311.1	0	Stone et. al. (2000)	
Ranges East of the Sierra Nevada in the Basin and Range Province									
Darwin Falls	Undivided	49	300.3	9.9	2.8			Stone et. al. (1989)	
El Paso Mountains	Mafic	24	3.600	10.1	4.0	21.1	17.5°CCW	Carr et. al (1997)	
Inyo Mountains	Felsic, Mafic	200	318.2	26.1	3.6			Stone et. al. (2000), Rinehart and Ross (1957), Conrad and McKee (1985), Elayer (1974), Kelley (1973), Stuart (1976), P.A. Stone (personal commun. 1998)	
White Mountains	Mafic	70	349.4	29.6	6.9			Ernst (1997), Nelson (1966b)	
N. Inyo Mountains	Mafic	41	354.2	34.8	10.7			Nelson (1966a)	
Santa Rita Flat	Felsic, Mafic	114	336.7	33.8	6.3			Ross (1965)	
N. Argus Range	Mafic	23	290.9	14.5	5.9	290.9	0	Prevost (1984)	
Panamint Butte	Undivided	43	305.8	13.0	3.9	305.8	0	Hall (1971)	
Southern Argus Range/Spangler Hills	Felsic, Mafic	960	312.4	19.3	1.2	312.4	0	G.I. Smith (personal commun. 1998) (Hopson, 2008)	
Northern Mojave Desert									
Granite Mountains	Mafic	70	337.7	9.3	3.1	274.2	63.5°CW	E.L. Schermer, (personal commun. 1998) Hopson, (2008)	
S. Avawatz Mountains	Mafic	29	332.9	6.0	2.2	269.4	63.5°CW	Aerial Photographs: The National Aerial Photography Program (6830-103, 6870-126, 6871-23, 6871-47)	
Eastern Mojave Desert									
Cowhole Mountains	Felsic, Mafic	86	319.9	40.6	8.7	290.3	29.6°CW	Wadsworth et. al. (1995)	
Colton Hills	Felsic, Mafic	34	334.8	6.3	2.5	305.2	29.6°CW	Goldfarb et. al. (1988)	
Providence Mountains	Felsic, Mafic	34	345.7	20.5	6.9	316.1	29.6°CW	Goldfarb et. al. (1988)	
Southern Mojave Desert									
Morongo Basin	Felsic, Mafic	30	320.2	23.6	8.4			Hopson (1996)	
Ord Mountains	Felsic	30	324.3	10.9	3.9			Diblee (1964a)	
W. Ord Mountains	Felsic	26	318.2	16.3	6.3			Diblee (1964a)	
Stoddard Range	Felsic	93	318.5	12.4	2.5			Diblee (1960)	
Eastern Transverse Ranges									
E. Pinto Mountains	Felsic, Mafic	217	343.3	23.6		3.1	301.9	41.4°CW	Howard (2002), Hope (1966)
W. Pinto Mountains	Felsic, Mafic	8	316.3	23.8		16.5	274.9	41.4°CW	Hopson (1996)
Eagle Mountains South	Felsic	81	343	8.0		1.7	301.6	41.4°CW	Powell (1981), Powell et. al. (1984a)
Eagle Mountains North	Felsic	19	50.6	16.2		7.8	9.2	41.4°CW	Powell (1981), Powell et. al. (1984a)
Chuckwalla Mountains	Felsic, Mafic	53	30.2	34.6		9.3	349.2	41.4°CW	Powell et. al. (1984b)

dominant northwest strike. Dikes within the Sierra Nevada showed no rotation with the exception of dikes at Mount Goddard. Dikes within the ranges east of the Sierra Nevada—the Benton, White, Inyo, Argus, Coso, El Paso ranges and Spangler Hill—also showed no rotation. Within the Mojave block, dike orientations were variable with an overall northwest strike; however, several localities indicated a northeast strike (Table 3). In the eastern Transverse Ranges, one domain showed a northwest strike while two other domains had a consistent northeast strike. Overall, dike orientations and paleomagnetic declinations support a clockwise tectonic rotation. Hopson et al. (2008) concluded that regional deviations of the dikes may be important indicators of possible tectonic rotations or of deviations in stress orientations during emplacement. Early Miocene rotations may account

for dispersions in dike orientations and the clockwise rotations are consistent with right shear within the eastern California shear zone.

In Summary

Research has shown that the Independence dike swarm is a useful geographic and geologic marker unit in helping to reconstruct the regional tectonic history of the Sierra Nevada and Mojave Provinces. Early studies were not based on plate tectonic theory, whereas the later studies were. Utilizing the dike swarm to examine plate convergence and extensional events during the Jurassic was a common thread in many of the referenced papers. There are many more relevant research papers; this article only summarized a few. It is certain that the list of papers focusing on and referencing the Independence dike swarm will

continue to grow. For now, the present-day Independence dikes will likely continue to be a significant key to the past.

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Paleozoic section at Mazourka Canyon, Inyo County, California: day trip guide

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Our 42.6 mile route (round trip) will traverse through Mazourka Canyon; a major synform located along the western side of the Inyo Mountains just outside Independence, CA. Mazourka Canyon, unlike most canyons, does not run directly into the Inyo Mountain crest at a right angle, but turns 90° and runs parallel to the mountains. In order to truly get a feel for the depositional history of the Paleozoic rocks in this region, an entire day should be dedicated to this trip. Extra time should be allowed if you plan on fossil collecting. 4WD is not necessary for the main trip, but it is recommended for many of the side roads. There are several spots within the canyon for dry

camping if you choose to make this a weekend outing: Santa Rita Flat, Barrel Spring, and Badger Flat.

The Paleozoic units of Mazourka Canyon show the transition from deep water deposition at Vaughn Gulch to shallow water deposition at Badger Flat. Many of the units contain fossils. Fossil preservation is variable due to intrusion of the adjacent Santa Rita Pluton. An erosional unconformity at Vaughn Gulch separates the deep water lithologies from the shallow water lithologies. However, as one drives up Mazourka Canyon, the transition from deep water deposition to shallow water deposition is evident. A modified stratigraphic column (Table 1)

Table 1: Generalized Columnar Section of Paleozoic Rocks in Mazourka Canyon and Independence Quadrangle, Inyo County, California (Modified from Ross, 1966)

Age	Formation	Thickness (feet)	Lithology
Mississippian	Rest Spring Shale	2500	Shale and siltstone, dark-grey, commonly weathers reddish brown; contains minor sandstone. Much of unit is metamorphosed to andalusite hornfels
	Perdido Formation	600	Clastic rocks, fine to coarse-grained, lenticular and variable. Chert is abundant as granule-to boulder clasts; quartzite fragments common. Calcareous quartz sandstone lenses abundant. Dark-reddish-weathering bold outcrops diagnostic
Silurian and Devonian (?)	Sunday Canyon Formation	683	Argillaceous siltstone and mudstone, dark-grey, interbedded with black chert, minor limestone, Argillaceous limestone, dark -grey to blue grey, yellow flaggy weathered slopes, minor to abundant coral-rich layers, medium grey calcareous shale, thinly bedded, <i>Monograptus</i> bearing horizon
	Vaughn Gulch Limestone	1518	Silty, dark grey thinly bedded limestone containing chert, corals near top, dark blue grey limestone, laminated, Argillaceous limestone, platy fracture, Bioclastic limestone rich in coral fragments, some yellow outcrops in the argillaceous limestones
Upper Ordovician	Ely Springs Dolomite	342	Greyish-black to medium grey to dark grey dolomite, well bedded to massive, black chert nodules, massive black chert north of Badger Flat.
Middle Ordovician	Johnson Spring Formation	210	Mixed lithology of interbedded quartzites, dolomites and limestones. Quartzite is white exhibiting some crossbedding, dolomites and limestones are dark grey, some corals and other fossils present in the middle limestone member. Exposures limited to Mazourka Canyon
	Barrel Spring Formation	157	Upper shale layer weathers a distinct red brown, fossils near base, medium grey limestone with minor siltstone interbeds, Impure quartzite interbedded with limestone
	Badger Flat Limestone (upper formation of the Mazourka Group)	590	Blue-grey Limestone, irregularly bedded calcarenite, with brown weathering silty lenses. Some thin quartzite beds, abundant fossils, some black chert
Lower Ordovician	Al Rose Formation (lower formation of the Mazourka Group)	400	Shale, mudstone and lesser limestone, red brown to brown weathering, <i>Graptolites</i> near top
Upper Cambrian	Tamarack Canyon Dolomite	910	Medium grey massive to thinly bedded dolomite, laminated, some black chert nodules
	Lead Gulch Formation	300	Limestone, siltstone, dolomite, chert and shale interbedded in thin layers, Brown weathering, <i>Trilobites</i> near base
Lower Cambrian	Bonanza King Formation	2823	Medium to dark grey to yellow grey dolomite, massive, poorly bedded to irregular thinly bedded, banding common throughout, abundant to moderate <i>Girvanella</i> as well as fucoids and other wormlike traillike markings in basal dolomite

summarizes the formations you will see in Mazourka Canyon. With the exceptions of the Lead Gulch and Al Rose formations and the Tamarack Canyon Dolomite, a more detailed description of lithologies will be found in the road log.

(0.0) Intersection of Hwy 395 and Mazourka Canyon Rd southern edge of Independence: [UTM coordinates: 11393287E / 4073044N]. Turn east onto Mazourka Canyon road.

Paiute Monument (Winnedumah) can be seen in the distance on the crest of the Inyo Mountains. It is a gigantic spire of granitic rock that juts upward. According to Paiute legend, medicine man Winnedumah was turned into this pillar of stone during a battle with invading Indians.

(2.0) (2.0) The road crosses the 233 mile long Los Angeles aqueduct. Construction of the aqueduct took place between 1905–1913. Los Angeles purchased the water rights and ultimately was able to gain complete control of Owens Valley.

(3.3) (1.3) The road crosses a 15-foot-high fault scarp from the 1872 Lone Pine earthquake. Seismic data indicates that the deepest part of the Owens Valley lies between this fault scarp and the face of the Inyo Mountains to the east (Stevens 1991).

(3.8) (0.5) The road dips down where it crosses the Owens River.

(4.4) (0.6) The paved road ends but the graded Mazourka Canyon road continues. As one looks north and south you can see the old roadbed of the Carson and Colorado Railroad (later the Southern Pacific), which served Owens Valley between 1883 and 1959. Kearsarge train station, named for the Civil War Union ship USS Kearsarge, used to stand here. The rails were removed during the 1960s and the old train station was dismantled.

(4.9) (0.5) Outcrop on south side of Mazourka Canyon Road. [UTM Coordinates 11400941E / 4074332N] This outcrop is part of the undifferentiated Owens Valley

Group/Keeler Canyon Formation (Ross, 1966). The lithology here consists of a reddish brown pebble conglomerate and light olive-green quartzite. The pebble clasts within the conglomerate are quartz.

(6.1) (1.2) Turn off the main gravel road onto a dirt road and continue east.

(6.3) (0.2) Turn off to Vaughn Gulch [UTM Coordinates 11402914E / 4075591N]. Follow the dirt road northeast toward the face of the canyon. Several roads crisscross through this area; however, if you stay on the most well-defined you'll reach the end of the road which deadends at the main wash. The road is rough and crosses a side wash that branches off of the main wash so drive with caution. High clearance and 4WD is recommended. Note: there is only room for two or three vehicles to turn around at the mouth of the gulch.

(7.0) (0.7) Entrance to Vaughn Gulch. The following describes what will be seen as you walk up canyon. Walking through the Paleozoic units at Vaughn Gulch will take some time and involves a bit of climbing. However, it is worth stopping to look at the entrance to the canyon, since the sediment-gravity-flows here are spectacular. In addition, the corals and ribbon chert within the Vaughn Gulch and Squares Tunnel are worth examining.

The Paleozoic section at the opening to Vaughn Gulch shows the transition from shallow water to



Figure 1: Coral replaced by chert in Vaughn Gulch limestone, northern side of Vaughn Gulch, Inyo Mountains, CA. Photo by Suzanne Baltzer



Figure 2: Debris flow brecciation within the Squares Tunnel limestone on the south side of Vaughn Gulch, Inyo Mountains, CA. Photo by Suzanne Baltzer

deep water. The lithologies at Vaughn Gulch consist of Silurian Vaughn Gulch Limestone, Devonian Squares Tunnel Formation, Mississippian Perdido Formation, and Mississippian Rest Spring Shale. To the east, this transition occurs in increasingly younger Paleozoic rocks (Stevens, 1986), reflecting the continued subsidence of the western North American continental margin. The Vaughn Gulch Limestone consists of calcareous, hemipelagic rocks and limestone containing a large array of shallow-water fossils. In the middle of

canyon development (Stevens 1991). Multiple gravity flow deposits within the Vaughn Gulch Limestone and Squares Tunnel Formation are evident within Vaughn Gulch (Figure 2).

The Vaughn Gulch Limestone is overlain by the Squares Tunnel Formation. The Squares Tunnel was named by Stevens et al. (1996) for Late Devonian rocks in the western Inyo Mountains near Independence. The Squares Tunnel Formation consists of dark

grey to black argillite intercalated with calcareous sandstone. This unit contains abundant black ribbon chert (Figure 3), light-grey phosphatic blebs, and radiolarians (Stevens 1991). Conodonts and radiolarians within the formation indicate deep water deposition.

At the Vaughn Gulch locality, the Squares Tunnel Formation rests upon the erosional unconformity cut into the Vaughn Gulch Limestone. A series of sediment-gravity-flow deposits within the Squares Tunnel Formation forms the south side of Vaughn Gulch. This sequence of debris-flow deposits and radiolarian chert is a temporal equivalent of the shelfal Lost Burro Formation (Stevens 1991).



Figure 3: Ribbon chert lenses within Squares Tunnel limestone, Vaughn Gulch, Inyo Mountains, CA. Photo by Suzanne Baltzer

The Perdido Formation conformably rests above the Squares Tunnel Formation.

The Mississippian Perdido Formation consists of four major units at Vaughn Gulch. The basal unit is a boulder conglomerate. The clasts consist of either quartzite or carbonate. In the higher reaches of the canyon, along the north side, a large boulder contains abundant crinoid columns. This boulder is believed to have been derived from the Hidden Valley Dolomite (Stevens 1991). The second unit within the Perdido is a quartzose siltstone, containing substantial amount of detrital pink garnet. The third unit is a limestone composed of large numbers of turbidite structures. The upper unit consists of shale gradational to the overlying Rest Spring Shale. The upper shale unit is characterized by well developed channelized turbidites and *Nereites* facies trace fossils. Maximum thickness for the Perdido Formation is 632 feet near Vaughn Gulch (Ross, 1966). The Perdido Formation thins to the south near Cerro Gordo where it is referred to as Mexican Spring Formation (Stone, 2004). A characteristic feature of the Perdido is the abundance of black chert fragments within the coarser clastic rocks.

At Vaughn Gulch the Rest Spring Shale conformably overlies the Perdido Formation. This formation is a sequence of black argillites locally interbedded with quartzose siltstone. The base of the Rest Spring Shale lies above the highest placed turbidite in the Perdido Formation in this locality.

Return to vehicles and Mazourka Canyon Road

(7.7) (1.4) Turn right at Mazourka Canyon Road.

(8.5) (0.8) Side road to the right goes to the Whiteside Mine. [UTM Coordinates 11403053E / 4076664N]

Mary DeDecker researched this mine and suggested that the entire operation was a fraud (Mitchell, 2003). The underground workings are inaccessible as the portal has collapsed and many of the supporting timbers have been burned.

(9.8) (1.3) Side road to the right goes to the Alhambra Mine, Black Eagle Mine, and the Betty Jumbo Mine at the head of Bee Springs Canyon. The climb is steep and 4WD is a must. Gold was found at the Alhambra and Black Eagle Mines. Ore deposition occurred as a result of intrusion of quartz monzonite into the pre-existing sedimentary rocks of the Tamarack Canyon Dolomite and Barrel Spring Formation. The geology is complicated by numerous faults in the area. The cabin at the Alhambra Mine remains in decent shape

and could still be used as shelter in a storm. The Betty Jumbo Mine, several miles farther up the canyon, lies within contact-metamorphic rocks where gold, silver, and scheelite (tungsten ore) were found.

(10.4) (0.6) Enter the Inyo National Forest. The dark grey dolomite on the east side of Mazourka Canyon Road between the Whiteside Mine and Squares Tunnel is the Cambrian Bonanza King Formation. The Bonanza King is a distinctive unit found throughout the southwestern Basin and Range province. In the Inyo Mountains, the formation is almost exclusively dolomite. The Bonanza King is easily recognizable in the field by its conspicuous color banding. The alternating bands are variable shades of light and dark grey. The total measured thickness in the Inyo Mountains is approximately 3,000 feet.

(10.9) (0.5) Squares Tunnel lies to the right of the road bed. This is the type locality for the Squares Tunnel Formation. The Mississippian Rest Spring Shale lies on the left side of the road. The Rest Spring Shale is the predominant rock type seen along almost the entire length of Mazourka Canyon. Normally it is described as a medium dark grey to black shale, mudstone, and siltstone with some minor sandstone and limestone. However, in Mazourka Canyon, the Rest Spring Shale has been metamorphosed to a hornfels by intrusion of the Santa Rita pluton during the Jurassic. Andalusite and spotted hornfels are the most common rock type in Mazourka Canyon.

(13.7) (2.3) Barrel Spring is found on the left at the mouth of Water Canyon. This is the type locality for the Barrel Spring Formation (Phleger 1933). The

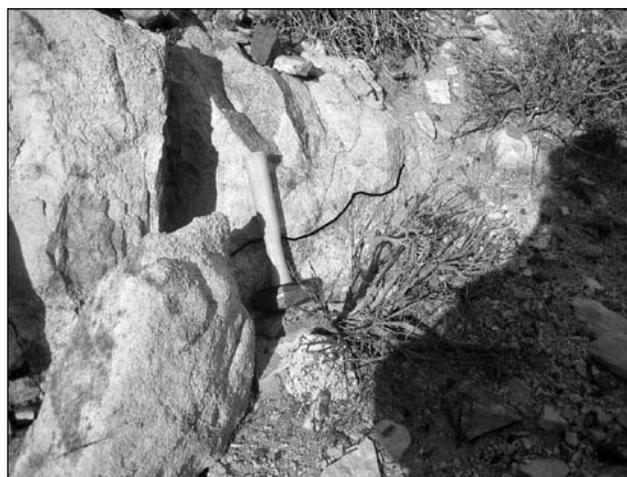


Figure 4: "Chilled" Margin in Keeler Canyon Formation limestone. The top of the chill zone is defined by line. The sledge rests upon the base of the chilled margin. Photo by Suzanne Baltzer

Barrel Spring Formation is a succession of quartzites, impure limestones and argillaceous shales of Middle Ordovician age. It crops out as a thin but widespread unit. In Mazourka Canyon, the Barrel Spring Formation is offset along strike by numerous cross faults. The stratigraphic thickness of the Barrel Spring is between 70 and 157 feet (Ross, 1966). The Barrel Spring Formation forms the red-brown rubbly slopes within the canyon. Brachiopods, trilobites, bryozoans, and graptolites are present in the Barrel Spring Formation, although they are not common. Many of the fossils are poorly preserved or have been silicified by intrusion of the Santa Rita pluton.

(14.2) (0.5) The side road to the right travels up Mexican Gulch to Johnson Spring. This particular road is classified a Class III 4WD road, meaning that high clearance and 4WD with limited slip differential is required (Mitchell, 2003). The type section for the Johnson Spring Formation is found here. The Johnson Spring Formation is a mixed sequence of quartzites, dolomites, and coral-bearing limestones. A crinoidal dolomite occurs in the lower part of the Johnson Spring Formation. This dolomite unit contains abundant pelmatozoan fragments along with some brachiopod and coral fragments. In addition, the coral-bearing limestone unit is located just north of the type section at Johnson Spring. However, the characteristic rock type of this formation is a white to grey orthoquartzite. The Johnson Spring Formation is considered a lateral equivalent to the widespread

Eureka Quartzite. Exposures of this formation are limited to Mazourka Canyon.

(14.3) (0.1) Turn left on Forest Road 13S05A. [UTM Coordinates 11403082E / 4081586N] This road takes you through a breach of the Pennsylvanian Keeler Canyon Formation. Shortly after you turn onto this road you will recognize the contact between the Rest Spring Shale and Keeler Canyon Formation; the Rest Spring Shale has a redder appearance. The Keeler Canyon Formation is a thin-bedded, medium to dark grey, impure, silty, and arenaceous to pebbly limestone and limy siltstone, with intercalations of pink or maroon fissile shale. Bedding is inclined to be platy or flaggy. It rests conformably upon the Mississippian Rest Spring Shale.

(14.9) (0.6) Contact between Jurassic Santa Rita pluton and Keeler Canyon Formation. [UTM Coordinates 11402696E / 4081981N] The thinly bedded light grey limestone and dark grey shales in this area contain excellent examples of chilled margins of the contact between the sedimentary units and the quartz monzonite of the adjacent pluton. These chilled margins form as the pre-existing sedimentary rocks are “cooked” by the pluton. As the rock cools, the sedimentary rock is literally fused with the underlying plutonic rock forming a “chilled” margin at the contact (Figure 4).

(15.3) (0.4) Santa Rita Flat [UTM Coordinates 11402139E / 4082117N]. The Santa Rita pluton is a quartz monzonite intrusive located along the western side of the Inyo Mountains. This pluton was one of many emplaced during the Jurassic Nevada Orogeny. It was also during this time that the Independence dike swarm was emplaced (see Baltzer, this volume). Mazourka Canyon is a large syncline and the units along the west limb were deeply metamorphosed as a consequence of pluton emplacement. The sedimentary units of the east limb also experienced metamorphic alteration, but to a lesser degree. The quartz monzonite of the pluton is coarse grained, containing large K-spar phenocrysts. As you drive across Santa Rita Flat,



Figure 5: Looking north up Mazourka Canyon at the Paleozoic stratigraphy. Photo by Suzanne Baltzer

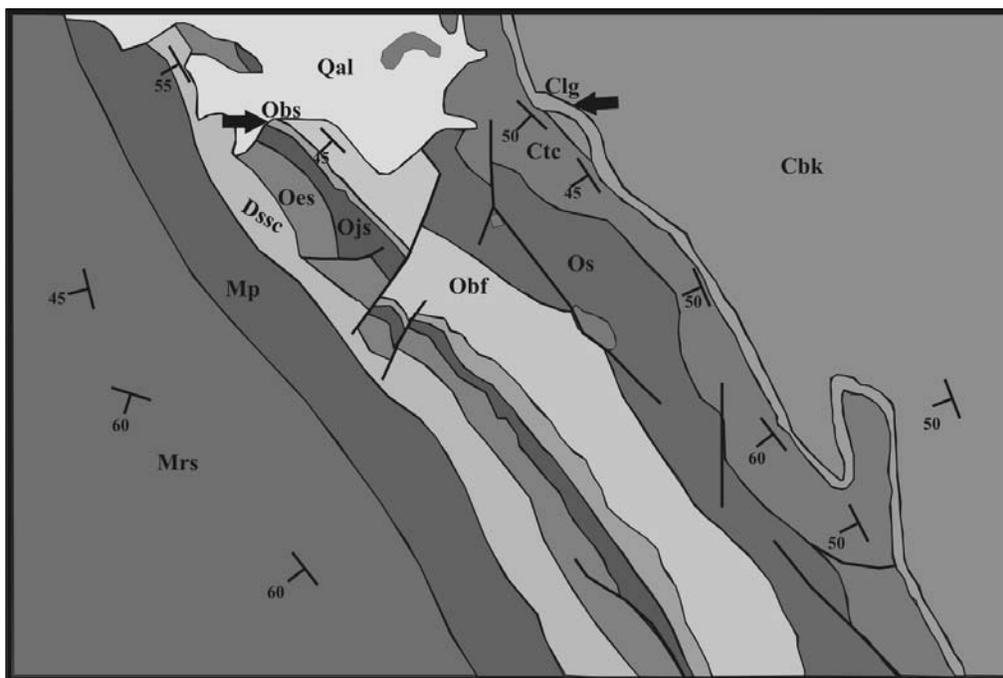


Figure 6: Simplified Portion of the Geologic Map of Mazourka Canyon Area (Ross, 1966) south of Badger Flat showing cross faults. Qal, Quaternary Alluvium; Mrs, Mississippian Rest Spring Shale; Mp, Mississippian Perdido Formation; DsSc, Silurian/Devonian Sunday Canyon Formation; Oes, Ordovician Ely Springs Dolomite; Ojs, Ordovician Johnson Spring Formation; Obs, Ordovician Barrel Spring Formation; Obf, Ordovician Badger Flat Limestone; Os, Ordovician Al Rose Formation; Ctg, Cambrian Tamarack Canyon Dolomite; Cbk, Cambrian Bonanza King Dolomite. (Note: map not to scale.)

note the spheroidal weathering pattern of the plutonic rocks. This is typical of granitic rocks as they weather. Turn vehicle around and return to Mazourka Canyon Road. (Note: one can continue on across the flat and head back to Mazourka Canyon Road via Pop's Gulch, but the road is much rougher).

(16.7) (1.4) Turn left back unto Mazourka Canyon Road. On the east side of the canyon, excellent exposures of a portion of the stratigraphic sequence in Mazourka Canyon can be viewed (Figure 5).

(17.6) (0.9) The side canyon on the left is Sunday Canyon. Shortly after passing Sunday Canyon there is a branch in the main road (~0.6 miles). Stay to the right on the main road.

(18.5) (0.9) The turn off to the left goes through Pop's Gulch. This is the back route to Santa Rita Flat. The road is much rougher than Forest Road 13S05A as it travels through a narrow steep wash. The route takes you through Al Rose Canyon. The road initially travels through Rest Spring Shale then passes a thin band of Perdido Formation. Beyond the contact between the Rest Spring Shale and Perdido, the route crisscrosses between Perdido Formation, Sunday Canyon

Formation, and Ely Springs Dolomite.

(21.1) (1.8) Fault block of Sunday Canyon Formation (?) against Ely Springs Dolomite (?). [UTM Coordinates 11403380E / 4083997N] Note the extreme fracturing and shearing of the light yellow-orange-brown shales of the Sunday Canyon Formation (Figure 6). This shearing occurred along one of the many unnamed cross faults in Mazourka Canyon. The Silurian Sunday Canyon Formation is laterally correlative with the Vaughn Gulch Limestone. The type

section is at Sunday Canyon, the small tributary canyon you passed earlier. The Sunday Canyon Formation is described as a calcareous siltstone, calcareous shale and argillaceous shale (Ross, 1963). In outcrop, the Sunday Canyon tends to be thinly bedded and weathers poorly on exposed slopes which are littered with thin shaly to flaggy fragments. These fragments are typically light grey to shades of yellow to orange. The Sunday Canyon Formation is highly fossiliferous, containing *Monograptus* graptolites which have been collected from several localities. These graptolites are considered to be restricted to the Silurian; however, some Devonian monograptids are also present.

The Ordovician Ely Springs Dolomite forms a thin discontinuous band along the length of Mazourka Canyon just north of Squares Tunnel to Badger Flat. The Ely Springs Dolomite conformably overlies the Johnson Spring Formation. The contact is abrupt, placing quartzite against dolomite. The Ely Springs Dolomite is predominantly a dolomite. Distinct dark and light grey banding of the dolomite allows it to be easily recognized in the field. At the north end of the outcrop belt, northwest of Badger Flat, the Ely Springs Dolomite grades into massive black chert (Ross, 1966).



Figure 7: Looking northwest from Mazourka Peak. The Sierra Nevada crest and impending storm clouds on the horizon. Photo by Suzanne Baltzer

(23.5) (2.4) The road climbs out of Al Rose Canyon and crosses Badger Flat. The outcrop of bluish-grey rock on the left is the Ely Springs Dolomite. Badger Flat is a high mountain valley at an elevation between 8,700 and 9,000 feet. The Badger Flat Limestone of Ordovician age was named after its type section locality at Badger Flat in the Independence quadrangle (Ross, 1963). Badger Flat Limestone is a shallow water blue-grey limestone with minor interbeds of light-grey-, orange-, and red-brown-weathering siltstone. It is highly fossiliferous, containing a variety of shallow water fossils. These fossils include gastropods as much as 3 inches across, brachiopods, trilobites, cephalopods, and corals. Pelmatozoan fragments are also abundant. Fossil assemblages indicate shallow water deposition interpreted to be part of a shelf environment (Stevens, 1991). Measured thickness of Badger Flat Limestone is between 511–586 feet (Ross, 1966). The Badger Flat Limestone at Badger Flat is the blue-grey limestone just above the red-brown Barrel Spring Formation. Follow the main road to the right to reach Mazourka Peak.

(25.5) (2.0) Mazourka Peak [UTM Coordinates 11400007E / 4092794N]. The summit of Mazourka Peak is elevation 9,400 feet. Even though the peak itself supports several antenna sites, the views of the Sierra Nevada crest and Owens Valley are breathtaking (Figure 7). On a clear day one can see for approxi-

mately 30 miles in either direction. For geologists, the geomorphic features of the valley floor are quite dramatic from this viewpoint. Mt. Williamson (elevation 14,375') can be seen to the south and Mt. Tom (elevation 13,652') can be seen to the north. (The view of Mt. Whitney is blocked by Mt. Williamson). Total distance to the peak from Independence is 21.3 miles. Return to vehicles and head back to Independence via the same route.

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Lower Cambrian–Precambrian stratigraphy and fossil fauna of the Westgard Pass area

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The succession of Lower Cambrian strata exposed in the White and Inyo Mountains and adjacent areas east and northeast represents one of the thickest accumulations of strata of this age in North America (Nelson and Perry, 1955). The type section at Waucoba Spring was originally described by Walcott (1908) (Figure 1). The 21,000 feet of Lower Cambrian and Precambrian strata are divided into eight formations, with the upper 12,000 feet of the succession lying above a significant regional unconformity. A generalized stratigraphic column is shown in Table 1.

The oldest rocks at Westgard Pass are part of the Precambrian Wyman Formation. The rocks of the Wyman formation lie 4000 feet below the oldest trilobite bearing zone of the Cambrian Campito Formation. The type section locality is Wyman Canyon in the Blanco Mountain quadrangle (Maxson, 1935). The formation has a minimum thickness of 9000 feet with no stratigraphic base exposed. The Wyman Formation is typically thin-bedded brown to dark grey argillite, quartzitic sandstone, siltstone and interbedded grey-blue limestone. The limestone beds are frequently dolomitized. The upper surface of the Wyman succession represents a regionally significant unconformity.

The Reed Dolomite was named by Kirk (*in* Knopf, 1918) for exposures at Reed Flat in the Blanco Mountain quadrangle. The formation is entirely massive dolomite, ranging from coarsely crystalline to aphanitic. It is typically white or slightly bluish, except it becomes very light yellow brown where it is weathered. Locally, oolitic lenses are present. Measured thickness of the Reed Dolomite is close to 2000 feet. In the northern part of the Bishop quadrangle, the Reed dolomite forms the axis of a large southward plunging anticline which disappears under the Campito sandstone. The Reed Dolomite forms a distinct long white ridge that stands out sharply against the darker basal sandstones in the Blanco Mountain quadrangle. Blanco Mountain is composed entirely of the Reed Dolomite and can be seen for miles. The only fossils present were poorly preserved forms that suggested *Girvanella*, a calcareous alga.

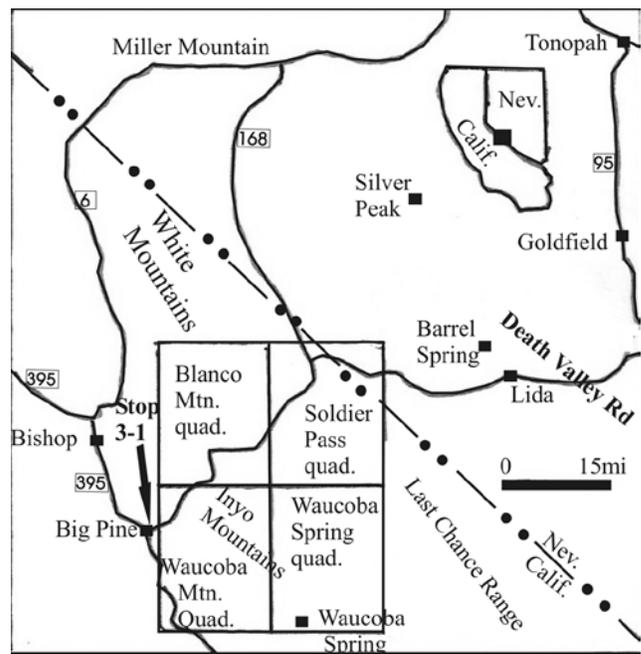


Figure 1: Index Map of Westgard Pass Area and Surrounding Vicinity (Modified from Nelson, 1962)

The Deep Spring formation unconformably overlies the Reed Dolomite and is overlain by the Campito Sandstone. The Deep Spring Formation is a series of interbedded sandstones and limestones with some minor shale horizons. The typically blue-grey to white sandstones are dominant in the upper part of the formation. Limestones are more dominant in the lower section, and are massive and pale yellow-white to dirty white in color. The shales, where found, are arenaceous with interbedded quartzitic sandstone. Measured thickness is approximately 1,600 feet.

The Campito Formation is about 3,500 feet thick and is broken into two members, the Andrews Mountain member and the Montenegro member. The lower Andrews Mountain member has a thickness of 2500–2800 feet and consists of massive, strongly cross-bedded, dark-grey to black quartzitic sandstone containing some grey siltstone and shale interbeds. When weathered, the sandstone becomes reddish-brown to deep purplish red. The massive sandstone is highly jointed, causing the sandstone to break down into angular blocks. A middle horizon in the An-

Table 1: Generalized Stratigraphic Column for the Westgard Pass Area

Age	Formation	Thickness (feet)	Lithology
Lower Cambrian	Mule Spring	1000	Massive to well-bedded dark-blue to grey limestone. Some interbeds of grey shales. Locally, some dolomitization of the limestone is present. <i>Girvanella</i> common.
	Saline Valley	850	Upper member: Light to medium brown quartzitic sandstone, grey limestone, and grey-green to black silty shales. Upper 10ft contains trilobite genera <i>Paedeumias</i> , <i>Bristolia</i> , and <i>Olenellus</i> . Lower member: Medium to coarse grained quartzitic red-brown sandstone. Locally becomes a vitreous quartzite. Thin blue-grey arenaceous limestone caps the underlying sandstones and quartzites.
	Harkless	2000	Grey-green shales and thin-bedded to platy fine-grained quartzitic sandstone and siltstone. Massive vitreous quartzite interbedded with shales. Quartzite contains <i>Scolithus</i> . Pisolithic limestone present locally. Contains the archeocyathid <i>Coscincyathus</i> , as well as some brachiopods and <i>Salterella</i> , a mollusk.
	Poleta	1300	Upper member: (200ft) Non-fossiliferous blue-grey limestone. Middle member: (500ft) Grey-green shales containing trilobite genera <i>Judomia</i> , <i>Fremontia</i> , <i>Laudonia</i> , <i>Nevadella</i> , <i>Holmia</i> and <i>Olenellus</i> . Also contains echinoderms. Lower member: (600ft) Massive blue-grey limestone with minor shale horizons. Contains 21 archeocyathid genera including <i>Archeocyathus</i> , <i>Protopharetra</i> , <i>Pycnoidocyathus</i> , <i>Cambrocyathus</i> , <i>Ajacycyathus</i> , <i>Rotundocyathis</i> , <i>Annulofungia</i> , and <i>Ethmophyllum</i> .
Precambrian to Lower Cambrian	Campito	3500	Montenegro member: (2500ft) Green-grey shales, fine grained quartzitic siltstones and sandstones. Minor limestone in the upper 200 ft. Contains trilobites genera <i>Daguinaspis</i> , <i>Fallotaspis</i> , <i>Nevadia</i> , <i>Holmia</i> , <i>Holmiella</i> , and <i>Judomia</i> . Also contains archeocyathids: <i>Ajacycyathus</i> , <i>Rotundocyathis</i> , <i>Ethmophyllum</i> , and <i>Pycnoidocyathus</i> . Andrews Mountain Member: (1000ft) Massive, strongly cross-bedded dark-grey to black quartzitic sandstone with some grey siltstone and shale interbeds. Upon weathering the sandstone becomes a deep reddish-brown to a deep purplish red. Middle section of this member contains the trilobite <i>Fallotaspis</i> . The trilobite zone marks the boundary between the Precambrian and Cambrian Periods.
	Deep Spring	1600	Upper Part: Blue-grey to white sandstones with minor arenaceous shale and quartzitic sandstone interbeds. Lower part: Massive pale yellow-white to dirty white limestone.
	Reed Dolomite	2000	Massive white to bluish dolomite, coarsely crystalline to aphanitic. When weathered, it becomes a tan to yellow-brown. Oolitic lenses locally present. Occasional poorly preserved <i>Girvanella</i>
	Regional Unconformity		
	Wyman	9000+	Thin-bedded brown to dark-grey argillite, quartzitic sandstone, siltstone and interbedded grey-blue limestone, frequently dolomitized.

Andrews Mountain member contains the olenellid trilobite *Fallotaspis*, (Seiple, 1984) representing the oldest trilobite horizon in North America. This horizon is considered by many paleontologists to be the base of the Cambrian Period. The Andrews Mountain units below this horizon are considered Precambrian in age. At the top of the member, thin lenticular archeocyathid limestone beds are present.

The upper Montenegro member is typically green-grey shales, fine-grained quartzitic siltstone, and sandstone with a measured thickness of 1,000 feet. Minor horizons of limestone occur in the upper 200

feet. This member also contains a larger assemblage of trilobite fauna than the lower Andrews Mountain. Trilobite genera found in the Montenegro member are *Daguinaspis*, *Fallotaspis*, *Nevadia*, *Holmia*, *Holmiella*, and *Judomia* (Nelson, 1976). In the upper limestone horizons of this member, the archeocyathid genera *Ajacycyathus*, *Rotundocyathis*, *Ethmophyllum*, and *Pycnoidocyathus* (McKee and Gangloff, 1969) are found. These marine, benthic animals were at first considered primitive corals based on their morphological similarity to some coelenterates. Later, they were placed into their own phylum, *Archeocyathia*. These ancient animals had a limited existence in time and are only found in strata of the Cambrian Period.

The Poleta Formation overlies the Campito Formation and contains a thick succession of archeocyathid bearing limestones, shales, and quartzites. The Poleta Formation is divided into three members. The lowest member has a measured thickness of 600 feet and is described as a massive grey-blue lime-

stone with abundant archeocyathids and minor shale horizons. This member contains 21 genera of archeocyathids, including *Archeocyathus*, *Protopharetra*, *Pycnoidocyathus*, *Cambrocyathus*, *Ajacycyathus*, *Rotundocyathis*, *Annulofungia*, and *Ethmophyllum*. Many of the specimens are fairly well preserved and some form local concentrations representing small reefs or bioherms. The lower member of the Poleta Formation grades into the middle member containing about 500 feet of grey-green shales. These middle shales contain several trilobite genera, including *Judomia*, *Fremontia*, *Laudonia*, *Nevadella*, *Holmia*, and *Olenel-*

lus. Six to ten species of echinoderms are also present. *Scolithus*-bearing quartzites rest upon the fossiliferous grey-green shales of the middle member. The upper member is a 200 foot thick non-fossiliferous blue-grey limestone which grades into the overlying Harkless Formation.

The 2,000 foot thick Harkless Formation consists of grey-green shales and thin-bedded to platy fine-grained quartzitic sandstone and siltstone. The formation contains massive vitreous *Scolithus*-bearing quartzite interbedded with shales. Locally, the Harkless Formation contains pisolitic limestone and archeocyathid-bearing limestone in the basal position. This limestone contains abundant large archeocyathids, genus *Coscinocyathus*, some of which are 10 inches long and 2 inches wide. Usually the archeocyathid specimens from the Campito and Poleta Formations are one-half to 4 inches long and one-eighth to one-half inch wide. Brachiopods and *Salterella*, a slender molluscan fossil around a quarter inch long, occur in some of the siltstones. The trilobite genus *Paedeumias* rarely occurs in the upper part of the Harkless Formation in the Westgard Pass area. Trilobites are more common in the Harkless Formation in Esmeralda County, Nevada.

The Saline Valley Formation is named from exposures in the Waucoba Spring quadrangle that overlook Saline Valley to the south (Nelson, 1962). The basal member is composed of medium to coarse-grained quartzitic red-brown sandstone which locally becomes vitreous quartzite. The sandstone and quartzite units are capped by a blue-grey arenaceous limestone. The upper member is a succession of brownish quartzitic sandstone, grey limestone and grey-green and black silty shales. The Saline Valley Formation is 850 feet thick. Most of the fossils are found in the upper 10 feet of a fissile olive-grey shale. Abundant trilobites of the genera *Paedeumias*, *Bristolia*, and *Olenellus* are present in these shales.

The Mule Spring Formation is a well-bedded dark blue to grey limestone measuring 1,000 feet. Locally, some of the limestones have been replaced by dolomite. There are also minor interbeds of grey shale. It contrasts sharply with the red and brown quartzites of the Saline Valley Formation. The Mule Spring Formation contains abundant *Girvanella* algal structures. In places, these algal structures make up about 40% of the rock. The Mule Spring Formation is the upper limit of the Lower Cambrian. Strata of the Middle Cambrian rests upon the Mule Spring Formation and are not discussed in this paper.

The fossil rich formations of Westgard Pass give important insight into the paleoenvironment of the region during the Cambrian and Precambrian Periods. The evidence found within the rock units fill in some of the missing pieces of Earth's history. It is an area popular for geology summer field camps due to this diversity. Spending time to examine the rocks' history at Westgard Pass is always a learning experience.

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The Olancha Dunes, Owens Valley, California

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The Olancha dunefield is a small elongate group of active reversing transverse ridges extending over 8 km², located near the community of Olancha in southern Owens Valley, California. It is bordered on the south by the Coso Mountains, on the west by the Sierra Nevada, on the east by the Inyo Mountains, and on the north by the partially dry bed of Owens Lake. The dunefield is situated on beach material along the passageway of the last overflow of Owens Lake that occurred 2,000 to 4,000 years ago. Research on this field, including the geomorphology of the dunefield and surrounding environs, the heavy-mineral content of the sand-sized grains, and current wind regimes, indicates that the immediate source of sand for the dunes is the beach material upon which the dunes lie.

The paleoenvironmental record preserved in a lake-bed core of Lake Owens reveals a history of at least eight overflow (transgression) episodes of the Owens River pluvial system during the past 800,000 years (Smith, 1993). Each overflow passed beyond the

southern end of Owens Valley, through the area where Haiwee Reservoir is currently located, and over the Haiwee Gap. Transgressions of Lake Owens

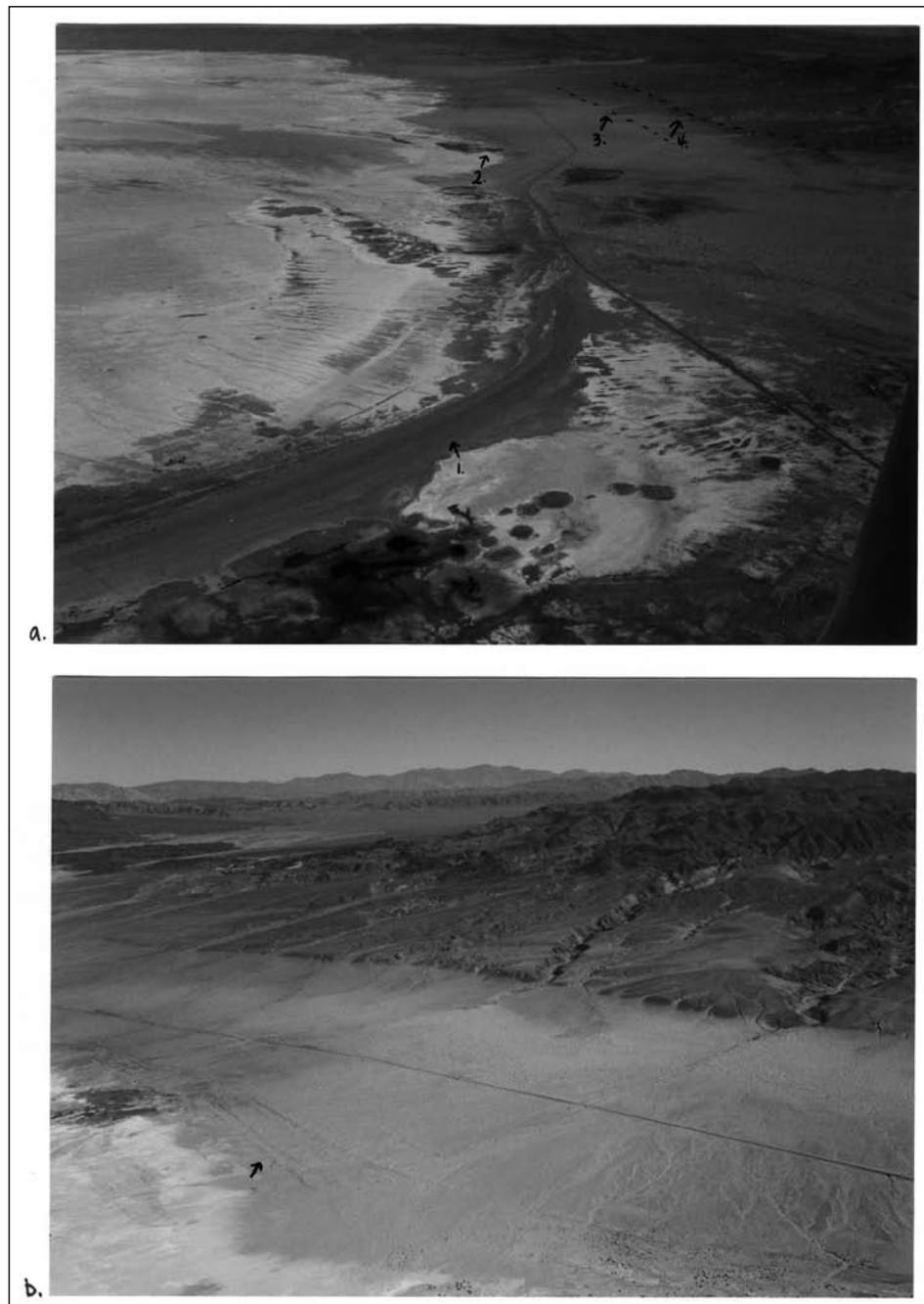


Figure 1. Pleistocene shorelines (2). Photos by author. A. Historic 1872 shoreline (Holocene). 2. Dirty Socks Spring. 3. Tioga shoreline (1144 m). 4. Tahoe shoreline (1183 m). Owens lake at the left. View to the east. B. Tahoe shoreline (1183 m) along a prominent fault scarp along the Coso Mountains. Arrow denotes historic shoreline. Centennial flat in background. View to the southeast.

This paper is derived from the Geography Masters thesis for California State University Northridge, written by the author in 1999 under her former name, Mary DeLaTorre.

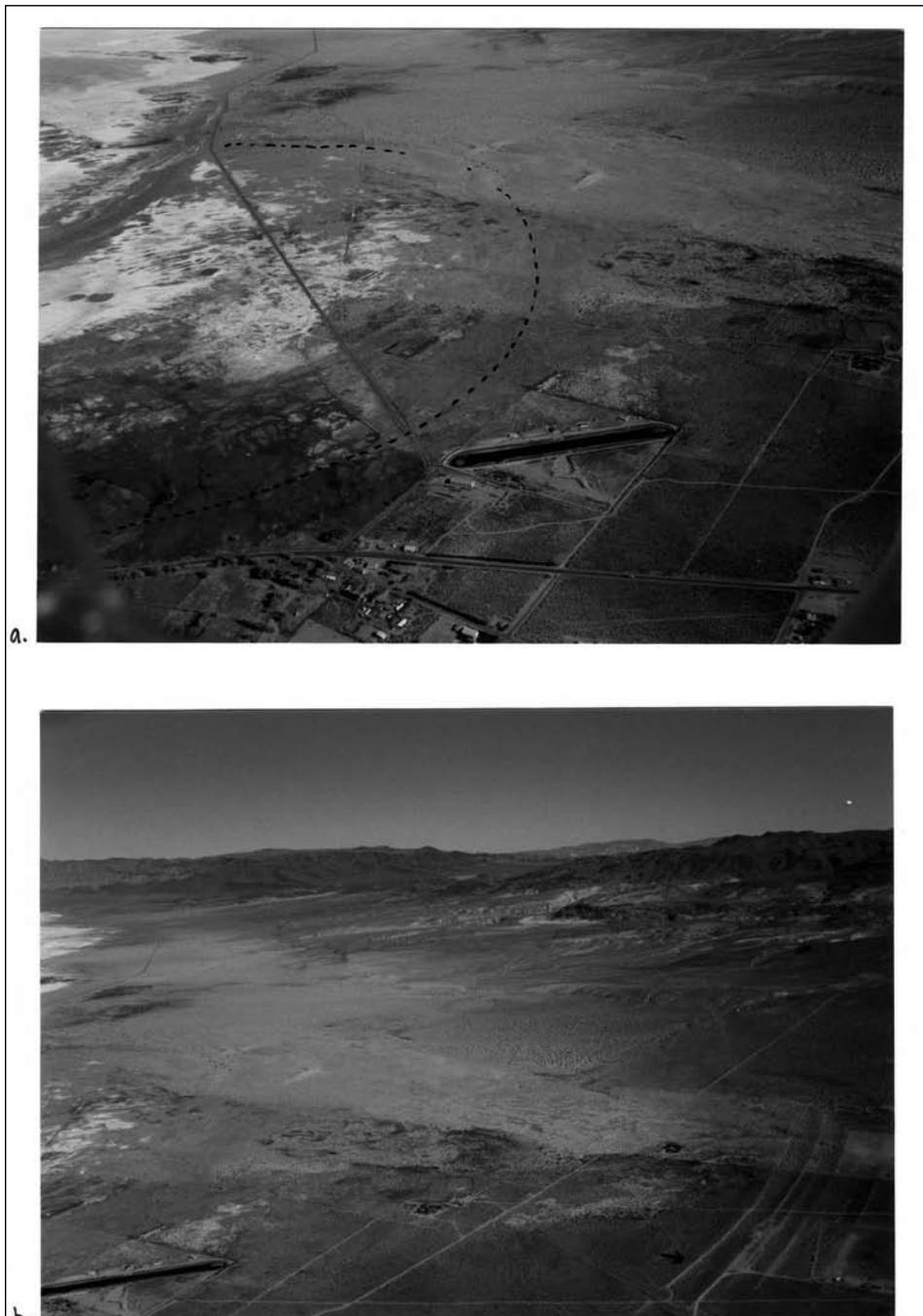


Figure 2. Holocene shorelines at the Olancha dunefield. Views to the east. Photos by author. A. Dashed line traces the obscure shorelines at 1102 m. B. Arrow denotes shorelines at the southern end of the dunefield at 1122 m.

during cold glacial conditions reworked alluvium that was deposited at low regressive lake levels during warm interglacial conditions. Addition of material and the reworking of sediment during the eight paleoclimatic fluctuations throughout the Pleistocene (1.8 Ma to 10ka) and the Holocene (10 ka to present), provided material for the 2- to 4-km wide broad, shallow beaches along the southern shore.

Lake Owens shorelines of many ages are discernible around the lake basin. Figure 1 shows the Pleisto-

cene shorelines along the south-southeast side of the basin. Those attributed to the lake's maximum highstand during the Tahoe glacial (65 ka) lie at approximately 1183 m elevation (Fig. 1 a-b). Those Tahoe shorelines are especially visible in the deeply incised pebble beaches and bar material deposited along the prominent fault scarp along the north flank of the Coso Mountains (Fig. 1b) (Carver, 1975; Saint Amand et al., 1986). Tioga glacial shorelines (20-18 ka) lie at 1144 m elevation (Fig. 1a). A ^{14}C date of $19,620 \pm 260$ BP was derived from a freshwater snail *Stagnicola* within those beach deposits (Orme, 1995).

Shorelines attributed to the final Holocene regression of the lake are clearly visible along the southern end of the dunefield at 1122 m and more obscurely discernible at 1102 m (Fig. 2a-b). The historic shoreline of 1872 is clearly visible in Figure 1a. The lake at this time (1872) was 76 meters deep. By 1926, water diversion by Los Angeles Department of Water and Power resulted in the lake's desiccation.

The Olancha dunes appear to have developed due to topographically controlled strong winds that blow both from the north and the south, hence, the east-west trending ridges of the dunes reverse seasonally. The Great Basin Unified Air Pollution Control District in Bishop gathered wind data between 1985–1997 from four sites around the lakebed. Analysis indicates that the prevailing wind at Olancha is southerly, with northerly wind

only accounting for about 4% of the total. However, the intensity of the infrequent north flow has been sufficient to balance the overall movement of sand. Other geomorphic features, such as ventifacts and older dunes along the north flanks and crests of the Coso Mountains, indicate that past wind regimes perhaps had a stronger northerly component than today.

Recent environmental concerns about Owens Lake have focused on the harmful dust emissions derived from the “salt fluff” that blows off the dry lakebed—particles so small (PM10 particulate matter <10 microns), they cause various levels of decreased lung function (Schade, 1994). Attempts to control those emissions from the crust, such as planting saltgrass, installing sand fences, and flooding portions of the lakebed, have significantly reduced the number and intensity of the dust storms.

Part of the process in determining mitigation strategies for the lakebed included researching the sources (provenance) of sand within the neighboring dunefields. This thesis studied the sources of sand in the Olancha dunes. The most reliable provenance indicators in sand are the heavy mineral grains that have a density greater than 2.96—heavier than quartz and feldspar. Field samples from 50 sites around the lakebed and its environs were gathered, then analyzed in the lab. The heavy minerals that provided the best source data for the Olancha dunes included zircon, rutile, dolomite, epidote (Coso volcanics), sphene (Cosos and Sierra Nevada plutons), and limonite, ilmenite, and magnetite—all three iron minerals which indicate plutonic sources (Cosos and Sierra Nevada). The combined evidence strongly supports the conclusion that Olancha dune sand originated from beach deposits along the southern shores that were reworked by waves, and derived from alluvium eroded from the Coso Mountains and Sierra Nevada. It is unlikely the sand was transported by wind across the lakebed from either the slopes of the Inyo Mountains or from the deltaic sediments deposited by the Owens River entering the north part of the lake.

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Triassic ammonoids from Union Wash, Inyo County, California

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For approximately the last century, since the time of original collections of ammonoid cephalopods in Inyo County by C.D. Walcott in 1896 and the publication of these fossils by J. P. Smith (1932), the Triassic marine strata (aggregate thickness of approximately 2800 feet) exposed on the western slope of the Inyo Mountains have been of intense interest to both amateur and professional paleontologists. While early studies of the Union Wash cephalopod fauna were necessarily descriptive and dominantly taxonomic in their focus, recent investigations of the taphonomy of the ammonoid remains (Woods and Bottjer, 2000; Mata and Woods, 2008) have been concerned with paleoceanographic conditions and the rebound of the early Triassic marine fauna which closely followed the Late Permian marine catastrophe. Although the Union Yes - Wash Canyon fauna was deposited on a low-lying coast of western Pangea, it is world-wide in its content, containing elements of the Boreal Ocean and Southeast Asian ammonite assemblages. Mega-fossils observed in the Union Wash section are almost exclusively ammonoids with a few broken scraps of monitid(?) bivalves.

The ammonoid fossils occur in a sequence of carbonate units with only local stringers of siliciclastic debris. Fossils tend to occur in widely separated layers within the sequence. The limestones range in color through all shades of gray, possibly reflecting varying degrees of oxygen content during deposition of the sediments. Laminations are observed sporadically throughout the darker units, again suggesting hypoxic conditions. Generally low rates of sedimentation are indicated. Moreover, Woods and Bottjer (2000) have postulated that the temporary disappearance of ammonoid conches high in the Union Wash sequence (above the *Meekoceras* Zone) may be related to a shallow aragonite compensation depth, while elsewhere in the section layers of concentrated ammonite debris may reflect anoxic events.

Ammonites from the Union Wash sequence are significant because they comprise the oldest Mesozoic molluscan assemblage in North America. The fauna dates from the Smithian–Spathian Stages of the Lower Triassic and is an exemplary sample of the

depauperate recovery (“bounce back”) faunas which are so common following the Late Paleozoic extinction event.

The Union Wash section is located on the western slope of the Inyo Mountains approximately ten miles northeast of the town of Lone Pine, California. Standard passenger cars can travel to a parking area at the mouth of Union Wash (and the first fossiliferous outcrops) with minimal discomfort but four wheel drive vehicles are necessary beyond this point. Note also that no vehicles are allowed beyond the boundary of the Southern Inyo Mountains Wilderness Area which is open only to hikers. Moreover, no disturbance of *in situ* fossil remains is allowed within the wilderness zone.

To reach the Union Canyon Wash, drive northbound on Highway 395 from Lone Pine. Turn right onto Narrow Gauge Road just before leaving town. Pass the former Southern Pacific Lone Pine Station and continue northbound on Owenyo Road past the old railroad stop of Owenyo (now largely demolished). About 1.2 miles beyond the ruins of Owenyo (and approximately eight miles from the turnoff from Highway 395) a road crosses the old canal and ascends the alluvial fan for approximately two miles. Return access to Highway 395 may be gained by continuing north on the Lone Pine/Owenyo Road, west on the Manzanar/Reward road to the Highway.

The first fossiliferous outcrops of the Triassic Union Wash Formation and the occurrence of the *Parapopanoceras haugi* Zone. The ammonoid fossils occur in dark grey, almost black, micrite and consist of molds and casts. Many of the available specimens have weathered from the outcrop and are found as fragments on the talus slope. Limited numbers of specimens, for noncommercial purposes, may be picked up from the surface.

At the branch in the wash, the east fork (topographically higher but stratigraphically lower) leads in about one half mile to the *Meekoceros gracilitatus* fauna (Zone), a worldwide indicator of Lower Triassic age. Again, the fossils are dark grey to black in color but here occur in a lighter gray, sometimes laminated and somewhat silty limestone and are more easily

recognized than the conches from the *Parapopanoceras* horizon. Although the fossils are easier to find, they may not be excavated because they lie within the Southern Inyo Mountains Wilderness Area.

The rocks of the *Meekoceras graciliatatus* Zone contain abundant cephalopods. James Perrin Smith (1932) reported, among other ammonoids, the occurrence of *Anasiberites*, *Juvenites*, *Owenites*, and *Xenodiscus* in the limestones of Union Wash, again attesting to the diversity of age-diagnostic ammonites along the shallow marginal sea of western Pangea.

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Archaeocyathids from Westgard Pass, Inyo County, California

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A classic locality for the collection of Lower Cambrian archaeocyathid fossils may be visited in Westgard Pass, Inyo County, California. Minimal, noncommercial collecting is permitted. The location occurs in a tan, orange-weathering, somewhat silty limestone interbedded with green shale. The Cambrian rocks outcrop along Deep Spring Valley Road about eleven miles from its junction with U.S. Highway 6 at Big Pine, California. The fossils are difficult to observe in unweathered matrix and many are partially destroyed due to post-depositional recrystallization (Okulitch, 1954).

The archaeocyathid remains occur in the middle part of the Middle Member of the Poleta Formation of Lower Cambrian age. This unit has a larger siliciclastic component than the otherwise carbonate shelf environment which comprises the bulk of the Poleta Formation (Dornbos and Bottjer, 2001). Associated fossils are trilobite fragments and disarticulated plates of the enigmatic echinoderm, *Helioplacus*, which first was described from this area by Durham and Caster (1963). The archaeocyathid fossils generally are small, rarely over three centimeters in length. Deeply weathered specimens show the characteristic double wall pattern (“cup within a cup”) which is particularly evident in slabbed and polished sections. The better specimens are found in talus slope material where weathering has enhanced their skeletal features. Typically, archaeocyathids are cup shaped—hence the name “ancient cups.” Some may be confused with small rugose corals because they exhibit short septae between the inner and outer walls of the cup. This condition is most apparent in weathered specimens.

The archaeocyathids comprise a problematic group which over the years has been placed among the sponges (pliosponges), rugose coral ancestors, and, more recently, a group (phylum) of their own, the Archaeocyatha. They are an example of a major taxonomic unit which has originated and then become extinct within the course of a single geologic time period—the Cambrian. Indeed, they are the hallmark of Lower Cambrian time although archaeocyathids persist into the Middle Cambrian in some areas. There is a disputed report of a single species of

archaeocyathid from alleged Upper Cambrian rocks in Antarctica.

Some workers consider the archaeocyathids to be the world’s oldest “reef-building” animals although that title is contested by sponges. There is evidence for sponge–archaeocyathid consortia in the Lower Cambrian but the two groups also appear to have built segregated bioherms later in Cambrian time. Both groups appear to have served as framework organisms. Archaeocyathids commonly are associated with carbonate sedimentation, but seem to tolerate some siliciclastic component, particularly in the silt-size particle range. This seems to be particularly true of their environment as exposed in Westgard Pass.

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Basaltic volcanism in the southern Owens Valley, California

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Abstract: Two Neogene basaltic volcanic fields lie within the southern Owens Valley, California. Each is characterized by a distinct petrology and geochemistry. The basalts of the Dove Spring Formation of the Ricardo volcanic field were emplaced between 8 and 12 Ma, and are comprised dominantly of quartz-normative tholeiitic basalt with interbedded rhyolite tuff. Olivine has undergone partial to complete alteration to iddingsite and calcium-siderite. The Coso field (<4 Ma) is bimodal, consisting of both basalt and felsic tuff in equal amounts. The basalts are generally alkaline, containing large phenocrysts of unaltered olivine.

The presence of iddingsite, a hydrous alteration product of olivine, in Dove Spring basalts suggests a complex evolutionary path for the magma. Mantle-derived magma rose along faults to shallow depths in the crust where it became gravitationally stable and ponded. Heat from the magma caused the surrounding rocks to partially melt, altering both oxygen fugacity and chemistry. Increased oxygen fugacity led to instability of olivine and formation of iddingsite, while changes in chemistry converted magma composition to that of a silica-saturated tholeiite. Reactivation of the faults resulted in breaching of the overlying rock units, providing a conduit for tholeiitic magma to reach the surface.

Dove Spring basalt was extruded during the initial stage of Basin and Range extension when rates of vertical crustal displacement may have been 0.1–0.2 mm/yr (Loomis and Burbank, 1988). In contrast, the Coso volcanic field formed in a transtensional basin created by Recent dextral shear (Monastero, 2005). Estimates of right-slip motion vary from 3 to 8 mm/yr, an order of magnitude greater than that for Basin and Range extension. As the rate of fault motion in the Coso field was significantly greater, faults serving as conduits for magma were more numerous and thus provided greater opportunity for open conduits to transport magma upward. In contrast, extrusion of Dove Spring basalt coincided with a period of slower crustal deformation. This resulted in periodic magma ponding and assimilation of crustal rock until a suitable conduit was provided.

Introduction

The Owens Valley of California is a 100 mile long fault-bounded basin lying between the Sierra Nevada range to the west and the Inyo and White mountains to the east. The Garlock fault forms the southern boundary and the Long Valley caldera the northern boundary. The Owens Valley lies at the western edge of the Eastern California Shear Zone (ECSZ). The ECSZ has experienced dextral shear at a rate of 15–25 mm/year for the past 3–6 million years. The southern Owens Valley has undergone intermittent volcanism since the Neogene. Volcanism began during the initial stages of Basin and Range extension (~10 Ma) and has continued into the Quaternary regime of right slip.

This research examines the geochemistry and petrology of basaltic volcanic fields in the southern Owens Valley (Figure 1). The Ricardo volcanic field

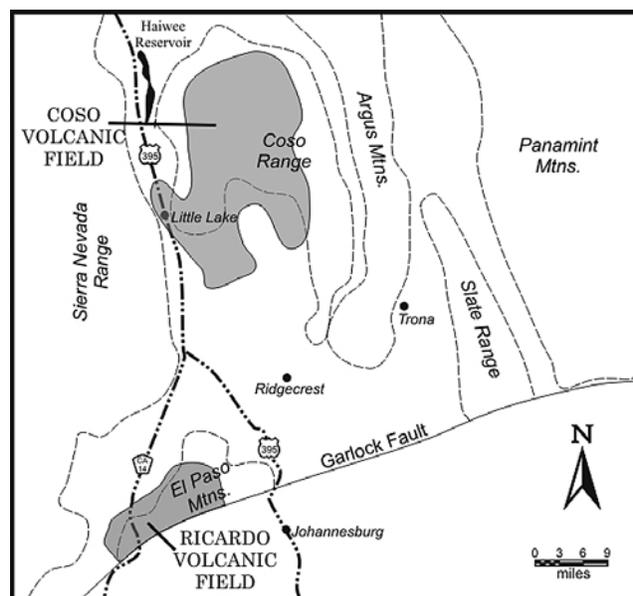


Figure 1. Index map showing the location of southern Owens Valley volcanic fields (shaded gray areas).

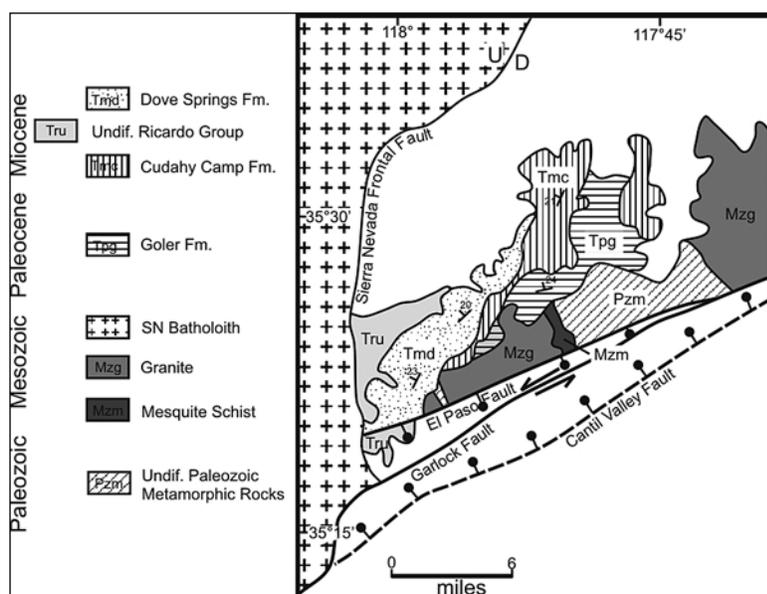


Figure 2. Geologic map of the southern El Paso Mountains (after Loomis and Burbank, 1988).

lies within the El Paso Mountains and is traversed by CA Highway 14. The Garlock fault is situated 1–3 miles to the south. The Ricardo Group of the El Paso Mountains is comprised of the Cudahy Camp and Dove Spring formations. The latter unit, the focus of this paper, experienced basaltic volcanism during the Miocene (11.7–8.1 Ma) (Frankel, et al., 2008).

The Coso volcanic field, situated 25 miles to the north-northeast, lies along the flank of the Sierra Nevada range. U.S. Highway 395 runs north–south along the western edge of the field. The China Lake Naval Weapons Station surrounds much of the area of active volcanism as well as the Coso Geothermal Plant. The field is comprised of an arcuate array of Pliocene to Quaternary rhyolite domes and basaltic cinder cones and flows covering 150 square miles. The volcanics were emplaced during two separate events; an early phase from 4.0 to 2.5 Ma and a more recent phase from 1.0 Ma to Recent time (Duffield, et al., 1980).

Geologic setting

The oldest rocks in the El Paso Mountains are a series of late Paleozoic sedimentary rocks, termed the Garlock Formation by Dibblee (1952; see figure 2). The Garlock Formation is well exposed in the northeastern El Paso Mountains. Mesquite Schist, of uncertain age, lies in unconformable contact with the Garlock Formation. Both units have been intruded by Mesozoic diorite and granite of the Sierra Nevada batholith. The Paleozoic and Mesozoic rocks are unconformably overlain by the Paleocene Goler Formation, a thick sequence of conglomerate and sandstone.

Merriam (1913) recognized a series of fossiliferous, Miocene age continental sedimentary rocks and volcanics near Red Rock Canyon and termed them the Ricardo Formation after the nearby town. Dibblee (1967) divided the Ricardo Formation into eight members with two distinct lithostratigraphic units: a lower volcanic unit of andesite, tuff, basalt, and conglomerate that lies unconformably on the Goler Formation; and a thicker upper unit of clastic sediments, chert, basalt, and tuff. A 1.5 million year disconformity separates the upper and lower units. Loomis and Burbank (1988) subsequently elevated Dibblee's Ricardo Formation to a Group, subdividing it into two formations.

The Cudahy Camp Formation is comprised of Dibblee's Members 1 and 2. It consists of a 1000 foot thick sequence of andesite, tuff, coarse clastics, and basalt. K–Ar geochronology of the basalts (Cox, 1982) suggests ages of 17.1 to 15.1 Ma. The 5400 foot thick Dove Spring Formation lies unconformably above the Cudahy Camp and consists of Dibblee's Members 3–8. The Dove Spring is comprised of fluvial and lacustrine conglomerate, sandstone, mudrock, and chert, as well as basalt and tuff. Dove Spring strata are exposed in a northeast-striking homocline near Red Rock Canyon. Four basalt flows have been mapped within the Dove Spring Formation by Loomis (1984); Tba1–3 and a fourth undifferentiated flow within Loomis's unit 4. Recent ages published by Frankel, et al. (2008) suggest the three lowermost basalts were extruded between 11.7 and 10.5 Ma. No age is available for the uppermost basalt flow, but Burbank and Whistler (1987) quote an age of 8.1 Ma for a rhyolite tuff immediately above the basalt. This brackets basalt extrusion between about 12 and 8 Ma. The Dove Spring is overlain by Pleistocene lacustrine silt and clay, alluvial conglomerate, and terrace gravel, also assumed to be Pleistocene in age (Dibblee, 1952).

The structural geology of the El Paso Mountains is typical of the Basin and Range region north of the Garlock fault. The western boundary is formed by a Sierra Nevada frontal fault that coincides with the western limit of Basin and Range extension. The dominant structural features of the area are the left-lateral Garlock fault and the nearby El Paso fault. The Garlock fault is a major transform fault which separates the Mojave Block to the south from the Basin and Range Province to the north (Davis and Burch-

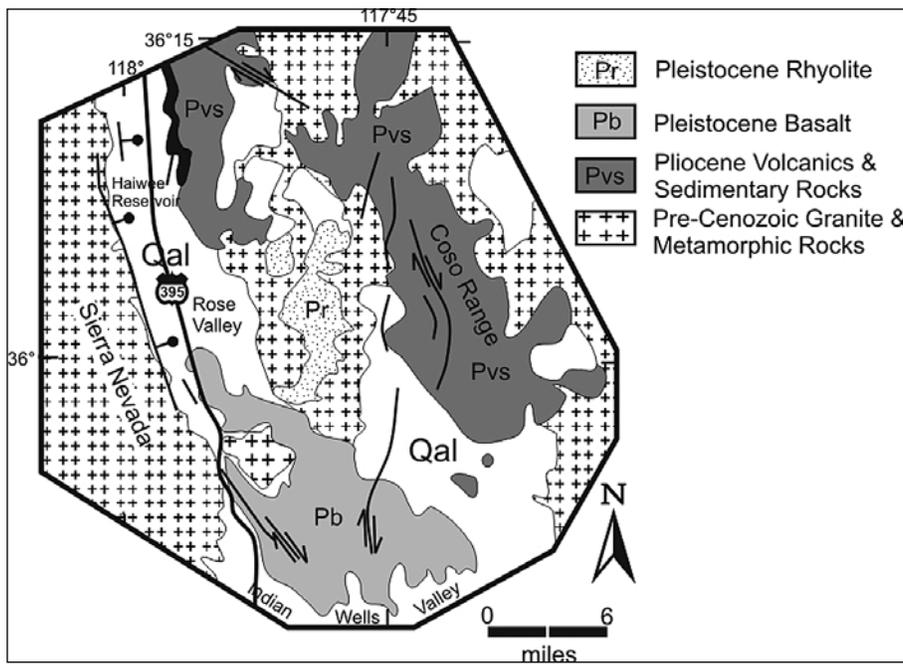


Figure 3. Simplified geologic map of the Coso Range (after Duffield, et al., 1980).

fiel, 1973). Subsequent dip-slip (south side down) motion along the El Paso fault has uplifted the El Paso Mountains.

Both the Goler Formation and Ricardo Group were deposited in an elongate trough, termed the El Paso Basin, that formed north of the Garlock fault (Loomis and Burbank, 1988). The Goler Formation was tilted to the north and eroded prior to Miocene sedimentation. Two east-west trending, steeply-dipping faults that juxtapose basement against the Goler Formation were active during deposition of the Goler Formation (Cox, 1982). The Dove Spring Formation is cut by several north-south striking normal faults that dip 30–40° to the east, suggesting that strata as young as 8 Ma have also been affected by extension in an east-west direction. The El Paso fault crops out as a series of discontinuous, essentially east-west striking segments that merge with the Garlock fault near the east edge of the El Paso Mountains. The fault appears to terminate before reaching the Sierra Nevada range to the west (Dibblee, 1952). Although mapping indicates that the El Paso fault is a strand of the Garlock fault, there is no evidence for left-lateral slip (Loomis and Burbank, 1988). Throughout most of its outcrop the Ricardo Group is tilted west-northwest toward the Sierra Nevada fault. Tilting is consistent with horizontal axis rotation accompanying east-west extension and is confined to late Miocene to Pleistocene, as evidenced by overlying untilted Pleistocene sediments.

Deposition of the Cudahy Camp Formation is thought to have occurred during a period of east-

west tension associated with the northward migration of the Mendocino triple junction (Glazner and Loomis, 1984). The tension produced no accompanying extension. The middle to late Miocene rocks of the Dove Spring Formation are distinctly different from the Cudahy Camp Formation and contain strong evidence of east-west Basin and Range extension. That evidence suggests Basin and Range extension began 1–2 million years after the passage of the Mendocino fracture zone. The sedimentary record of this extension is marked initially by renewed basin growth and sedimentation north of the Garlock fault,

culminating with accumulation of detrital sediment from the Sierra Nevada beginning about 8 Ma.

Loomis and Burbank (1988) conclude that the evolution of the El Paso basin and extrusion of the Dove Spring basalts was coincident with the onset of sinistral slip along the Garlock fault and the east-west Basin and Range extension. The Miocene sequence records: volcanism and north-south tension without net extension at 17–15 Ma (Cudahy Camp); relative uplift about 15–13.5 Ma (resultant disconformity); the onset of sinistral slip on the Garlock fault and east-west Basin and Range extension about 10–9 Ma (Dove Spring Formation); and emergence of the Sierra Nevada as a topographic high by 8 Ma.

In many respects, the stratigraphic relationships within the Coso volcanic field mirror those of the Ricardo volcanics 25 miles to the south-southwest. The Coso Range is underlain by Mesozoic plutons and Paleozoic metamorphic rocks (Figure 3). Cenozoic volcanic and sedimentary rocks overlie the basement complex. The sedimentary rocks range in age from Miocene to Pleistocene and vary in lithology from coarse to fine fluvial and lacustrine clastics. Holocene gravels and conglomerates have been deposited along recent normal faults. Pre-Cenozoic rocks consist of granitic, dioritic, and gabbroic plutons, and metamorphic rocks. The Jurassic to Cretaceous age plutons are possibly related to the Sierra Nevada batholith to the west. Numerous north-northwest-trending dikes of the Independence dike swarm crosscut the plutons.

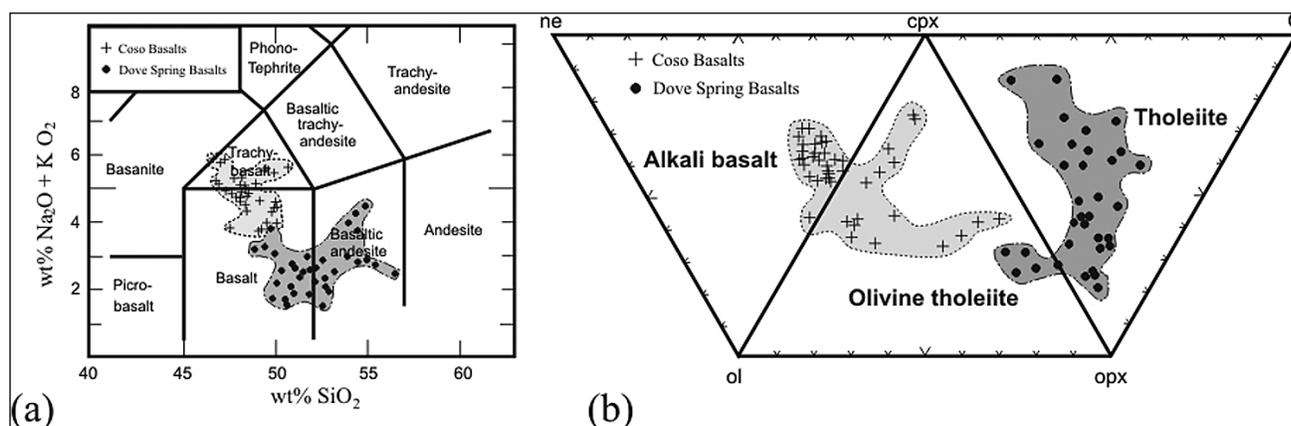


Figure 4. (a) TAS diagram for the Ricardo (Dove Spring) and Coso basalts; and (b) modified basalt tetrahedron for the Dove Spring and Coso basalts (some data for the Coso field from Groves, (1996) and DePaolo and Daley, 2000).

Pliocene volcanic rocks crop out in an arcuate chain extending from Haiwee Reservoir to Volcano Butte at the southeast end of the Coso Range. Widespread, generally alkaline basalt flows represent the earliest phase of Pliocene volcanism followed by the more local extrusion of volcanics of variable composition from basalt to rhyolite (Duffield, et al., 1980). K–Ar dates suggest the onset of volcanism at 4.0 Ma, continuing to 2.4 Ma (Duffield, et al., 1980). Pleistocene volcanism was characterized by bimodal eruption of alkali basalt and rhyolite (Babcock, 1977). Basalt is usually older than rhyolite, but locally the emplacement sequence is reversed. The locus of Pleistocene basaltic volcanism lies along the western margin of the field with rhyolitic vents centered on the Pleistocene basalt flows (Figure 3). K–Ar ages for the Pleistocene basalts range from 1.1 Ma to 40 ka (Duffield, et al., 1980).

Sedimentary rocks of the Coso Formation crop out along the west and north flanks of the Coso Range. Rock types range in composition from conglomerate to arkose and fine-grained, tuffaceous lake-bed sediment. Locally, rhyolite tuffs occur within the section, as at Haiwee Reservoir. Schultz (1937) suggested a Plio–Pleistocene age for the Coso Formation on the basis of vertebrate fossils. The youngest sediments in the Coso Range include fluvial gravel and aeolian sand, silt, and clay.

Weaver and Hill (1979) suggested that volcanism and deformation within the Coso Range were due to a transtensional environment created by a releasing bend in a dextral strike-slip fault system. Monastero, et al. (2005) place the principal boundary faults of the strike-slip system as the Little Lake fault in Indian Wells Valley to the southwest, and in the Wild Horse Mesa area to the northeast. They believe that a buried dextral shear fault is present near Wild Horse Mesa

and that this fault continues northward beneath Rose Lake and southern Owens Lake. Presumably, it would represent a southern extension of the Owens Valley fault system, exposed to the north near Lone Pine. The arcuate Airport Lake fault represents a cross-basin fault, analogous to the Black Mountain detachment in Death Valley, formed by the step over. Duffield, et al. (1980) point out that many of the Coso Pleistocene basaltic cinder cones lie astride arcuate dip-slip faults. It is possible these faults represent the surface expression of the transtensional accommodation zone.

Christiansen and Lipman (1972) describe the Coso Range as one of many Cenozoic volcanic fields in the Basin and Range that are “fundamentally basaltic” and the product of tectonic extension. The Pleistocene volcanic rocks of the range are a classic example

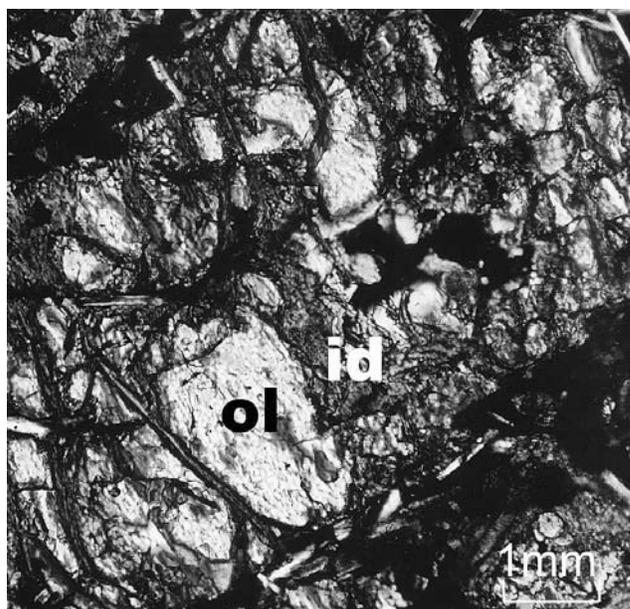


Figure 5. Large olivine phenocryst that has been veined and replaced by iddingsite. The olivine (ol) remnants are the light-colored grains while the darker amorphous material is iddingsite.

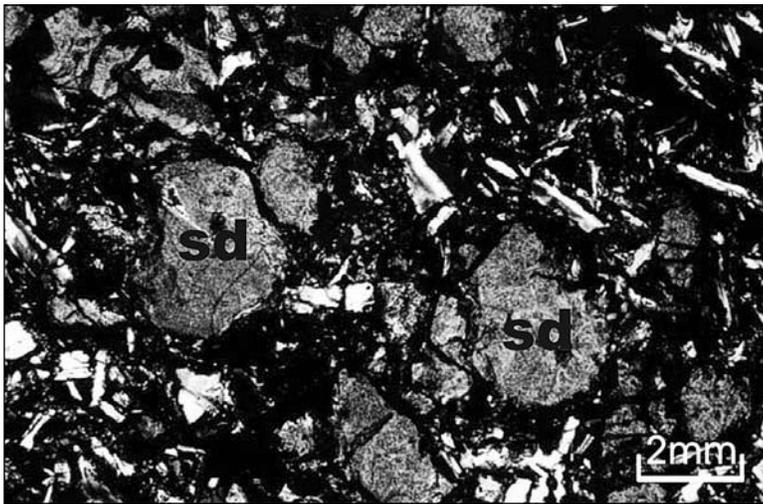


Figure 6. Ca siderite (sd) replacing and pseudomorphing olivine.

of the bimodal basalt-rhyolite association that characterize these rock suites. The Pliocene volcanic rocks, however, are more compositionally diverse and may or may not represent the “classic” rock suites as described by Christiansen and Lipman. Nonetheless, the association of both Pliocene and Pleistocene volcanism in the Coso Range with crustal extensional is indicated by emplacement of flows and cones along dip-slip faults.

Basalt geochemistry and petrology

This research sampled four basalt flows (Tba1–Tba3 of Loomis (1984) and a fourth undifferentiated flow within unit 4 of Loomis), as well as dacitic to rhyolitic tuffs of the Dove Spring Formation emplaced between about 12 and 8 Ma. While a small number of basalt/rhyolite samples were collected from the Cudahy Camp Formation for the sake of comparison, limited geochronology as well as uncertainty of tectonic setting precluded detailed examination.

Figure 4a is a TAS (total alkalis vs. silica) diagram for the basalts of the Dove Spring Formation and Coso volcanic field. Dove Spring (DS) basalts average approximately 53% total SiO₂ and 3% total alkalis. In contrast, basalts of the Coso field (CV) average 48% SiO₂ and 5% total alkalis. Figure 4b is a modified basalt tetrahedron based on normative mineralogy. The use of normative mineralogy allows a more sensitive discrimination of basalt types. As can be seen from the diagram, DS basalts are generally tholeiites while CV basalts span a more diverse compositional range, but are mostly alkali basalt. It is worth noting that Groves (1996) concentrated her sampling within the Pleistocene basalts. They plot in a restricted compositional range of the alkali basalt subtriangle. Samples of both Pliocene and Pleistocene basalts by DePaolo

and Daley (2000) and the authors display a more diverse compositional range encompassing both alkali basalt and olivine tholeiite subtriangles.

Plagioclase is the most common phenocrystic phase in DS basalts, followed by skeletal clinopyroxene and rare olivine. Olivine phenocrysts appear common in outcrop, but thin section analysis reveals that much of the olivine is actually iddingsite, a cryptocrystalline mixture of hydrated iron oxides and magnesium clays. The iddingsite veins, embays, coats, and in many cases completely replaces olivine grains (Figure 5). The presence of iddingsite has also been reported in the Big Pine field by Darrow (1972) and in Darwin basalts by Lusk (2007). Its significance to basalt petrogenesis will be discussed below.

Anderson (2005) noted that some olivine phenocrysts appeared to be pseudomorphed by dark brown to black calcium siderite (Figure 6). The association of calcium siderite, iddingsite, and olivine has not been previously documented in the Owens Valley, but was noted, curiously, in Martian meteorites by Vicenzi, et al. (2001). Its presence was attributed to the weathering of the Martian surface by CO₂-saturated fluids or vapors. The reason for its presence in DV basalts is uncertain.

Pliocene Coso basalt is characterized by small phenocrysts of plagioclase, olivine, and ophitic clinopy-

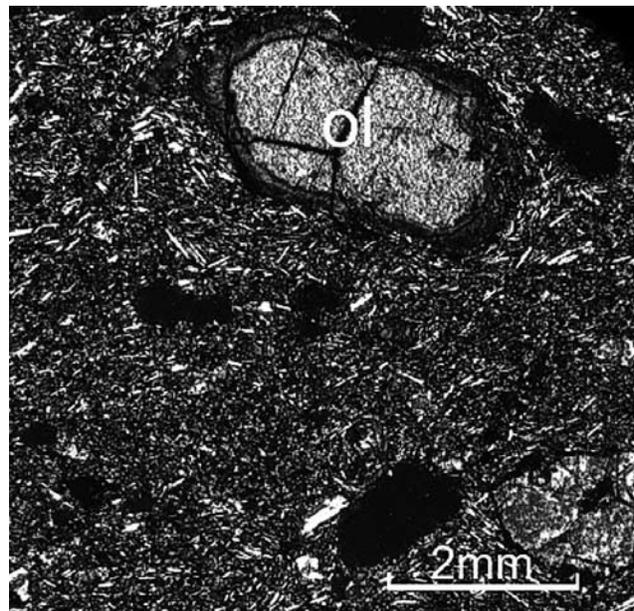


Figure 7. Iddingsite reaction rim coating olivine (ol) from the Darwin, CA volcanic field.

roxene, whereas Pleistocene basalt is strongly porphyritic, containing varying proportions of olivine, plagioclase, clinopyroxene, and opaque oxides (Duffield, et al., 1980). Many flows reportedly contain quartz grains as large as several millimeters in diameter. This latter observation presents a paradox. The crystallization of quartz requires a silica-oversaturated magma; however Duffield, et al. (1980) characterize both Pliocene and Pleistocene basalts as alkali basalts. Formation of alkali basalts necessitates an undersaturated magma that does not yield quartz as a primary mineral phase. We suggest two possible explanations for the presence of quartz. Because the Coso Range is an active geothermal field, the quartz may represent a secondary mineral phase deposited from silica-saturated hydrothermal fluids. A second possibility centers on our geochemical data. It reveals that many Coso samples are nepheline normative. Although the authors have not observed nepheline as a modal phase, it is possible the quartz of Duffield et al. (1980) might be nepheline, which can easily be mistaken for quartz in hand sample.

The occurrence and significance of iddingsite

An important difference between the Coso and Ricardo volcanic fields is the apparent lack of iddingsite in Coso basalts, while it is a common phase in Ricardo basalts. Iddingsite, a mixture of various hydrous silicates of iron and magnesium, is formed by alteration of olivine.

Sun (1957) concluded from x-ray studies of iddingsite that only a single crystalline phase, goethite, was present. Other substances occurred in amorphous grains or clots. As such, he regarded it as a complex alteration product of olivine rather than a true mineral. The alteration of olivine consists of the addition of Fe₂O₃ and the removal of MgO (Gay and Le Maitre 1961). The chemical formula for iddingsite is approximated as MgO•Fe₂O₃•3SiO₂•4(H₂O) where CaO can be substituted by MgO in a ratio of 1:4 (Edwards, 1938; Ross and Shannon, 1925). Iddingsite normally occurs as a reaction rim or corona coating olivine (Figure 7), but can also vein or embay and corrode olivine grains. The color of iddingsite varies from red-brown to brown or green.

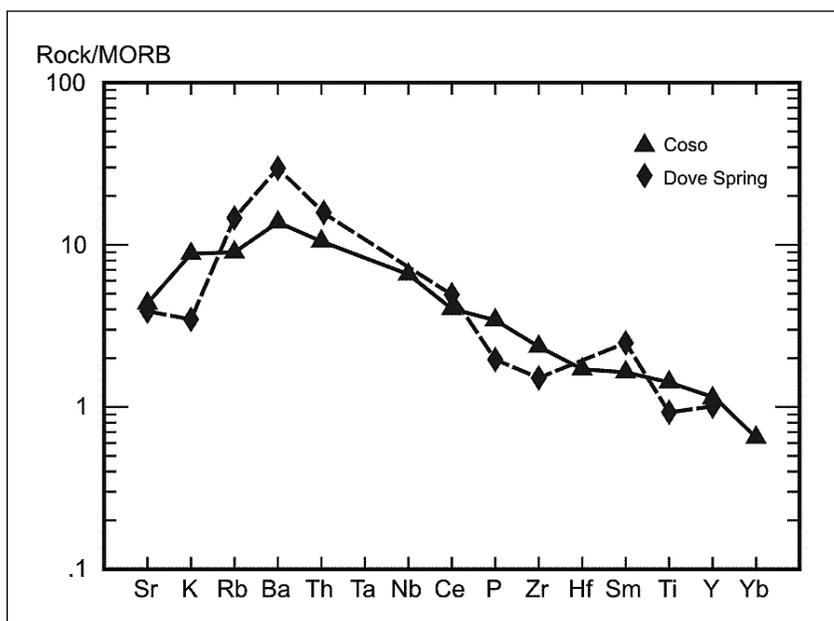


Figure 8. Spider diagram for the Coso and Dove Spring basalts utilizing the MORB standard of Pearce (1982). (Some data from Groves, 1996).

The occurrence of iddingsite is controversial. It appears to be limited to olivine-bearing hypabyssal or extrusive rocks and absent from plutonic rocks. Ross and Shannon (1925) were the first to suggest that iddingsite is a deuteritic mineral formed during the final cooling of basaltic lava. The oxidation of iron from the ferrous to ferric state plays an important role in the instability and conversion of olivine to iddingsite. Edwards (1938) suggested the alteration occurred when the fugacity of oxygen in the magma chamber increased, which he attributed to a circulating thermal solutions. He suggested the addition of water may have occurred during magma mixing. Baker and Haggerty (1967) concluded that alteration could occur at both elevated temperatures and during the weathering process. They felt that any process that resulted in oxidation and the addition of water would produce iddingsite. Sun (1957), however, stated "it may be concluded generally that the alteration of olivine to iddingsite occurs most likely in a highly oxidizing solution, at high temperature and under high pressure." Furgal and McMillan (2001) reached a similar conclusion, and cautioned that alteration required elevated temperatures, not those of typical circulating groundwater.

The presence of iddingsite in the basalts of the Ricardo Group and its absence in the Coso volcanics is important to an understanding of Owens Valley basalt petrogenesis. Apparently, the Ricardo basaltic lavas experienced an increase in oxygen fugacity while the Coso magmas did not. This increase could

have resulted from magma mixing, assimilation of hydrous crustal rocks, an encounter with circulating meteoric water or simply weathering at the earth's surface. We will reexamine this topic in the Discussion section below.

Discussion

Ringwood (1976) suggested that melting depth was an important constraint on magma composition. Partial melts from the upper mantle and lower crust tend to be tholeiitic, whereas magmas generated from deeper in the mantle are alkaline. Anderson and Jessey (2005) concluded that the DS tholeiites represented a shallow melt, perhaps subducted East Pacific Rise, while the olivine-bearing, alkaline Coso basalts were derived from primitive mantle at greater depth. Brown, et al. (2008), however, pointed out that iddingsite is a common mineral phase in DS basalts requiring that olivine was at least initially an important constituent of the DS magmas. Therefore, DS magmas may have originally been as alkaline as those of the Coso volcanic field and thus melting depth might not be an adequate explanation for the observed compositional differences.

The presence of iddingsite places additional constraints on any genetic model. While iddingsite can be the product of weathering; textural and geochemical relationships for the Big Pine and Darwin basalt fields tend to discount this possibility (Lusk 2007). It is more likely the iddingsite formed during the latter stages of magma crystallization when changes in oxygen fugacity led to instability of olivine. The most common mechanism for such a change is an encounter with shallowly-circulating meteoric water. This does not, however, account for the significant compositional variation between tholeiitic DS and alkali CV basalts. While meteoric water may result in the oxidation of ferrous iron and instability of olivine, it will have limited affect on bulk rock chemistry. The large difference in chemistry between the two fields requires either magma mixing or assimilation. However, the effective mixing of mafic and felsic magmas has always presented chemical and physical problems due to density and viscosity disparities. The abundance of basalt and rhyolite, but the near absence of their mixed product, andesite, in the Coso and Ricardo fields is further evidence for the inherent problems with the mixing model.

Assimilation requires a stalling or ponding of the basaltic magma at depth in the crust with heat from the magma melting or assimilating crustal rocks.

Obviously, the amount of material assimilated must be small, as large scale melting would result in sufficient heat loss to cause crystallization of the basaltic magma. However, melting of only a few percent of the surrounding crustal rock could account for the observed compositional differences in the magmas, and alter oxygen fugacity. The ponding of the magma would result from gravitational stability of the mafic melt. It would rise to shallow levels in the crust and stall when its density equaled or exceeded that of surrounding rocks. While ponded, heat from the magma would melt or assimilate country rock until an open conduit presented itself, perhaps by fault movement. The basaltic magma would then rise to the surface and be extruded.

If the assimilation model has merit, it implies that the alkali Coso basalts must have risen quickly to the surface with limited, or no interaction with the shallow crust. In contrast, the tholeiitic Dove Spring basalts were ponded melting crustal rocks that changed magma oxygen fugacity and composition. Evidence to support assimilation comes from both isotopic and trace element data. Rb/Sr and Nd/Sm isotopic data, (see Jessey, et al., this volume) reveals that Coso basaltic magmas are primitive mantle melts that have not been contaminated by a crustal isotopic overprint.

Figure 8 is a spider diagram for trace elements from the DS and CV basalts. The most incompatible elements lie near the left-center of the diagram (K thru Ta). Note that DS basalts are distinctly enriched in incompatibles (Rb, Ba, Th). This could occur if the DS magmas were ponded assimilating crustal rocks that characteristically contain the greatest concentration of incompatible elements.

Loomis and Burbank (1988) argue that the Dove Spring Formation was deposited during the onset of Basin and Range extension. Rates of vertical crustal displacement during the late Miocene are uncertain, but Le, et al. (2007) found that Holocene extension in the eastern Sierra Nevada occurs at a rate of 0.1-0.2 mm/yr. In contrast, the Coso volcanic field formed in a transtensional basin created by dextral shear along the Owens Valley fault zone. Estimates of rates of motion across the Owens Valley fault zone vary from 3 to 8 mm/yr, more than an order of magnitude greater than the rate of vertical displacement.

This may provide a possible explanation for the differences in basalt composition. As the rate of motion is much greater in the Coso field, faults serving as conduits for magma would remain open. However, in the Ricardo field, Basin and Range extension resulted

in much slower rates of displacement, causing periodic magma ponding until a suitable conduit was created by dip-slip fault movement.

We conclude:

- Dove Spring basalts are quartz normative tholeiites while Coso basalts are nepheline/olivine normative alkali basalt.
- Ricardo basalts are characterized by partial to complete replacement of olivine by siderite and iddingsite while Coso basalts contain unaltered phenocrysts of olivine.
- Differences in chemistry and petrology were caused by the ponding of Dove Spring magmas at shallow levels in the crust whereas Coso magmas were largely unimpeded during their ascent to the surface.
- Ponding of Ricardo magmas was a consequence of slower rates of vertical uplift associated with Basin and Range extension as compared to the more rapid dextral shear regime that characterizes the emplacement of Coso basalts.

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Vegetation of the Owens Valley and the White Mountains

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Introduction

Owens Valley is a trough-shaped depression 90 miles long, lying between steep-fronted mountain blocks rising 7,000 to 10,800 feet above the valley floor on either side. The Sierra Nevada rise to the west and the White Mountain Range lies to the east, with only 20 miles separating their crests (Putman, 1995). Other significant features include the Owens River and Owens Lake, with valley floor elevations ranging from 3,500 to 4,500 feet.

The White Mountain Range is about 110 miles long with a maximum width of approximately 22 miles east of Bishop (Hall, 1991). White Mountain Peak rises to 14,246 feet, giving it the distinction of being the third highest peak in California and making the Whites the highest mountain range of the Great Basin Province. Though sometimes referred to as the White-Inyo Range, the separation of the White Mountains from the Inyo Mountains along Westgard Pass Road is of no real geological or topographical significance; many authors choose to call the range simply the White Mountains. The rise from the Owens Valley floor to elevations of over 14,000 feet is abrupt, occurring within a 12 mile line. The rapid changes in elevation create corresponding changes in habitats and species composition.

Owens Valley lies at the juncture of three very different bioregions: the Sierra Nevada, the Great Basin, and the Mojave Desert—each with distinct climate variables and assemblages of plants



Figure 1. White Mountain Peak, the third highest peak in California. Photo by Leah Gardner.

(Putman, 1995). With elevations ranging from 3,500 feet to over 14,000 feet, a variety of plant communities can be encountered over short distances. Traveling south through the Owens Valley in Inyo County,



Figure 2. View from the White Mountains down to Bishop in the Owens Valley and across to the Sierra Nevada crest. Photo by Leah Gardner.

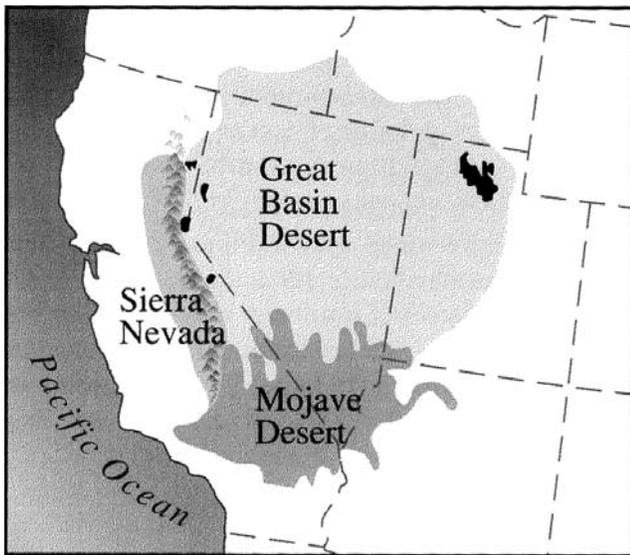


Figure 3. The biogeographic units of the Great Basin, the Sierra Nevada, and the Mojave Desert intersect in the Owens Valley. from Putman and Smith.

you will notice a shifting assemblage of plants as you leave behind shrubs characteristic of the Great Basin, such as sagebrush (*Artemisia tridentata*), and start to see more of those found in the Mojave Desert, such as creosote bush (*Larrea tridentata*). Great Basin shrubs also make up much of the vegetation of the lower foothills along the eastern Sierra front. Travelling east or west up the steep mountain slopes, a montane flora is encountered, arranged in elevational bands up to the alpine zone.

Plants from the different bioregions and elevation zones usually blend together rather than stopping and

starting at discrete boundaries, resulting in greater biodiversity along these ecotones. An exception to this normal blending pattern occurs in the White Mountains where different geological substrates come together and are reflected in abrupt changes in vegetation patterns and species composition. The sagebrush scrub growing on sandstone abruptly gives way to open forests of bristlecone pine (*Pinus longaeva*) on exposures of white dolomitic limestone, a dramatic example of the effect of substrate on plant distribution (Kruckeberg, 2006). Granitic and metavolcanic substrates also occur in these mountains.

Due to this meeting and mixing of bioregional elements, elevational gradients, and geological substrates, the Inyo region is considered to be the floristically richest region in transmontane California and an important center of endemism. Over 3,500 native plant taxa have been described for the Inyo Floristic Region (Hall, 1991), 454 of which occur only within this region in California, and 199 are endemic to the state (Davis, et al., 1997).

According to the geographic subdivisions described in the Jepson Manual, this region lies outside of the California Floristic Province and falls within the Great Basin Province. The Owens Valley is part of the section labeled "SNE" for East of the Sierra Nevada, extending along the eastern edge of the Sierras to the southern end of Owens Valley, where it transitions to Mojave Desert scrub. The White and Inyo Mountains, labeled "W & I", are considered a separate sub-region due to the presence of subalpine bristle-

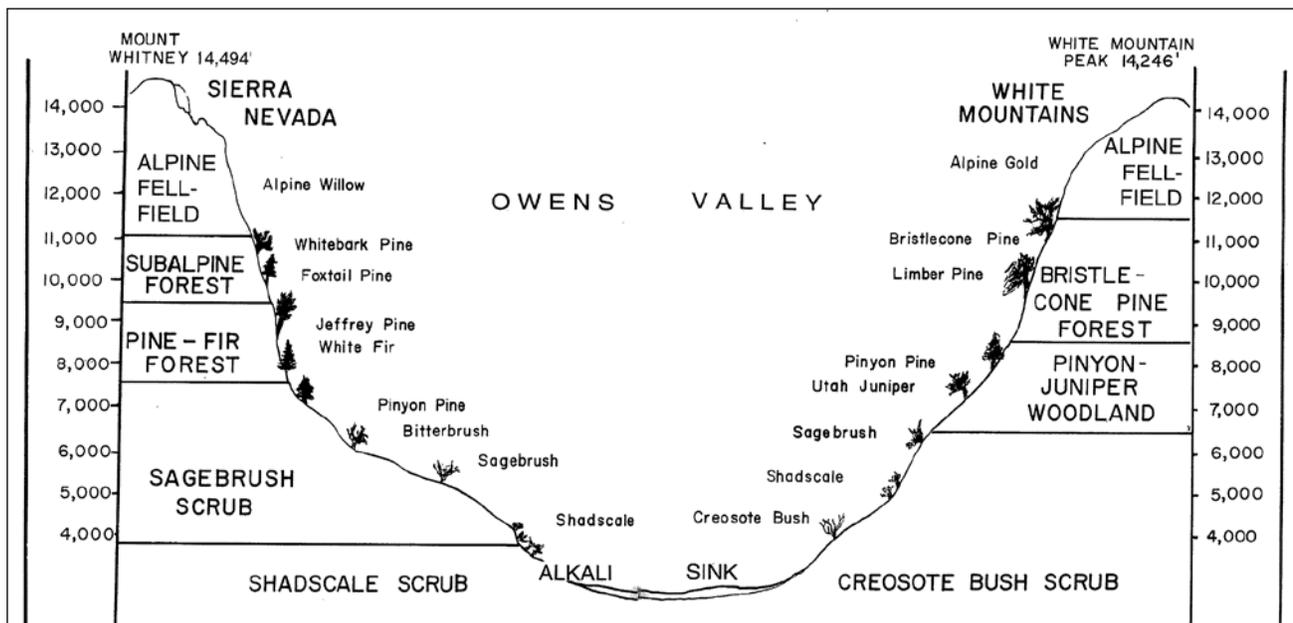


Figure 4. A cross-section of the plant communities across the Owens Valley. from Putman and Smith.

cone and limber pine woodlands and alpine vegetation (Hickman, 1993).

Climate

The aridity characterizing the region's climate is due to the rain shadow effect. The high wall formed by the Sierra Nevada traps most of the moisture released by Pacific storms, allowing very little to travel further east. Annual precipitation typically ranges from 4 to 6 inches in the Owens Valley, increasing to 20 inches along the crest of the Whites, more than half of which falls as snow. By contrast, the mountains of the Eastern Sierra receive an average of 10 to 20 feet of snow each winter (Putman, 1995). The general regional pattern for precipitation is one of increasing aridity from north to south and from west to east. Air descending from the Sierra Nevada warms 4–5 degrees for every 1000 feet of elevation loss, resulting in nearly constant strong winds through the valley and across the White Mountains.

Vegetation Types

Shrub Communities

Upland xerophytic shrub communities occur on the broad bajadas and alluvial fans which spread out from canyon mouths of the Sierra Nevada and the White Mountains. These appear the same from a distance, but they differ in species composition on closer inspection.

Sagebrush Scrub and Sagebrush Steppe—Seven species of sagebrush (*Artemisia*) occur in the White-Inyo Range, forming a dominant part of the vegetation from the valley floor up to the alpine zone (Spira, 1991). All are characterized by silvery-grey, small, aromatic foliage. Great Basin sagebrush (*Artemisia tridentata*) is the most abundant



Figure 5. Sagebrush scrub is common from the valley floor up to the alpine zone. Photo by Leah Gardner.

species from the valley floor to middle elevations. Common associates include bitterbrush (*Purshia tridentata*), desert peach (*Prunus andersonii*), rabbitbrush (*Chrysothamnus naseosus* and *C. viscidiflorus*), and Mormon tea (*Ephedra viridis*). In the subalpine zone and above, alpine or dwarf sagebrush (*Artemisia arbuscula*) becomes the dominant species, comprising an extensive shrub-steppe vegetation type on sandstone (Sawyer and Keeler-Wolf, 2007).



Figure 6. Alkali scrub or alkali meadow occurs in low desert basins high in salts and pH. Here, salt grass dominates the meadow around Owens Lake. Photo by Leah Gardner.



Figure 7. A common vegetation type in the valley, shadscale scrub is comprised of several members of the *Atriplex* genus. Photo by Leah Gardner.

Desert Scrub Vegetation: Alkali Sink Scrub, Shadscale Scrub, and Creosote Bush Scrub. 3,500-6,500 feet.

Alkali Sink Scrub, Alkali Marsh and Alkali Meadow—In the lowest desert basins, water collects and evaporates, leaving behind salts and soils with a high pH. Plants that can live along the margins of the dry barren lake beds (playas) in these alkali sinks are called “halophytes,” literally meaning salt-loving. Examples of common halophytes include saltgrass

(*Distichlis spicata*), arrow-scale (*Atriplex phyllostegia*), iodine bush (*Allenrolfia occidentalis*), bush seepweed (*Suaeda moquini*), and yerba mansa (*Anemopsis californica*). Additionally, many rare plants occur in the alkali marsh, playa, and meadow habitats. This vegetation type is common around Owens Lake and throughout the Owens Valley.

Shadscale Scrub—This vegetation is frequently associated with moderately alkaline soils and is found in the northern Mojave, in the Owens Valley and along the lower slopes of the White Mountains. Dominant species are comprised of several members of the *Atriplex* genus, including shadscale (*A. confertifolia*), allscale (*A. polycarpa*), and saltbush (*Atriplex canescens*).

Common associated species include hop-sage (*Grayia spinosa*), winter fat (*Krascheninnikovia lanata*), budsage (*Artemisia spinescens*), and yellow rabbitbrush (*Chrysothamnus viscidiflorus*).

Creosote Bush Scrub—Creosote bush is an evergreen shrub, able to withstand a wide range of conditions,

but typically occurring in areas receiving between 2 to 8 inches of rainfall. The common vegetation type dominated by widely-spaced creosote bush shrubs covers 60–70% of the lower elevations of the Mojave Desert. Common associates include burro bush (*Ambrosia dumosa*), brittle bush (*Encelia farinosa*, *E. actoni*, *E. virginensis*), and cheesebush (*Hymenoclea salsola*). Driving south along Highway 395, you first notice the appearance of creosote bush in the vicinity of Ash Creek, west of Owens Lake. It also grows on the lower foothills of the White Mountains.



Figure 8. Desert scrub, often dominated by creosote bush, covers most of the low elevations of the Mojave Desert and the foothills of the White Mountains. Photo by Leah Gardner.



Figure 9. Joshua trees, unique to the Mojave at elevations from 2,500 to 4,500, are found just south of Owens Lake surrounding Haiwee Reservoir. Photo by Leah Gardner.

Woodlands: Joshua Tree Woodland, Pinyon–Juniper Woodland, Subalpine Woodland

Joshua Tree Woodland

Joshua trees (*Yucca brevifolia*) are unique to the Mojave Desert and are found between 2,500 and 4,500 feet elevation in areas receiving 6 to 15 inches of rainfall. You can find them at the extreme southern end of the Owens Valley in the area surrounding Haiwee Reservoir. Other plants found here among these dramatic Mojave indicators are cheesebush (*Hymenoclea salsola*), sweetbush (*Bebbia juncea*), boxthorn (*Lycium andersonii*), and wire lettuce (*Stephanomeria pauciflora*).



Figure 10. Pinyon is the only single-needled pine. Pinyon-juniper woodlands are dominant in the White Mountains from 6,500 to 9,500 feet. Photo by Leah Gardner.

Pinyon–Juniper Woodland

In the White Mountains, pinyon–juniper woodland is the dominant vegetation type from 6,500 to 9,500 feet. *Pinus monophylla* is the Great Basin singleleaf pinyon and the only single-needled pine. Pinyons are very cold-tolerant and drought-resistant, inhabiting semi-arid areas as far north as Idaho. They tend to grow as short-statured, broad-crowned trees that form woodlands of widely spaced trees in association with Utah junipers (*Juniperus osteosperma*). The large seeds of the pinyon, called pine nuts, are eaten by many types of birds and mammals. During the last glacial advance, pinyon–juniper woodlands were widespread throughout the Mojave Desert. They expanded

northwards with the warming interglacial climate, spreading at a rapid rate (Lanner and Devender, 1998). Shrubs commonly associated with this woodland include Great Basin sagebrush (*Artemisia tridentata* and *A. nova*), bitterbrush (*Purshia tridentata* and *P. glandulosa*), green ephedra (*Ephedra viridis*),

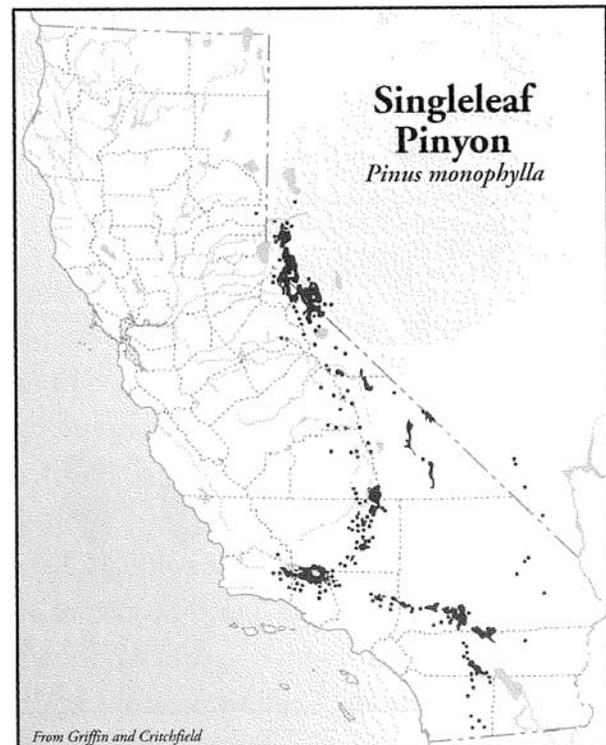


Figure 11. Distribution map of pinyon pines. From Griffin and Critchfield.



Figure 12. Subalpine woodland composed of limber pines and bristlecone pines. Notice the sharp delineation between sagebrush scrub on sandstone soil and subalpine woodland on dolomite. Photo by Leah Gardner.

and rabbitbrush (*Chrysothamnus viscidiflorus*) (Spira, 1991).

Subalpine Woodland: Bristlecone Pines and Limber Pines

Though the White Mountains are best known for the presence of bristlecone pines, limber pines (*Pinus flexilis*) are actually more numerous and widespread. These two pine species grow together in open woodlands on dolomitic soils. Limber pines are not restricted to these sites, but bristlecones are.

Limber pines are found from 8,000 to 11,500 feet across the high peaks of the Great Basin. In California they occur in the White Mountain Range, the southeastern Sierra, and in the Transverse and Peninsular ranges. Bristlecone pines are found only in Mono and Inyo counties in

California where they grow in isolated groves between 10,000 and 11,500 feet. Outside the state, they range into the high peaks of Nevada and Utah (see distribution maps).

Several features help an observer distinguish between these two pines. While both have five needles to a bunch, those of the bristlecone pine are a darker green and are shorter and more tightly clustered on the branches than the limber pine, giving the bristlecone a denser, “bottlebrush” appearance. The limbs of

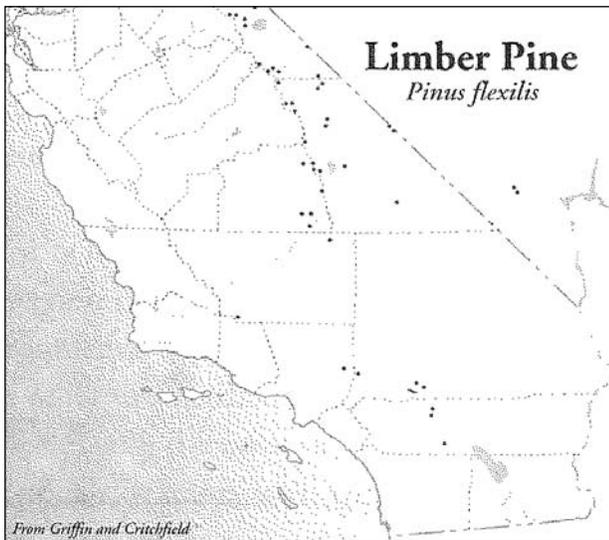


Figure 13. Distribution map of limber pines. From Griffin and Critchfield.

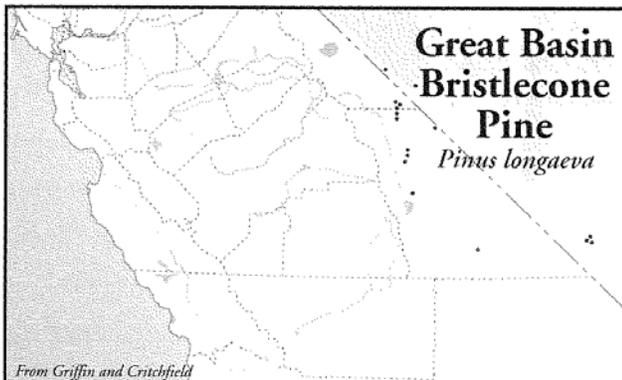


Figure 14. Distribution map of bristlecone pines. From Griffin and Critchfield.



Figure 15. Differences between the limber and bristlecone pines are recognizable in the field. Both have 5 needles but those of the bristlecone are darker green, shorter, and more densely clustered. Photo by Leah Gardner.



Figure 16. The cones of the two pines are distinctive. Note the sharp-tipped barb at the tip of each scale on the bristlecone. Photo by Leah Gardner.



Figure 17. Bristlecone woodlands are best developed on harsh, exposed sites underlain with dolomite limestone substrates where trees reach lifespans exceeding 4,000 years. Photo by Leah Gardner.

limber pine are more open-branching and upward-arching. Finally, the cones are distinctly different. Bristlecone cones have sharp-tipped barbs at the tips of each scale, hence the name “bristle cone.”

Both species are long-lived. While bristlecones are well known for their venerable lifespans, which exceed 4,000 years, limber pines can reach a maximum of 2,000 years. The oldest bristlecone pine trees live on the harshest sites. Due to cold temperatures,

dry soils, high winds, and a short growing season, the trees grow very slowly. Their wood is dense and resinous, resistant to invasion by insect pests or fungi (Lanner, 1999).

The bristlecone pines grow almost exclusively on dolomitic limestone and are considered an indicator species for the presence of dolomite. Soils derived from dolomite are alkaline and of low mineral nutrient value, especially deficient in phosphorus. The light-colored dolomite reflects more sunlight and retains a higher moisture content than the surrounding sandstone; thus they have a cooler root zone (Wright and Mooney, 1965).

Other Dolomite Endemics

In addition to the bristlecone pines, the dolomitic substrate supports a number of other endemic herbaceous plants and dolomite-tolerant species. These include raspberry buckwheat (*Eriogonum gracilipes*), White Mountain horkelia (*Horkelia hispidula*), White Mountain ivesia (*Ivesia lycopodioides* var. *scandularis*), dolomite milk-vetch (*Astragalus kentrophyta* var. *implexus*), and Mono clover (*Trifolium monoense*). Clokey’s daisy, (*Erigeron clokeyii*) is a common herb in the understory of the bristlecone pines, but can occur infrequently on sandstone (Kruckeberg, 2006). In addition to growing in the company of the bristlecone pines in the subalpine woodlands, these edaphic endemics are also found on dolomite barrens above treeline in the alpine zone.

Riparian Vegetation—The riparian vegetation found along the Owens River, streams coming down from the Sierras, washes, and springs contrasts dramatically with that of the surrounding arid landscape. Modern irrigation canals, stock ponds, and reservoirs also provide habitat for riparian plants. Cottonwood



Figure 18. Many other plants are associated with the bristlecone pines on dolomite. Shown here, dolomite endemics include the raspberry buckwheat and Clokey's daisy. Photo by Leah Gardner.

(*Populus trichocarpa* and *P. fremontii*) and yellow willow (*Salix lutea*) are abundant in riparian areas, as are large perennial herbs such as tules and cattails. Water birch (*Betula occidentalis*) forms thick stands in riparian areas from 6,500 to 8,500 feet on the east side of the White Mountains as well as following drainages descending from the Sierra Nevada. At higher elevations in the subalpine zone, quaking aspen (*Populus*



Figure 19. Riparian vegetation composed of cottonwoods, willows, and other water-loving plants occurs along the Owens River as well as around springs, streams, and irrigation canals. Photo by Leah Gardner.

tremuloides) takes the place of the cottonwoods along montane riparian corridors, around springs, and on other moist soils. A genetic dwarf aspen with a shrublike appearance occurs in localized sites in the Whites (Elliott-Fisk, 1991).

Alpine Vegetation

The alpine zone of the White Mountain Range is the most extensive in California, composed of a broad, undulating plateau (Sawyer and Keeler-Wolf, 2007). Above the treeline, alpine vegetation is characterized by low-growing shrubs, grasses, and herbs. Differences across the landscape are dependent on local conditions of moisture, soil and rock characteristics, and slope aspect. The harsh conditions of extreme temperature fluctuations, nearly constant winds, high insolation and ultraviolet radiation, and a short growing season have been mitigated by plants in a variety of ways. Most plants are low-growing perennials forming mats or cushions. Root-to-shoot ratios are high, with much more biomass beneath the surface than above. Leaf adaptations include having a reduced or divided shape to reduce surface area and having silvery-grey coloration or hairs to reflect more UV radiation. Flowers are typically large and showy during their short bloom season to because pollinators are limited in number.

Vegetation types in the alpine zone of the White Mountains are characterized as sagebrush steppe, alpine meadows, and dolomite barrens.

Sagebrush steppe consists of a dwarf shrubland dominated by Alpine sagebrush (*Artemisia arbuscula*). A mix of other shrubs, herbs, and even small cacti grow with the sagebrush on sandstone.

Alpine meadows are composed of a wide assortment of grasses, grass-like plants, and herbs. The wildflower season is short but spectacular here, delighting hikers



Figure 20. At higher elevations, montane riparian vegetation includes additional species such as quaking aspen. Photo by Leah Gardner.



Figure 21. The alpine zone of the White Mountains is the most extensive in California, composed of a broad, undulating plateau covered with low growing vegetation. Photo by Leah Gardner.



Figure 22. Alpine meadows are composed of an assortment of grasses and herbs. Many are cushion- or mat-forming perennials with showy wildflowers. Photo by Leah Gardner.



Figures 23 and 24. Most seasonal wildflowers bloom in spring or summer. An exception is the rabbitbrush, which paints the valley in yellow-golden hues during the autumn months. Photo by Leah Gardner.

when the cushion plants are covered with bright blossoms in July and August.

Dolomite barrens—see “other dolomite endemics” section above.

Seasonal Wildflowers

Though not a distinctive vegetation type, the presence of seasonal wildflowers can put on quite a show in many different plant community assemblages. Herbaceous annuals and perennials as well as flowering shrubs can add a temporary burst of color to an otherwise drab plant palette. Flowering time varies with elevation, starting with spring flowers in the valley and proceeding up the mountains to the cushion plants of the highest alpine zone in mid- to late summer. The valley floor bursts into color again in the fall when rabbitbrush blooms in golden-yellow profusion.

Two genera deserve special mention: *Eriogonum* and *Astragalus*. *Eriogonum*, or wild buckwheat, is the largest dicot genus in California. No less than 32 taxa (species and subspecies in this case) occur in the White Mountains. The Whites are home to at least 12 types of *Astragalus*, or milk-vetch, which is also a species-rich genus with 94 taxa in California (Hickman, 1993). Many of the species and sub-species of *Eriogonum* and *Astragalus* found here are rare and are endemic to the region, occurring nowhere else.

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The rush from the Kern River mines into the Owens Valley

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The Place

Allekey W. Von Schmidt, a forty-niner from Latvia, should have known better. While surveying the eastern boundary of California in July of 1855, he dismissed the Owens Valley as a worthless land with a worthless climate.

Yet of all the places in the Great Basin, the Owens Valley was the most hospitable to settlement: it contained an abundance of water, timber, and land. Flanked by the Sierra Nevada on the west, and the Inyo and White Mountains on the east, the valley is more than 80 miles long and from 4 to 10 miles wide; its elevation ranges from 3,700 to 4,100 feet. Volcanoes, glaciers, and erosion have left several thousand feet of sediment on the bedrock. A strip of land along the foot of the Sierra is especially well drained and fertile. Several major streams, notably Oak and Bishop creeks, flow into the Owens River, which originates in the Mono Lake Basin to the north. Originally averaging 50 feet wide and 15 feet deep, the river emptied into what was then called Big Owens Lake. Today, the lakebed is dry, dusty, and desolate, but in 1859 it was 19 miles long and from 4 to 8 miles wide. Its water, however, was so murky and alkaline that it could support little more than saltwater algae, brine shrimp, and the larvae and pupae of the brine fly, which migratory waterfowl would feed on.

The People

Humans might have arrived in the valley as early as 3500 B. C., but the direct ancestors of today's Native Americans settled there much later, perhaps between A. D. 600 and 1000. They called themselves *nümü*, meaning "person" or "the people" in their language, sometimes called Mono, which is part of the Uto-Aztec linguistic family. Most *nümü* lived in seven districts, or homelands, in the Owens, Round, and Benton valleys. The districts, in turn, contained clusters of hamlets in which the people lived as a community.

To obtain much of their fat and protein, the *nümü* depended on two insects. The caterpillar, or larva, of the Pandora moth (*Coloradia pandora*) provided an

important source of fat and protein. The larvae fed on the needles of Jeffrey (yellow) pines in extensive forests between Mono Lake and the Owens River watershed. In late June and early July, the caterpillars would descend the trees to pupate. Every two years, foraging parties would camp in these forests and, using smudge fires, would drive them down into trenches dug around each tree. Fires then would be started in the trenches, and the larvae would be baked. A party could quickly prepare a ton or more of larvae. Afterward, the larvae would be eaten or stored in elevated bark shelters. The other insect was the brine fly (*Ephedra hians*), which lived in Owens Lake. In late summer, the pupae would wash up onto the shore, where they would pile up in windrows. The people would gather hundreds of bushels of the pupae, which they would dry in the sun and shell (by rubbing) and then store for the winter or trade with groups far to the north.

Seeds provided the other major source of protein. One important seed was the pine nut, borne by the single-leaf piñon (*Pinus monophylla*). Piñon grew along the eastern Sierra Nevada, along the western slopes of the White Mountains and Inyo Mountains, and in the Benton Range, between 6,000 and 9,000 feet. They bore nuts every other fall. Only a few small stands of oak grew along the eastern slope, mostly along Division and Oak creeks. The *nümü*, therefore, would trade salt and pine nuts with the Monache, who lived along the western Sierra, for acorns and acorn flour. Using bedrock mortars, the women would grind the acorns into mush and then leach out the tannic acid with hot water. The mush would be eaten at once or poured into cold water, where the mixture would solidify into cakes for later consumption. The *nümü* would finish their harvesting by mid-summer, when the families would return to their base camps. If the harvests were large, small groups would spend the winter in the piñon forests, where they would store their crops in covered, grass-lined pits. If the pine-nut crop failed, the people would spend the winter in the valley, where they would eat seeds (especially Indian rice-grass) that had been gathered during the summer and fall.

Another staple was the Owens River pupfish (*Cyprinodon radiosus*). Usually no longer than 2½ inches, the fish lived in the shallow and slow-moving waters of the river and its tributary creeks. By damming and poisoning streams, the nümü would catch many of the fish, which they would dry in the sun and eat during the winter.

The nümü also cultivated wild plants. Apparently, the small bulbs of nut-grass (*Brodiaea capitata*, in the sedge family), were the main crop, although the corms (enlarged bases) of wild hyacinth and the seeds of one species of spike-rush (genus *Eliocharis*) were important, too. The growing season would begin in the early summer, when the melting snowfields in the Sierra Nevada flooded the meadows. Using only their hands and digging sticks, the people in several districts would dam the creek and dig irrigation canals. Elected irrigators then would divert the floodwaters from one plot to another. In the fall, the people would destroy their dams, harvest the crops, and replant the bulbs.

The main areas of cultivation were Pine, Freeman, Baker, and Bishop creeks (between Independence and Bishop). The marshy meadows along lower Bishop Creek made up the largest district, in which a network of ditches was used to irrigate the night-grass. Two fields there covered at least 6 square miles. In the summer of 1859, a military expedition came upon mile after mile of domesticated nut-grass that was being irrigated with great care.

The Arrival of the Whites

The discovery of the Comstock Lode in what was then western Utah Territory attracted adventurers searching for another lode, the Lost Gunsight, which, according to legend, was fabulously rich in silver. In late 1860, Dr. Samuel G. George left Visalia with a small party that included William T. Henderson, W. B. Lilly, and E. McKinley. The men prospected in the Panamint Mountains. Three miles southeast of Wildrose Spring, Dr. George and Henderson found a thick, silvery vein on Christmas day, which they called the Christmas Gift. A snowstorm caught them, however, and they barely made it out alive. In Visalia, the ore yielded mostly antimony sulfide (stibnite) but it contained a tantalizing amount of silver. Dr. George decided to return in the spring. But first, Horace P. Russ organized a separate expedition, the New World Exploring and Mining and Company. The party left San Francisco in early March, 1861. One of its members was Oscar L. Matthews, who became a

[J] Bullion (Russ District) Gold and Silver Mining Company, Office No. 622 Montgomery Street, San Francisco, July 13, 1863.— Notice is hereby given, that on the Thirty-First DAY OF AUGUST NEXT, at twelve o'clock noon, there will be sold at Public Auction, at the Salesroom of John Middleton & Son, No. 400 Montgomery Street, a sufficient number of shares of Stock in the above named Company, standing in the names of the following persons, to pay the assessment of Ten Cents per share, levied March 23d, 1863, and the accruing costs of advertisement and sale thereof.

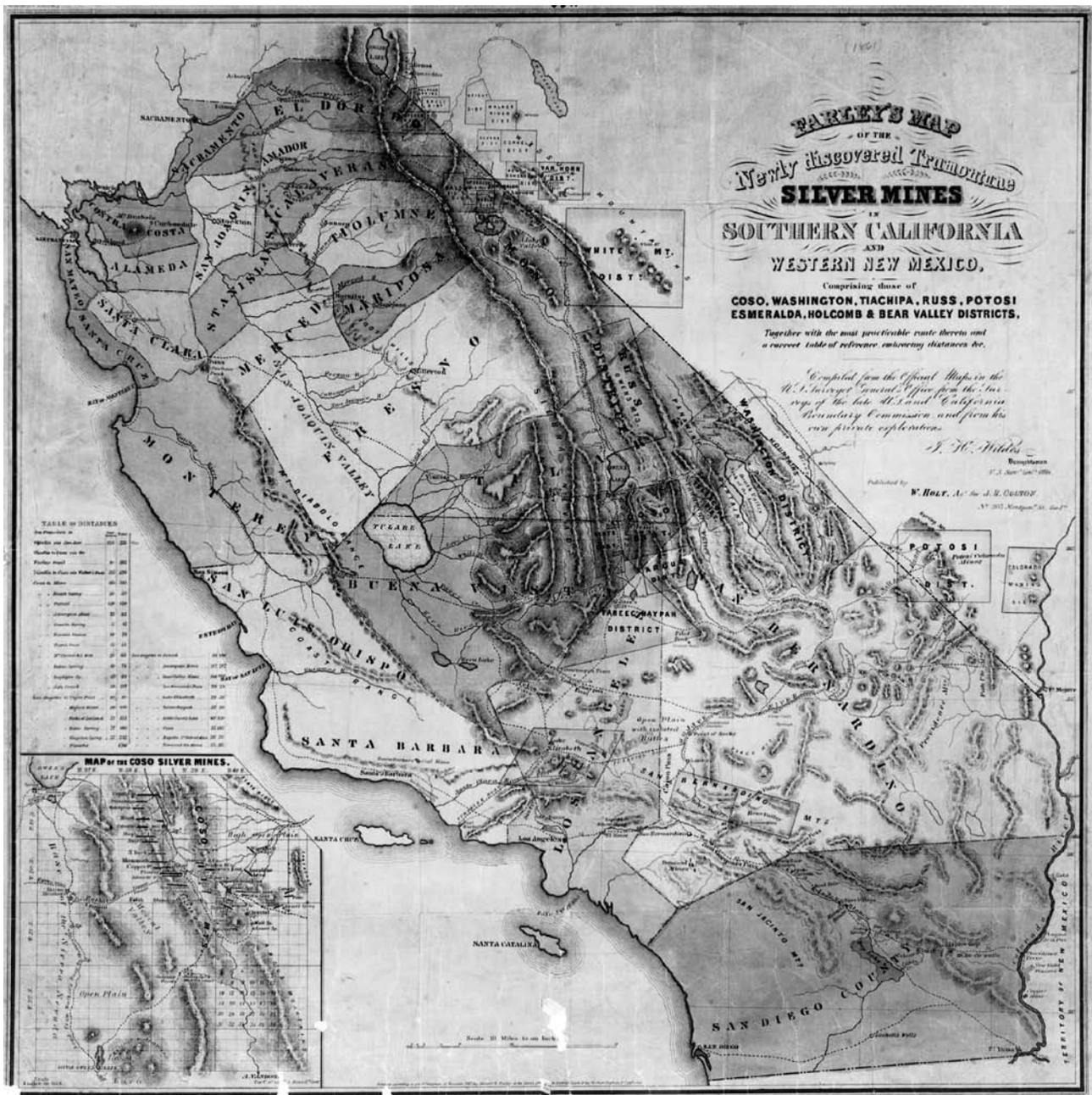
Names.	Shares.	Amount.
H C Badger.....	18½	\$1 85
Jas R Bell.....	300	30 00
Horace Bellows.....	18½	1 85
R N Berry.....	324	32 40
Z Birdsall.....	31	3 10
Miss Ellen S Booth.....	25	2 50
Wm Bosworth, Cert. 77.	150	15 00
J Boilleau.....	20	2 00
W J Bryn.....	15	1 50
A H Clark.....	60	6 00
Justinian Caire.....	26	2 60
T Campbell.....	15	1 50
Aaron Creamer.....	350	35 00
Mark Cornish.....	55	5 50
C C Dall.....	162	16 20
Peter Dean.....	15	1 50
E H Doty.....	643	64 30
W R Doty.....	150	15 00
Jas W Emburg.....	60	6 00
J F Greenman.....	21	2 00
Peter Goodhue.....	5	50
Amelia Goldstone.....	50	5 00
John N Harris.....	68	6 80
Wm Hewitt.....	20	2 00
Geo V Hight.....	150	15 00
Samuel Higgins.....	5	50
J Hirschfeld & Co.....	280	28 00
Samuel E Holcombe.....	156	15 60
Samuel E Holcombe.....	7	70
W W Lawton.....	50	5 00
O L Matthews.....	80	8 00
Laura Mitchell.....	10	1 00
G M Mower.....	10	1 00
C F Powell.....	138	13 80
Nelson Rolf.....	141	14 10
Isaac Rowell.....	100	10 00
Wm Role.....	25	2 50
Phoebe S Robertson.....	19	1 90
J H Slater.....	5	50
Timothy Sargent.....	20	2 00
Solomon Sweet.....	45	4 50
G R Spinney.....	5	50
Chas A Schmitt.....	310	31 00
John Wilson.....	70	7 00
Anna B Wilson.....	50	5 00
Miss Belle Wilder.....	5	50

By order of the Board of Trustees.
DAVID WILDER, Secretary.

July 23.

Im.

Notice of auction to pay assessments, Bullion (Russ District) Gold and Silver Mining Company. Visalia Delta, August 26, 1863: 1:6.



Farley's Map Of The Newly Discovered Silver Mines Of Southern California And New Mexico ... (San Francisco: W. Holt, 1860). Farley was Minard Farley, a mining promoter from Oroville, California. This is the first reasonably accurate map of the newly discovered mines of eastern California and what is now Nevada, which was then part of Utah and New Mexico territories. The dark areas are mining districts, which, in the original map, are overprinted in various colors. [Bancroft Library]

merchant in the Owens Valley. In April, Dr. George left Visalia with another company, which again included Henderson and Lilly. The two groups met at Walker's Pass and decided to combine. After the adventurers reached the eastern slope, they divided into two groups. Dr. George and his men explored the Panamint Mountains again. Upon their return, they rejoined Russ and entered the Owens Valley. They established a base camp just north of the later town-site of Independence. Using a field glass, Dr. George

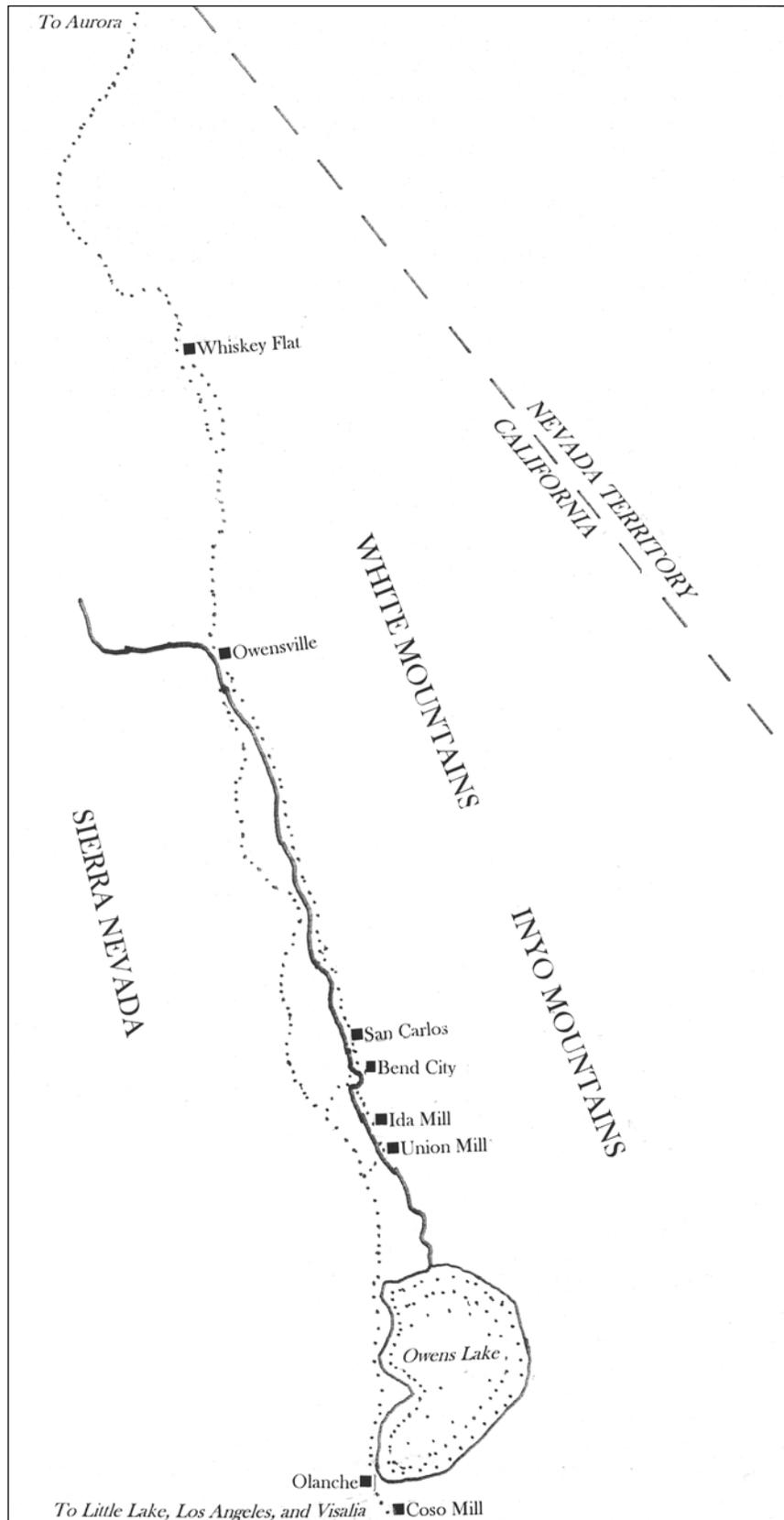
observed bold outcroppings in the Inyo Mountains, where the prospectors located the Union, Eclipse, and the Ida veins. In late April, they organized the Russ Mining District. Assayed in San Francisco in July, the ore yielded \$180 a ton in silver and gold, "liberally diffused throughout the quartz."

Meanwhile, the development of the mines at Aurora, north of the valley, led to another boom. To supply the market for meat, ranchers began moving cattle into the valley in 1861. The cattlemen let their

herds overrun the irrigated meadows used by the nümü and take over their creeks and ponds; the settlers and miners also cut down the piñons, which furnished the nümü with one of their main foods. The nümü customarily stored surplus food to tide them over during famines, but the destructive practices of the whites left them unprepared. Facing a man-made famine, the nümü began to fight back and even take cattle, but the heavy immigration continued. The fighting became so persistent that, in July 1862, the army established a post known as Camp Independence near Oak Creek. Even so, the attacks continued. In April 1863, many whites suspended work and moved into several fortified camps. When Captain Moses McLaughlin arrived with a detachment of soldiers in May, he distributed arms to the whites. The nümü ceased their attacks (although not their raids on cattle), and the farmers and miners returned.

By June 1863, prospecting parties were roaming the hills. They found gray silver sulfides (argentite) “shining with gold,” and even copper ore, all diffused “by the hand of luscious nature.” Several weeks later, Dr. George returned to Visalia with specimens of minerals. A correspondent in San Carlos, a new camp, reported in early September, “Our principal business has been prospecting. Everybody goes prospecting here. It is as fashionable for a man to put a pick over his shoulder and start for the hills as it is for a city belle to put on her silks and sally out for a walk on Montgomery street [in San Francisco] on a summer afternoon. . . .”

Prospecting soon gave way to mining and milling. Among others, Richard M. Wilson, the clerk of Mono County; E. S. Sayles, the new superintendent of the San Carlos



A guide to the settlements shown on Holt's Map of the Owens [Owens] River Country (following pages). Note: Montgomery City was laid out after the publication of this map.

mining company; and A. K. Warner and a well-supplied crew of carpenters and machinists arrived in September and October 1863. In February 1864, a correspondent in San Carlos gushed: "... Drills and powder are doing their work at a great rate, and the timid are often awakened from their midnight slumbers by an explosion loud enough to raise the head, from which they infer that night and darkness form no barriers to those intrusted [entrusted] with the development of the resources of this rich district. ..." About 10 miles north of San Carlos, for example, work was going on night and day in the tunnel of the Inyo mining company, which extended about 120 feet.

Mills and Smelters

During the 1860s, the reduction of silver and gold ores became increasingly sophisticated. In districts containing little timber, an all-metal crushing assembly known as the Bryan battery was popular. Although it was heavy, it was durable and easy to set up. If the ores were free of metallic sulfides such as galena (lead sulfide), pyrite (iron sulfide), and arsenopyrite (arsenic-iron sulfide), gold (and sometimes silver) could be extracted easily with mercury. The presence of metallic sulfides, however, interfered with amalgamation. In some districts, pyrite or arsenopyrite was the most common sulfide. For such ores, quartz mills containing special amalgamators were used. They consisted of steam-heated metal tubs in which close-fitting paddles, called *mullers*, rotated and mixed the ore, water, and chemicals, including mercury and salt. The tubs were called pans, separators, and settlers; Varney and Wheeler were popular brands. This system of reduction was called the "Washoe pan process." It proved effective at extracting gold, but not silver; a great deal of mercury was also lost. To extract silver from galena, which was common in some of the veins near San Carlos and at Montgomery, smelting was required, using lime as a flux. (Smelting later was used to reduce the silver-lead-zinc sulfides at nearby Cerro Gordo, far above the Owens Valley, and elsewhere in the Great Basin.)

LAKE CITY (ALLENDALE)

In November of 1862, W. B. Lilly and T. F. A. Connelly laid out a townsite named Lake City near the southern shore of Owens Lake. A creek ran through the property and timber stood only 8 miles away. Lake City served as a milling camp for the Coso Silver Mining Company, which owned a mine in the Coso

Mining Notices.

Union Gold and Silver Mining Company. Office 410 Montgomery street, San Francisco. July 28th, 1863.—At a meeting of the Board of Trustees held on the 24th day of June last, an assesment of One Dollar per share was duly levied on the capital stock of this company, and made payable immediately in U. S. gold coin, notice is hereby given, that on TUESDAY, the 1st day of September next, at 11 o'clock A. M., there will be sold at auction by E. S. Spear & Co., 433 California street, so many shares of the capital stock of said company, standing in the names of the following persons, as will be necessary for the payment of the assesment thereon, levied as aforesaid, and now due and unpaid, together with expenses of advertising and sale:

Names.	Shares.	Names.	Shares.
Samuel Aitker...	1	E Jacobs.....	25
T R Anthony...	34	W S Jewett....	100
L B Adsit.....	20	James T Jones..	5
Ellen S Booth...	25	A Johnson.....	50
F Blanchard....	350	J J Kely.....	75
John W Bell....	100	B F Libby.....	10
H Bellows.....	12½	B Lagourde....	30
D P Belknap....	25	Parfalt Lematre.	10
Thos Brown....	10	E C Morton....	7
R N Berry.....	100	Hall McAllister.	100
L A Brown....	25	Jos L Moody....	50
C J Bonsteil...	47½	Chas B Marvin... 200	
O H Beyring....	85	Horace Morrison	50
O P Blackman..	200	W H Mitchell..	10
Geo S Brown....	4	Henry Moor....	30
J W Brown....	8	J R Monley....	41
Lottie E Brown.	3	J R Mead.....	49
John Brazer....	25	J P Newmark....	10
F Boilleau....	100	C F Powell.....	50
C J Brenham....	50	Geo Proudfoot..	50
F S Carter.....	2	Jules Pellinot..	5
Aaron Creamer.	4	Peter Peterson..	73
J A Clark.....	5	N Rolfe.....	55
Mark Cornish..	25	E A Rowe.....	5
P B Cornwall... 627		H P Russ.....	7
J B E Cavallier.	4	Antonio R Ruge	5
John Cook.....	5	Isaac Rowell... 25	
J Cannin.....	40	Paul Roussett... 110	
E Cazales.....	9	Francis Rouge..	10
A B Coffin....	50	James Rothfeld.	5
C C Dall.....	36	Thos Richards.. 100	
C Druailler....	12	A B Rayner....	25
H F Dorrance..	25	C R Stinger....	1
John L Durkee.	5	R G Spinney....	1
W H Eaden....	10	B L Schmitt....	50
J O Eldridge... 155		A B Schalu....	50
Frank Eastman.	50	Chas Stephens..	20
G R Eckley....	30	E S Sleeper....	26
A P Fuller....	20	H Sullivan.....	20
Mrs Helen Foster	2	A G Turner....	41
Wm Fleming... 12½		Paul Forquet... 1	
C Grilliet.....	4	A O Thoms....	25
S Gregg.....	9	J D P Teller... 50	
W H Gray....	6	J H Voorhees.. 15	
S D Hassey....	4	Theo Vosin....	10
Joseph Haine.. 75		B F Whitson....	25
J B Hesbut....	50	J R Whiteman.. 5	
J J Haley.....	117	Jas W White... 15	
Fred Hall.....	49	Weins h e n k &	
F A Hassey....	35	Lankersheim.. 100	
S E Holcombe.. 127		W E Wood....	70
T L Horn.....	50	A Wolf.....	25
G W A Jones... 10		Edmund Worth.. 75	
E H Jacqueline.	25	L P Waller....	20
C C Jones....	50	J C Winans....	50
Geo Pen Johnston	50	Harry Wheeler.. 100	
J T Zorn.....	150	G Forest Walter.	42

By order of the Board of Trustees.
a13-td. JOHN B. RUSSELL, Sec'y.

Notice of auction to pay assessments, Union Gold and Silver Mining Company, Visalia Delta, August 27, 1863, 1:6.

Range several miles to the south. The mill probably was completed in the spring of 1863; its superintendent was Isaac S. Allen, for whom the place was sometimes called Allendale. The camp contained a house, a blacksmith shop, office buildings, a ditch, and a flume. Erasmus Darwin French also lived there. The Tulare County board of supervisors established an election precinct there in August, designating French's house as the polling place. Allen, who held a mortgage on the property, sued the company in August 1864, and gained ownership of the property, which was auctioned off in March 1865. The next owner was the Olanche Gold and Silver Mining Company, owned by Dr. Samuel George and Minard Farley. They mined for a while, but in early 1866 Farley moved the mill to Claraville in the Piute Mountains, near the Kern River.

UNION

The Union Gold and Silver Mining Company owned what would prove to be the richest lead, the Eclipse, at the foot of the Inyo Mountains. But the threat of attacks from the Indians was so great that the company left the machinery at Walker's Pass, about June 1862, awaiting the arrival of soldiers from Los Angeles to secure the route to the Owens River. In the summer, the Union shipped out a ton of rich silver and gold ore to San Francisco, apparently for sampling, and then completed a steam-powered, 8-stamp mill that stood 17 or 18 miles northwest of Owens Lake. The superintendent was Robert S. Whigham. In late December, Whigham brought \$10,000 in amalgam, the product of 9 tons of ore, to San Francisco. Problems with the mill's silver-recovery process, however, soon forced it to shut down, and attacks from the nümü forced all mining to suspend in April 1863. The Union and the nearby Ida mill, however, soon resumed work. (In August 1863, the Tulare County supervisors created an election precinct at the mill.) Although the shafts and tunnels of the Eclipse extended no farther than 100 feet, the company took out 500 tons of very rich ore in the summer and worked it in the Ida mill. The mill was re-equipped in December; afterward, work resumed round the clock. In late February 1864, a 75-ton lot of ore yielded at least \$9,000 in gold. The mill operated at least into August.

Ida

The Ida Gold and Silver Mining Company owned the Nevada lead, above the Eclipse, and a millsite 1½ miles south of the Union mill and 7 miles south of Bend City. One of the company's trustees was Robert S. Whigham, who was the superintendent of both

Ida Gold and Silver Mining Company,
Russ District, Tulare County, Office No. 522
Montgomery Street San Francisco, July 27th
1863.—Notice is hereby given, that on the THIRTY-FIRST DAY OF AUGUST NEXT, at 12 o'clock noon, there will be sold at Auction, at the salesroom of John Middleton & Son., No. 400 Montgomery Street, a sufficient number of shares in the above Company standing in the names of the following persons, to pay the assessments now due thereon, and the cost of advertising and sale.

Names.	Shares.	Assessments.
D Wallach (J J Loun)	12	\$ 36 00
Thos G Beasley	7	21 00
S G Gregg	33¼	100 50

By order of the Board of Trustees.
DAVID WILDER,
Sec'y

jj30-1m.

Notice of auction to pay assessments, Ida Gold and Silver Mining Company. Visalia Delta, July 30, 1863, 2:6.

the Union and Ida mills. In April, 1863, Whigham shipped out a boiler, an engine, four Wheeler pans, a miniature amalgamating apparatus, and a complete assaying outfit. After the arrival of Captain Moses McLaughlin and his troops in May, the company completed the mill and began shipping out ore. In early June, Whigham returned with 50 yoke of heavy cattle, several large wagons, and a large amount of freight from Visalia. Afterward, the Ida began working ore from the Eclipse. During the winter of 1863–1864, the company enlarged its mill to 12 stamps and drove an 80-foot tunnel into the Nevada.

THE OWENS RIVER CANAL

As an adjunct to the Union, Ida, and San Carlos mills, in August 1863, Robert Whigham and others incorporated the Owen's River Canal Company, which proposed to build a canal to transport ore and to provide power to the mills. Apparently, the company dug 2,700 feet. After the collapse of silver mining in 1864, Whigham hired a consulting engineer in San Francisco to report on the prospects for a longer canal and went to New York to sell stock in a new corporation. According to a prospectus, the company would dig the first section of the canal from San Carlos to the Eclipse Mine, about 10 miles away.

SAN CARLOS

The San Carlos Mining and Exploring Company began building a reduction works in the summer of 1863, when it dug a 2,700-foot ditch. A superintendent, S. E. Sayles, arrived in September. Milling ma-

chinery and the castings for a smelter arrived during the winter of 1863–1864. Using lumber cut along Oak Creek and brick and lime produced at the millsite, crews completed a smelting furnace in May and a mill by early July. The mill stood along the river, near Romelia and Silver streets, in San Carlos. The plant consisted of a steam-driven Bryan battery containing five 750-pound stamps; a copper-plated “apron” coated with mercury; four large settling vats; two Wheeler pans; and a Wheeler separator. Spare parts were stored in an adjacent blacksmith shop. Nearby was a well, which had been sunk 8 feet below the level of the river; a pump drew the water into tanks behind the boiler room. The mill started up on July 4.

CHRYSOPOLIS

The Chrysopolis was another mill, although information on it is fragmentary. In January of 1863, a prospecting party led by Robert Whigham struck several promising veins near the Owens River about 15 miles northeast of Ida Camp. In the spring, others found several nearby gold-bearing leads about 10 miles above San Carlos; one was named Chrysopolis, the Greek word for “golden city.” Whigham’s company soon located a millsite about 6 or 8 miles north of San Carlos at a narrow point on the Owens River where a dam could be built. A camp was reported there in September. Little else is clear: a mill was in fact built, perhaps by the Hooker and Burnside mining companies or by the Owens River Mining Company. Apparently, the plant operated only sporadically and shut down, perhaps in early 1864. A revival took place after the Civil War; a post office was established in May, 1866, but it soon closed in March 1867.

OWENS RIVER MINING COMPANY (SANTA RITA)

The Owens River Mining Company owned the Santa Rita claim in Mazourka Canyon. A half ton of ore was shipped out in late May of 1863. Meanwhile, the company began shipping milling machinery to an

unidentified millsite, and timbers for the building arrived several months later. Nothing else about the mill was reported. Discoveries in an adjacent lead, the Owens River, “made it a favorite” among stockbrokers in San Francisco in November; in one week, the price of the stock rose from \$40 to \$85 a share. According to one report, some of the ore was yielding as much as \$1,000 a ton in gold, which was easy to extract. Meanwhile, the price of shares in the Santa Rita rose from \$10 to \$25. The company later sank an 80-foot shaft on the Santa Rita lode, in which silver predominated at 70 feet.

CONSORT

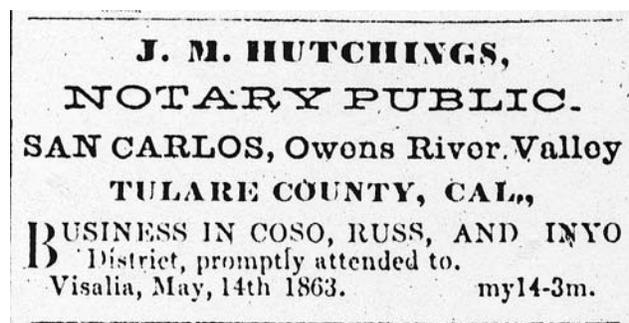
This might have been a phantom mill. According to one report, the Consort Gold and Silver Mining Company shipped out a 10-stamp mill to Owensville via Aurora in November, 1864. Most likely, the plant was never installed there.

Camps and Towns

At first, the only villages were those of the *nümü*. But on July 4, 1862, Colonel George Evans, commanding 157 men, established a military post near Oak Creek; he called it Camp Independence. Camps soon grew up at the Union and Ida mills. Several townsites also were laid out in the spring of 1863, but only three towns were built. The largest ones were San Carlos and Bend City, 3 miles apart, standing east of the present town of Independence. In the north end of the valley stood Owensville, 4 miles east of Bishop’s Creek near the Owens River. The railroad town of Laws was later built at the site.

SAN CARLOS

The founder of San Carlos was the San Carlos Mining and Exploring Company, which was incorporated in early February 1863. Among the stockholders were James M. Hitchens and Henry Hanks, the company’s assayer. Acting upon a motion by Hanks, the miners soon organized the Inyo Mining District. Meanwhile, the company laid out a townsite near the river about 8 miles north of the Union mill. The main street, aptly named Galena, ran only a few hundred feet from some of the claims. Attacks by the *nümü* forced all mining to suspend during the spring, but after the arrival of troops, life resumed. The first business in the camp appeared in May, when Hitchens began advertising his services as a notary public. In July, John Lentell and Oscar L. Matthews, a former member of the Russ expedition, shipped out a load of goods from Visalia and built a trading post. When the county supervisors created a precinct at San Carlos in Au-



J. M. Hitchens advertised his services as a notary public in San Carlos; the newspaper spelled his name wrong. Visalia Delta, July 2, 1863, 1:5.

gust, they designated the store as the polling place. By early September, the business district contained two stores; two butcher shops; two assay offices owned by Matthews and Hanks; Edward Kenson's express office; and a saloon. Most of the establishments stood along a single street. At first, many residents had to live in tents, but after lumber began to come onto the market, probably from sawmills along Oak and Big Pine creeks, substantial stone, adobe, and frame houses were built. By late 1863, the town contained at least 30 buildings, including four stores, owned by Matthews and Lentell, Hiram Ayers, Loomis Brothers, and H. P. Garland; and a physician, Homer L. Matthews, the brother of Oscar. J. L. Hobart and H. T. Reed, who owned a hotel and trading post at Little Lake, opened a branch store in March of 1864. Construction continued as late as April. A post office, with Henry Hanks as postmaster, was established in June.

BEND CITY

The only serious rival of San Carlos was Bend City. The settlement was founded about July 1863 in a bend of the Owens River between the foot of Mazourka Canyon and the later site of Independence. The townsite was laid out in a grid composed of 300x220-foot blocks, which were separated by streets 100 feet wide. Most of the buildings were made of adobe. Bend City developed rapidly during the summer. Among the businesses by early October were William S. Morrow's boardinghouse, an express office, a lager brewery, and, "of course, the usual quantum of chain-lightning and tangle-foot whiskey dispensaries." By December, Bend City contained at least 25 neat and comfortable—though widely scattered—adobe houses. By early 1864, the business district included five stores stocked with as much as \$15,000 in goods; two eating houses; two blacksmith shops; a barber shop owned by a "contraband," or freed slave, the only African-American in the valley; Uncle Abe's butcher shop; a cobbler's shop; a saddle- and harness maker; a tailor's shop; several bars or saloons; a stock exchange; a circulating library where books could be rented for 25¢; a laundry run by a Chinese couple; a community coffee mill that served "most of the unfortunate bachelors"; and a hotel, the Morrow House; meanwhile, a doctor from Aurora was building a drugstore. John R. Hughes & Company completed a second hotel, the Husseth Exchange, in late January 1864. Though somewhat dusty, Bend City continued to bustle. By March, at least 30 buildings stood there.

GRAHAM AND OWENSVILLE

Mineral deposits also were found in the northern Owens Valley, which was then in Mono County, where the Keyes Mining District was organized.

The first camp in the district was Graham, four miles from Bishop's Creek, below the mines of the Keyes district. The place was named after D. S. Graham, a mining superintendent. In December 1863, Graham contained several houses, including buildings used by two mining companies. From dawn till dark, a correspondent wrote, "the sound of the mason's trowel and the carpenter's plane can be heard, mingled with the music of the prospector's pick in the adjacent hills." Graham was soon renamed Riverside.

Graham might have served as the nucleus for Owensville, for the name appeared in late 1863. Owensville stood about 40 miles north of San Carlos at the foot of the White Mountains, near the present site of Laws.

Even though the market for mining stocks collapsed in early 1864, Owensville remained active for a while. A correspondent for the *Aurora Times* noted: "Owensville not only still lives, but has cast off her swaddling clothes, has ceased to creep, now walks alone, and before the fall winds shall strew the plains and valleys with the yellow leaf, will be full grown and prosperous. . . ." In April, in fact, nearly 23 pounds of ore from the Twin Brothers yielded \$520 a ton in gold and silver; by then, from 30 to 40 miners were at work in the district. Meanwhile, corner lots in Owensville were selling for as much as \$1,500, at least for a while. The business district included a saloon, store, and restaurant; from 15 to 20 other buildings made of stone, adobe, and lumber were under construction. Seventy-four voted in the precinct in early November. Owensville was still considered "a smart little town." As late as August of 1865, the place still contained 10 buildings, including two blacksmith shops, owned by J. M. Haskell & Company and George M Hightower; a hotel and bar, owned by J. Curry; J. B. White's saloon; and, perhaps, a store, owned by D. Disborough. A post office was even established in March 1866; Thomas Soper, the superintendent of the Consort mining company, was appointed postmaster.

LITTLE LAKE

Little Lake was the most important way station between Walker's Pass and the Owens Valley. In the summer of 1862, a "sojourner" built a stone hotel at Little Owens Lake, which contained fresh water and was home to a variety of waterfowl. In late May of 1863, the Little Lake Mining District was organized

**COSO, ARGUS, RUSS, INYO,
TELESCOPE, & SLATE RANGE
MINING DISTRICTS.**

TO PROSPECTORS, MILL OWNERS and others. The advertisers would respectfully call the attention of all parties interested in the country east of the Sierra Nevada's, to the fact, that they have in Store, at Little Lake a full and complete assortment of

**Staple and Fancy Groceries,
Provisions, Wines,
Liquors, Hard Ware,
Clothing, Boots,
Shoes, Steel,
Iron, Blasting Fuse,**

and everything in their line of general merchandise, while they are prepared to sell at Wholesale or Retail, at San Francisco prices, with the addition of cost of transportation. Particular attention paid to outfitting prospecting parties, and also to filling the orders of Mining Companies, and Mill Owners. To parties visiting our section of country. We would say that every thing required, can be purchased here, and at prices which will be satisfactory.

By strict attention to business, we hope to merit the patronage which we most respectfully solicit.

HOBART & REED.

Little Lake, June 11th, 1863. j18-63.

Advertisement for the Hobart and Reed store at Little Lake. *Visalia Delta*, July 2, 1863, 3:2.

there. J. L. Hobart, the chairman of the meeting, and a partner, H. T. Reed, operated a trading post there. In June, they advertised a complete stock of groceries, hardware, clothing, boots, shoes, and supplies, including blasting fuses. The place offered "ample and pleasant accommodations," a traveler reported in December. Since the station commanded all the trade from Visalia and Los Angeles, Hobart and Reed sold more than \$10,000 in merchandise in 1863. Leander Ransom, the editor of the *Delta*, considered Little Lake "the only civilized spot I've seen since I left Tulare Valley. To see the herds of cattle, sheep, goats, flocks of turkeys, chicken and pigeons; the sleek old cat and the playful kittens, the dignified and manly old mastiff and the frolicsome younger ones, makes one fancy himself on a well ordered farm in the best part of the Eastern States. . . ." The merchants also owned a sailboat; and, "there being a nice breeze," Ransom and Reed "had a delightful time" sailing on the lake.

Local Government

At least a dozen mining districts were organized east of the Sierra from 1860 through 1864, including Coso, Russ, Telescope, Inyo, Olancho, Keyes, Saline Valley, Argus, Slate Range, Lake, Alabama, Kearsarge, White Mountain, and Montgomery. But their powers were limited, and most of them failed to last.

Both Tulare and Mono counties maintained jurisdiction in the Owens Valley, in which Big Pine Creek served as the county line (later moved). In August, 1863, the Tulare County Board of Supervisors created four election districts east of the Sierra: Coso, Owens Lake, Owens River (at the Union mill), and San Carlos, and the Mono County supervisors established a precinct at Allen Van Fleet's ranch north of Big Pine Creek. Mono County later created precincts at Owensville, Montgomery, and Big Pine. To maintain some semblance of order, the counties also established townships, which contained two justices of the peace and two constables. (The judges also were authorized to hold coroner's inquests.) In December 1863, to save the residents a 100-mile trip to the county seat, which was then at Aurora, the Mono County supervisors appointed Will Hicks Graham, a

**OWENS RIVER STAGE
LINE AND EXPRESS.**

Via White River, Linn's Valley, Greenhorn, Keysville, Coso, Big Lake, Union Camp, and San Carlos, to Camp Independence.

The Stage leaves Visalia every Monday and Thursday mornings.

The Express leaves Visalia every Thursday morning, arriving at Camp Independence on the following Monday.

Leave Camp Independence on Thursday, arriving at Visalia on Wednesday night.

Letters, Coin and Packages safely carried, and promptly delivered, and all business properly connected with an Express line transacted in proper style.

WHITE, RICE & SCHLEIGH.

JAS. C. WHITE.

CHAS. RICE

CHAS. H. SCHLEIGH.

Visalia June 25th, 1863.

j25-63.

Owens River Stage Line and Express advertisement. The stage ran via White River, Linn's Valley, Greenhorn, Keysville, Coso, Big Lake, Union Camp, and San Carlos to Camp Independence. *Visalia Delta*, July 2, 1863, 2:6.

lawyer from San Francisco, as deputy county clerk and recorder for the southern part of the county. But for those living in Tulare County, the county seat was 250 miles away. In response to a petition, the state legislature finally authorized the formation of a county, named Coso, out of Tulare County's portion of the valley. The legislature, however, failed to give the legally required time before an election could be held, and the measure died. (Two years later, however, the legislature created Inyo and Kern counties.)

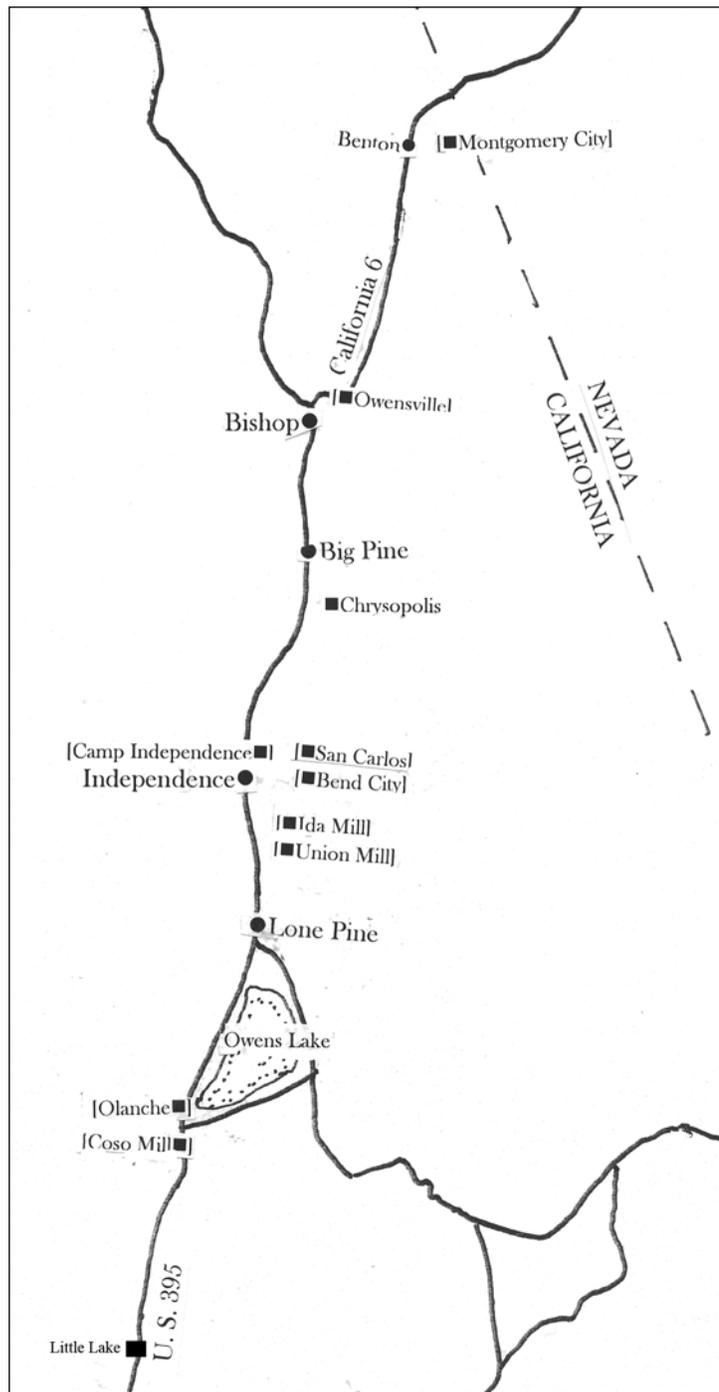
Travel, Transportation, and Communications

ROADS AND BRIDGES

Though the floor of the valley was fairly flat, the Owens River, which had steep banks, was an obstacle to freighting. In early December of 1862, the Mono County Board of Supervisors granted a franchise for the establishment of a ferry or the construction of a bridge at Graham's Crossing in the northern end of the valley, where the Owens River was swift and deep. Apparently, it was never built. But by April 1863, the San Carlos mining company was running a free ferry—the only one on the river. By late December, ferries were operating at the Union mill, 8 miles south of San Carlos, although teamsters considered its toll “a rather onerous tax,” and at Bend City. The ferry there was so crude—a tule raft attached to a rawhide line stretched across the river—that the citizens began to fear a loss of trade and began raising money and cutting timber to build a free bridge. The contractor, Thomas Passmore, completed the span in early 1864 at a cost of \$2,000.

STAGE SERVICE

For several years, travel to the Owens Valley was difficult. The lack of stage service, Robert S. Whigham complained in May 1863, forced travelers to carry their own blankets “and sleep out in the open air, or what is worse, sleep in a corral with the mules, and fight the fleas . . .” The coaches from Visalia, in fact, probably stopped in Linns Valley. In January 1864, Amos O. Thoms, a stage operator in Visalia, announced plans to run a coach to Bend City; most likely, the line was never established. In late July, however, J. M. Haskell, a blacksmith, began running a



Modern overview of early settlements.

stagecoach from Owensville to Aurora. And in early October, Daniel Wellington started a semiweekly line from Aurora to Montgomery; it was supposed to reach Montgomery in 13 hours.

EXPRESS SERVICE

To some degree, express lines made up for the lack of postal service. In June 1862, T. G. Beasley began running an express from Keysville, on the Kern River, to the Coso Mining District. Between May 1863, and

November 1864, a series of lines began running from Visalia (via the Coso Range) to the settlements of the southern and central Owens Valley, including Big Owens Lake, Camp Independence, Camp Union, San Carlos, and Bend City: James C. White and Charles (Charley) H. Schleigh; Thomas M. Heston; Thomas Derby and a partner named Nash (who also served the Slate Range Mining District); Charles Rice, James White, Charles Schleigh; Henry A. Bostwick (who made the round trip in 6 days); and Tom Tilly. The northern Owens Valley was more closely connected to Aurora. In January, 1864, Albert Dekay (also listed as Dokay) and Thomas K. Hutchison established an express line from Bend City and San Carlos to Aurora. In February, Edward Kenson moved the terminus of his line from Visalia to Aurora; the run took 30 hours. Josiah F. Parker, a mine owner, began running an express from Montgomery to Aurora in late September.

MAIL SERVICE

But express service, which cost senders and recipients from 25¢ to 50¢ apiece, was no substitute for mail service. In March, 1863, the state legislature urged its representatives in Congress to “use all honorable means” to establish a weekly route from Keysville to the Union mill via Walker’s Pass. Finally, a post office was established at San Carlos, with Henry Hanks as postmaster, in early June of 1864.

Domestic Life

Like the *nümü*, the whites were careful to observe the rites of life and death. When Gustavus A. Warner, a former miner from Tulare County, died of an infection in February 1864, the community gave him a dignified funeral. The mining company at San Carlos lowered its flag to half-staff, and its superintendent conducted “an impressive Episcopal service” at which he delivered some “touching remarks.” The first marriage in the valley took place that same month. In July, sons were born to Mr. and Mrs. John W. McMurry in Owensville, and to the wife of Levi Dean, 58, at the Owens River.

Lavish dances were popular on Christmas Eve, Washington’s Birthday, May Day, and Independence Day. A ball at a hotel in Bend City on Christmas night of 1863 “passed off in a style that would have done credit to older and more noted localities.” In the northern valley, the first social event was a dance given on May Day, 1864, in a small adobe cabin in Owensville. Independence Day was the most festive holiday. At Owensville, in 1864, about 150 Unionists and seces-

sionists alike crowded into a large stone building where they heard, first, an oration delivered by Will Hicks Graham, a lawyer; then the traditional reading of the Declaration of Independence; an invocation delivered by Thomas H. Soper, the superintendent of the Consort company; and the reading of a poem written for the occasion. Joined by Northerners and Southerners alike, seven women sang patriotic songs while Soper played his portable melodeon. A sumptuous banquet at the Owensville Hotel concluded the festivities. Equally elaborate was the celebration at San Carlos that day. At dawn, a militia battery fired a salute “to the great danger of windows and discomfort of late sleepers.” Gaily dressed residents watched a parade of military companies, as martial music was played and gleeful children set off firecrackers. The procession marched to a grove and presented the orator with a bouquet, listened to the blasts of the steam whistles at the mills for two hours, danced away the evening, stopped for a sumptuous supper at midnight, and then resumed dancing nearly until dawn.

Isolated as they were, the people liked to read. Circulating libraries were started at San Carlos and Bend City, where “cheap literature is dispensed at reasonable rates.” Henry Hanks owned the largest private library in the valley, comprising a well-selected collection of professional and other books. Meanwhile, the express companies were bringing in an “astonishing” number of newspapers, especially the *Sacramento Union* and the weekly edition of the *Alta California* (and probably the *Visalia Delta*). In March 1864, Leander W. Ransom, the publisher of the *Delta*, announced his intention to establish a newspaper at Bend City. This was the *Owens River Valley Herald*, which proposed to report on both mining and farming, carry domestic and foreign news, and favor the “Constitution as it is, and the Union as it shall be when this infernal rebellion is crushed and peace once more reigns within our borders.” Apparently, the first issue appeared on June 11. The *Herald* was quoted only once again, in late August.

The Development of Agriculture

Even the earliest prospecting parties could see opportunities for farming. A 4-mile strip of well-watered, fertile soil extended along the foot of the Sierra Nevada. Though sagebrush, bunch grass, and scattered mesquite covered the surface, creeks watered the land, which could produce a wide variety of crops. At first, the main agricultural areas were Round Valley, Bishop’s Creek, and Big Pine, where many of the *nümü* lived.

Widespread settlement began in the spring of 1863. By May, despite the threat of attacks from the *nümü*, farmers and ranchers were streaming into the valley, bringing their cattle, wagons, tools, and families. They were “men of the right stamp,” intending to live there permanently, some to farm and others to mine. By September, settlers had claimed more than 15,000 acres. In early December at Adobe Meadows, north of the valley, one correspondent came upon an encampment of 56 men, 82 yoke of oxen, 15 wagons, 14 teams of horses, many saddle animals, and a family with “four tow headed little urchins.” During the winter, the settlers recklessly began burning the dry grass to prepare for their spring crops, although the fires also destroyed most of the natural feed needed by the livestock and even killed many of the animals themselves. As one correspondent approached Camp Independence in early March of 1854, “little adobe and sod houses seemed to spring up out of the ground”; he counted 43 dwellings between the Union mill and Black Rock, a townsite 15 miles north of the Ida mill. “. . . All are busy; some ditching, some fencing, a few ploughing; others planting vineyards. The canons [cañons] of the Sierras resound with the echo of falling timber, as they labor to get out material for fencing, or, chop the fallen logs into fire-wood. Some are burning coal, while others are whip-sawing lumber. . . . We are beginning to love these mountains, and have long ceased to be surprised that the Indians fought for the valley. . . .” During the summer, ranchers from the drought-stricken western Sierra drove their starving cattle into the valley, where tall grass was growing in abundance.

Usually, supplies were plentiful. In early January, 1864, two mercantile companies in Aurora shipped out a total of 38,000 pounds of freight. In the meantime, while the *nümü* starved, flour, potatoes, and beef were available at Bend City and Camp Independence. Some travelers, according to one report, even picked off cows in the large herds. During the summer, the farmers at Bishop’s Creek and in Round Valley were selling butter, hay, and a great variety of vegetables, including peas in the pod, turnips, cucumbers, radishes, ears of roasting corn, and potatoes.

Lumber, however, was scarce. For a while, all of it had to be cut by hand, using whipsaws. In September, 1863, several teams were sent to Aurora to bring back lumber. Four loads from Mono Lake soon arrived at San Carlos. Some whites lived in tents, although most of them built their houses of adobe bricks. At San Carlos, a former mining superintendent and his wife lived in a house made of tules; it contained two “spa-

cious and comfortable rooms.” In early 1864, several businessmen in San Carlos completed a sawmill on Big Pine Creek, about 30 miles north of Oak Creek, and in the summer another company built a sawmill on Bishop Creek.

The Decline

Investing in mining then was risky at best. Try as they might, no inventor or engineering genius could perfect the Washoe pan process, especially in the recovery of silver. (Other methods, such as chlorination and smelting, were effective, but they were slow and expensive.) Many of the corporations engaged in shady practices. One form of fraud was the “freeze-out” scheme, in which corporate officials would levy frequent or heavy assessments, sometimes to develop their mines but too often to enrich themselves. If a stockholder fell in arrears, the company would auction off the shares to other suckers. Auctions also allowed company officials to buy the stock at bargain prices and increase their control. The Ida company levied five assessments and then, in January of 1864, auctioned off the delinquent shares. The Chrysopolis mill was shut down about then. The “rush has been by no means as great as was expected by some,” one resident conceded in April. The lack of a competent amalgamator idled one of the valley’s best mills, possibly the San Carlos, where Henry Hanks was appointed to take the place of E. S. Sayles, in August. Considering the “universal mining panic,” a visitor bluntly declared in September, “the whole country is ‘gone in’—it may be for years . . .” Bands of Indians were still driving miners out of the White Mountains in the fall. By December, most of the miners were gone. “. . . No mills in the country, tend, of course, to no work—and selling feet is played out.”

The mining towns, meanwhile, were turning into ghost towns. When William Brewer and his geological party arrived in Bend City in July 1864, he found “a miserable hole, of perhaps twenty or twenty-five *adobe* houses, built on sand in the midst of the sagebrush, but there is a large city laid out—on paper. . . .” The camp was dull, the heat was intense, and the residents were fearful of attacks by the *nümü*. (In fact, they burned the Union mill in June, 1865.) By January 1866, the only people left in San Carlos and Bend City were soldiers stationed there to prevent the Indians from burning the buildings and the Ida mill. The post office at San Carlos was discontinued in late February and was moved to the new town of Independence. A masonry smokestack stood there for a long time.

When the geologist W. A. Goodyear visited the valley in the spring of 1888, the Union mill was gone, and the Ida and San Carlos mills, their machinery still intact, were falling into ruin. He counted the ruins of 33 houses at Bend City and 26 at San Carlos; “but now they are entirely deserted, and there is not a soul living there, nor a roof upon a house at either place. . . .”

Several camps managed to hang on for a while longer. By August, 1864, Owensville had become very dull. The Consort mining company sold its blacksmith shop, which was moved to Bishop Creek where it became the first building there. Other buildings were torn down and the lumber was rafted down the Owens River for use at Independence and Lone Pine. By August 1865, only 10 buildings remained. Perhaps the camp enjoyed a revival, for a post office was established in March, 1866; Thomas Soper was the postmaster. After a brief change of name, the office was moved to Bishop Creek in February, 1870. Owensville lost its last resident in 1871. At Lake City, Isaac Allen, who held a mortgage on the property, sued the Coso company in August 1864, and was awarded a judgment. The property was auctioned off in March, 1865. The next operator was the Olanche Gold and Silver Mining Company, owned by Dr. Samuel George and Minard Farley. They mined for a while, but in early 1866 Farley moved the mill to Claraville, in the Piute Mountains near the Kern River. At Chrysopolis, the mill shut down in early 1864, although the operation enjoyed a brief revival. A post office operated there from May 1866, to March 1867.

Meanwhile, the valley developed a much more stable and solid economy. Beginning in 1864, fairly large ore deposits were found in the Kearsarge Mining District, west of Independence; near Benton Hot Springs, to the north; and at Cerro Gordo, east of Owens Lake. The first buildings at Lone Pine, Bishop's Creek, and Big Pine, all centers of agriculture, appeared in 1864. Independence was laid out in 1865 and became the seat of Inyo County in 1866. But by then, the Kern River rush was over.

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Regional Histories

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memories of many of the early settlers; his history touches on everything in this article.

The People

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Montgomery City

Alan Hensher, 593 Collins Drive #3, Merced, CA 95348

For those unable to make their pile elsewhere, they could look forward to another excitement. In early 1864, perhaps in February, rich “monster ledges” were discovered in the northwestern foothills of the White Mountains. Henry Hanks, an assayer at San Carlos, was one of the earliest claimants.

The site offered several advantages. The strikes were made near the mouth of a canyon, at about 6,500 feet. Aurora, the nearest supply center, stood about 55 or 60 miles to the north; Owensville stood about 30 miles to the south. The winters were milder than at Aurora, and the summers were cooler than in the Owens Valley. A creek ran down the canyon, and extensive stands of piñon grew in the vicinity.

Though specimens full of wire silver, assaying several thousand dollars a ton, were displayed in Aurora and San Carlos in April and May, a boom failed to develop at first. But in the summer, the mines at Aurora began to lay off workers. Although the *Esmeralda Union* considered the reports “somewhat exaggerated,” a rush began in early September, when one of several lots of samples assayed \$13,378 a ton, mostly in native silver. The excitement stripped Aurora of its wagons and animals, even skeletal “emigrant horses, scrag-tailed mustangs, galled mules and burros,” at exorbitant prices. By late September, despite “the stringency of the money market,” some claims were commanding high prices: a share in the Gorgonia, the leading claim, was selling for \$20. “For the last week about half of the town has been & still is in a great statement of excitement about the Montgomery District,” Laura Sanchez, the wife of a banker in Aurora, wrote in early October. She hoped that the reports of “exceedingly rich” ore were true, “for the sake of a great many poor men who have gone there. . . .” A mining district was organized several weeks later. In the meantime, the Mono County Board of Supervisors organized an election precinct there.

In late September, a camp named Montgomery City began to rise. Richard M. Wilson, a surveyor from Aurora, laid out a townsite. Thirty-one cabins were built between mid-October and mid-November. By late November, 84 lots in town had been sold. Although a fierce rain and snowstorm soon knocked down many of the camp’s “rude huts and canvas houses,” the miners were prepared to spend the winter there and rebuilt better dwellings. Meanwhile, about

half a dozen families had arrived, and others were on the way. “. . . Little children also toddle around our streets, and men begin to talk of the necessity of a school house.”

Apparently, the business district was very small. Among the establishments there by late November were two saloons: the Exchange, owned by Peter B. Comstock, and the Pioneer, operated by Patrick Reddy, who would go on to become a noted lawyer. The camp already had one lawyer: William Hicks Graham, the former recorder of the Keyes Mining District, at Owensville.

The best-known business, however, was its newspaper. After the re-election of Abraham Lincoln as president, Robert Ferral, the editor of the secessionist *Aurora Times*, suspended the newspaper, moved a small press and some old type to Montgomery City, and, on November 26, published the first issue of the *The Montgomery Weekly Pioneer*. The *Pioneer* contained four pages, measuring 6x8 inches. Ferral was the editor-in-chief, local reporter, compositor, pressman, devil (assistant), and carrier. “. . . We have no apologies to offer for the size of the paper. It is big enough for the winter season in [a] camp not yet three months old. . . .” Ferral sold his entire run that day and had to print a second edition, of 200 copies. Among the first advertisers were the Montgomery Smelting and Refining Works and Will Hicks Graham. Ferral’s former rival, the *Esmeralda Union*, in Aurora, called the *Pioneer* “a neat little paper.” In San Francisco, the *Daily Alta California* was astonished to receive a “bantling” newspaper from a “bantling town,” adding, “But Gods! What a sheet!”

Though the camp never had a post office, it did enjoy other services. In late September, Josiah F. Parker, a mine owner, began running an express from Montgomery to Aurora, and in early October, Daniel Wellington began sending out a semiweekly coach from Aurora; it was supposed to reach Montgomery in 13 hours.

The ore bodies, which contained various amounts of native silver, silver glance, copper, and antimony, were too complex to allow a single method of reduction. The native silver, usually called *wire silver*, could be extracted in an arrastra. Silver glance, now called argentite, was a sulfide, which required the use of the Washoe pan process. Copper- and antimony-bearing

ore had to be smelted. A miner, Frank Hutchinson, built the first smelter, in early October. Next, Dr. Daniel Munckton, who owned drugstores in Carson City and Aurora, formed a partnership with Charles Isenbeck, a reputed chemist from a mining college in Germany. They built a smelter—possibly the small furnace erected by the Montgomery Smelting and Refining Works—and Isenbeck produced two silvery bars, weighing 55 pounds, in early December. As it turned out, the bullion contained nothing but the basest of metals (possibly lead). After nearly being lynched, Isenbeck headed to Aurora and other camps, defrauding others along the way. Fortunately, the next smelterman, a Spaniard hired by Sol Carter and others, was honest. He completed a furnace in early January of 1865, but the ore lacked galena (lead sulfide), which was essential to smelting. Carter put up an improved furnace, which produced two silver bricks, in February. A cold snap, however, soon forced it to shut down.

Some mine owners didn't even bother to work their ore. In early January, a ton of very rich ore was sent to Liverpool, England, for reduction. Several weeks later, two large lots from the vein were shipped to San Francisco, where one sample yielded as much as \$1,800 a ton.

Other miners milled their ore. Charles Plummer crushed 10 pounds of ore in a hand mortar, amalgamated it, and sent the amalgam to Aurora, where it assayed \$43, or \$8,560 a ton. A few weeks later, a miner identified only as Morton improvised a hand-cranked two-stamp mill, using mercury flasks for stamps; a small belt, connected to the crank, turned the mullers in an amalgamating pan, which stood atop a brick furnace. Two men in Aurora completed a light two-stamp mill, in early February; each stamp weighed only 100 pounds. They were waiting until spring, when freight rates would drop, to ship it. About then, Daniel Munckton and another partner began operating an arrastre and amalgamator; reportedly, their plant was "paying handsomely."

The pockets of the richest ore were probably very small, but they were rich enough to support several small-scale miners and their families through much of 1865. Despite the winter, enough miners remained to cast 99 votes for district recorder in mid-February. (Patrick Reddy won the election.) One attrastra recovered \$684 from a ton of silver ore as late as June.

By then, however, the district was fading. In August, Montgomery City contained only seven houses and two arrastras. Peter Comstock moved to Benton, a new mining camp near a noted hot springs. The

county supervisors abolished the election precinct after April, 1866.

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The Blue Chert Mine: an epithermal gold occurrence in the Long Valley Caldera, California

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Introduction

The Vista Gold—Long Valley Property, also known as the Blue Chert Mine, lies within the Inyo National Forest, 8 miles to the east-northeast of Mammoth Lakes, CA (Fig.1). The prospect is situated near the center of the Long Valley caldera, 3 miles north of U.S. 395. Access is via well maintained Forest Service roads.

Standard Industrial Minerals, owner of the nearby Huntley kaolinite mine, was the first to report gold on the property in the early 1980s. Standard optioned the property to Freeport Minerals in 1983. Freeport undertook a drilling program outlining three zones of mineralization. Royal Gold leased the property from Standard in 1988 and continued the ongoing drilling program. By 1997, over 600 reverse circulation drill holes had been completed. After an aborted joint venture with AMAX the option reverted to Standard Industrial Minerals in August, 2000. In 2003, Vista Gold purchased the claims from Standard. Total measured and inferred gold reserve is 68M tons @ 0.018 oz/ton

(Source: Technical Report, Long Valley Project, Mono County, California USA, Vista Gold Corp., 2008). As of the summer of 2008 Vista Gold had undertaken no further exploration, nor had they green-lighted development.

The author wishes to acknowledge Neil Prens and Thomas Dyer of Mine Development Associates (MDA), the source of much of the technical information on the property. Their excellent technical report is available on line at [http://www.vistagold.com/technical_reports/Long Valley Preliminary Assessment, January 9, 2008.pdf](http://www.vistagold.com/technical_reports/Long_Valley_Preliminary_Assessment_January_9_2008.pdf). I would also like to thank Dr. Steve Lipshie, L.A. County Department of Public Works and Bob Reynolds, LSA Associates for their editorial comments and suggestions.

Geologic setting

The oldest volcanic rocks associated with the Long Valley caldera are the Glass Mountain rhyolite flows. They were extruded along a ring fracture system bordering the northeast margin of the caldera from

2.1 Ma to 1.3 Ma. Draining of the underlying magma chamber resulted in the initiation of subsidence at 1 Ma and culminated with eruption of the Bishop Tuff, at 760,000 B.P. (Van den Bogaard and Schirnick, 1995). Further catastrophic collapse ensued, creating an elliptical depression measuring 10 miles north-south by 18 miles east-west. Caldera subsidence totaled 2 miles of which 3500 feet is reflected in the present topographic relief, the remainder buried by post-caldera basin fill.

Following caldera subsidence, the central part of the basin underwent resurgent doming. Rhyolite and rhyodacite were

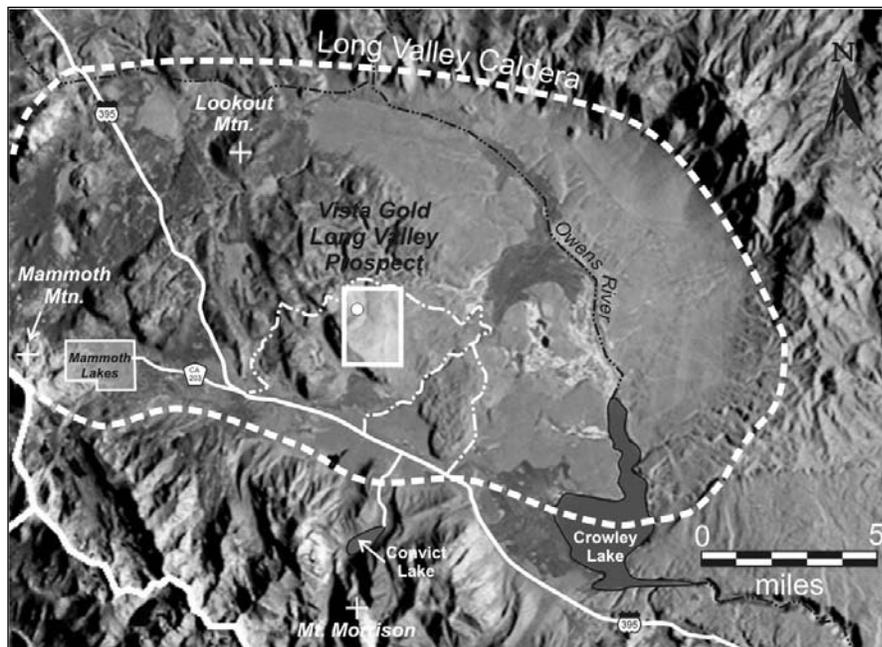


Figure 1. Index map showing the location of the Vista Gold—Long Valley (Blue Chert) prospect. White circle shows the approximate location of the Blue Chert Mine.

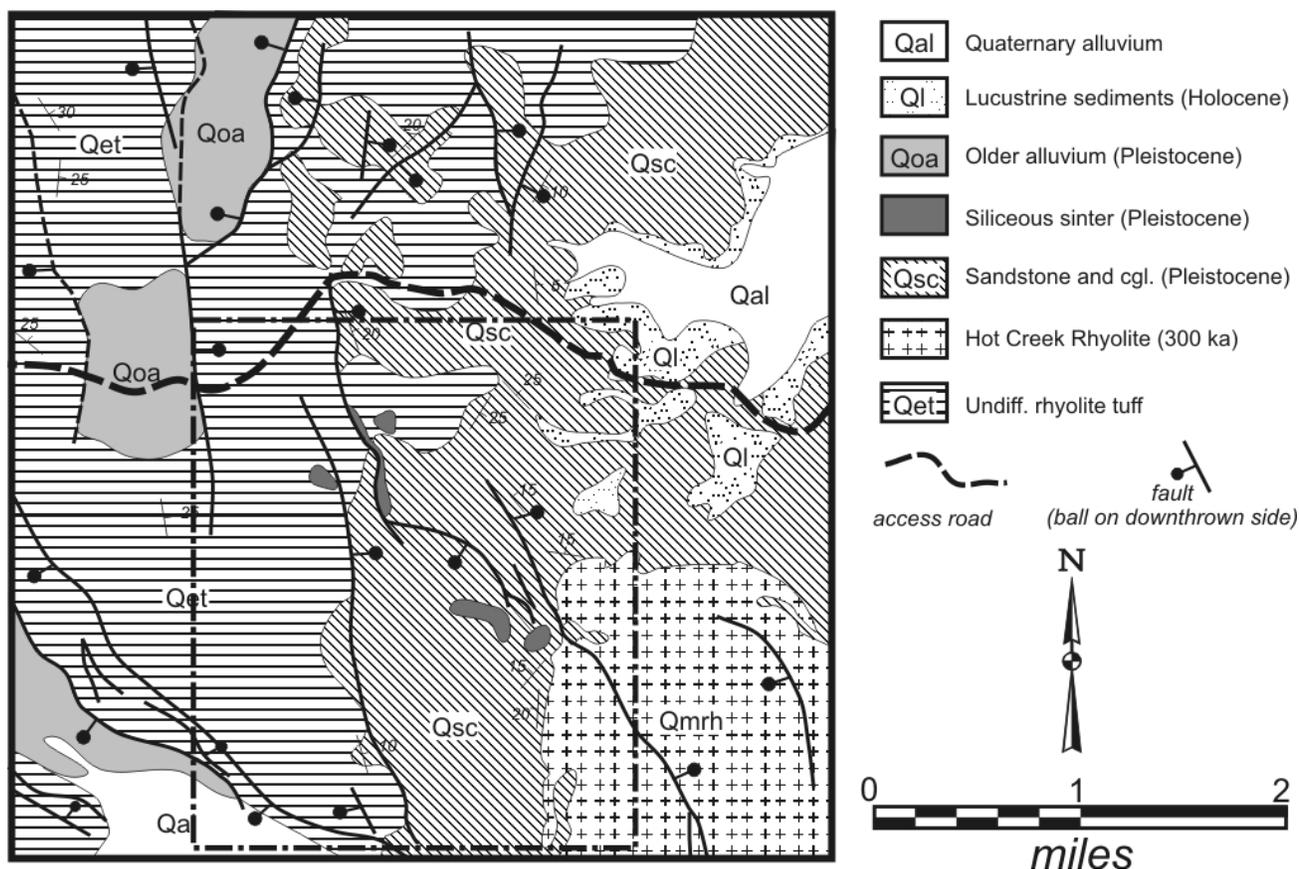


Figure 2. Geologic map of the Blue Chert property. Dashed line shows the approximate outline of the claim block. Geologic map adapted from Bailey, 1989.

emplaced from 12 vents during an interval of approximately 100,000 years, beginning 680,000 years ago. Evidence suggests that rhyolite doming had ceased by about 510 ka. The next phase of volcanism involved the emplacement of moat rhyolites in the ring fracture system around the periphery of the caldera in three separate events at approximately; 500,000, 300,000 and 100,000 years B.P. A later stage of volcanism produced rhyodacites from at least ten vents in the western part of the caldera. The main mass of hornblende-biotite rhyodacite is Mammoth Mountain. The Mammoth flows range in age from 180,000 to 50,000 years. Overlapping the rhyodacite flows are basalt and andesite extruded along the west rim of the caldera. The ages for these rocks ranges from 220,000 to 60,000 years B.P.

The most recent eruptive activity has been Holocene rhyolite volcanism that formed the Inyo Craters and domes in the northwest quadrant of the Long Valley. These volcanic features appear to be aligned along

a north-trending fissure extending from the Long Valley caldera to the Mono Craters. Activity within this zone has occurred sporadically for at least the last 40,000 years; the last eruption about 200 years ago.

The Blue Chert property, near the center of the caldera, is underlain by post-Bishop Tuff volcanics and sediments related to caldera formation and subsequent resurgence (Fig. 2). The oldest volcanics (Qet) underlie much of the western half of the property. They are composed of light gray to white, massive,



Figure 3. Fumarolic mound marking the blue chert outcrop. View looking southwest.



Figure 4. Kaolinite boulder veined by quartz, Huntley Kaolinite Mine.

rhyolitic tuff containing xenoliths of Bishop Tuff and granitic and metamorphic basement. Age is uncertain, but the unit must post-date Bishop Tuff emplacement. Depositional environment varies from subaerial along the flanks of resurgent domes to subaqueous throughout much of the caldera. Drilling suggests a maximum thickness of 1800 feet (Bailey, 1989). The southeast corner of the property is underlain by the Rhyolite of Hot Creek (Qmrh). It is a massive, pink, rhyolite containing only sparse phenocrysts. Locally, it has been bleached and hydrothermally altered, as along Hot Creek, and cut by siliceous hydrothermal vents. Mankinen, et. al (1986) have published K-Ar ages of 330-290 ka.

Much of the eastern half of the property is underlain by lacustrine detrital rocks that vary in composition from siltstone to poorly sorted sandstone and conglomerate (Qsc). Rock fragments are composed mainly of rhyolite and rhyodacite from the surrounding volcanic domes. This unit is thought to post-date resurgence, suggesting an age of 500 to 100 ka. Holocene alluvium, colluvium and lacustrine silts and clays occur locally along the eastern and southwestern margins of the property.

Perhaps the most interesting rocks are several siliceous sinter domes near the center of the property. These units mark the locus of fossil fumarolic activity that was most likely responsible for the gold deposition. They appear as a series of low hills or mounds (Fig. 3) that stand in relief due to their high silica content and resistance to erosion. Silica sinter is porous, fine-grained to cryptocrystalline rock composed mainly of opal, jasper, chert and chalcedony.

The north-south trending Hilton Creek Fault Zone (HCF) defines the eastern margin of the resurgent dome within the central part of the Long Valley Cal-



Figure 5. Closeup of the blue chert outcrop. Chert (gray) crosscut by secondary veins of quartz.

dera, continuing southward beyond the caldera rim. Clark and Gillespie (1993) state that movement along the HCF fault averages 1-2 mm/yr and that it can be characterized as a range-front normal fault, east side down. Historic earthquakes along the fault are numerous, the most recent a M=5.5 in July, 1998. Splays of the fault crosscut the property; two in particular appear to control the location of silica sinter mounds and have acted as conduits for the gold-bearing fluids.

Ore deposit

Five zones of low grade gold mineralization have been delineated on the Long Valley Property, termed the North, Central, South, Southeast and Hilton Creek. The latter three are part of a contiguous arcuate block with a strike length exceeding two miles. The ore bodies are flat-lying to gently (10-15°) east dipping with an average width of 1000 feet and thickness of 125 feet in the Hilton Creek zone and 75 feet in the Southeast zone (MDA, 2008). Mineralization in the South and Southeast zones is exposed at the surface, while the Hilton Creek zone lies beneath 20 to 50 ft of alluvium and colluvium.

The ore bodies have been oxidized to an approximate depth of 200 feet, roughly coincident with the current water table. Ore grades do not appear to be affected by the red/ox boundary. Mineralization is continuous throughout the ore zones; however, numerous high grade pockets (0.050 oz Au/t) exist, particularly along the Hilton Creek zone. The high grade zones are coincident with fault intersections or increased fracture density. Host rock lithology does not appear to be a major controlling factor in the distribution or grade of gold mineralization.

Mineralized zones correlate with areas of advanced argillic alteration and/or silicification. The predomi-

Table 1. Some Characteristics of Epithermal Deposits in Volcanic Rocks

	Low Sulfidation	High Sulfidation
Structural Setting	Commonly within a caldera	High level, crater of a volcano
Size	Variable, but some are large	Quite small
Host Rocks	Rhyodacite	Rhyolite to andesite
Timing of Ore and Host	Ages distinct	Similar
Mineralogy	Pyrite, native gold, base metal sulfides	Pyrite, native gold, enargite, Cu sulfides
Alteration	Potassic (adularia), sericitic	Advanced argillic, alunite
Salinity	Low	Moderate to high
Source of Fluids	Meteoric	Meteoric/magmatic

nant clay mineral is kaolinite with lesser alunite (Fig. 4), while the silicification consists of chalcedony, chert, quartz, or opal (Fig. 5). Multiple periods of brecciation and silicification have occurred, as evidenced by cross-cutting veinlets and silicified breccia fragments.

As stated previously, ore grade mineralization appears to be geometrically and spatially related to faulting. Splays of the north-south trending Hilton Creek fault zone can be projected to trend through the central part of the Hilton Creek and Southeast zones. These faults acted as conduits for the ascending hydrothermal fluids, which then flowed laterally into the more porous and permeable horizons.

The predominant gangue sulfide in unoxidized ore is pyrite, with lesser arsenopyrite. Sulfides occur as disseminated grains and clusters. They are commonly fine-grained (5 to 70 microns) and locked in a matrix of silicate minerals including quartz, feldspar and kaolinitic clay (MDA, 2008). Pyrite is poorly crystalline and generally framboidal, with lesser well crystallized, euhedral grains. Gold occurs as disseminated 1-6 micron grains, but more commonly is locked within the framboidal pyrite. Visible gold has not been observed. A significant portion of the gold resource is present in material which has been at least partly oxidized. The pyrite is altered to iron oxides (goethite) releasing the gold as submicroscopic grains.

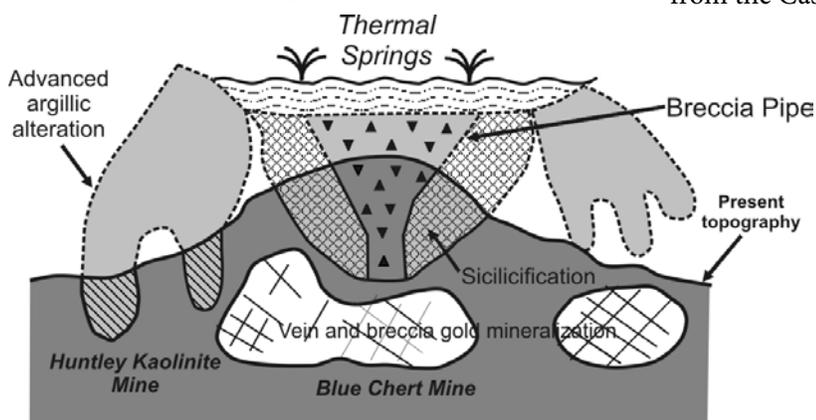


Figure 6. Idealized model for the Blue Chert Mine. Based upon the model for Round Mountain, NV of Tingley and Berger (1983).

Discussion

The association of gold with fossil hot springs systems has been recognized for over half a century. In the 1980's the U.S. Geological Survey focused their resources on epithermal gold deposits to better characterize and model the mineralization. Table 1, modified from Hayba et al. (1985), subdivided epithermal deposits into

two groups; low (quartz-adularia) sulfidation and high (acid-sulfate) sulfidation. This grouping was based largely on geochemical constraints. Complexes increase the solubility of gold dramatically; cyanide is the best known example. Of course, natural cyanide solutions are unknown. However, bisulfide (HS^-) can also increase the solubility of gold significantly under reducing conditions and chloride-rich brines can have a similar effect in neutral or near neutral pH solutions. The importance of these two complexes results in the two-fold breakdown of epithermal gold deposits. High sulfidation or acid-sulfate gold deposits are those in which gold has been carried as a bisulfide complex, while low sulfidation or quartz-adularia deposits are those dominated by chloride complexes.

MDA (2008) states that "the type of gold and silver mineralization present at the Long Valley property appears to fall under the general classification of an epithermal, low sulfidation type of deposit". This is consistent with a structural setting within the Long Valley caldera, an association with geothermal (meteoric) water, the low sulfide content and the moderate size of the deposit (~70 MT). However, the alteration sequences (advanced argillic and alunite) are indicative of acid-sulfate mineralization. Unfortunately, definitive fluid inclusion data is unavailable. Brines from the Casa Diablo geothermal plant can be quite corrosive suggesting locally acidic pHs for the present geothermal system.

The proposed model for the Blue Chert property envisions a fossil geothermal system analogous to the present day system within the Long Valley caldera (Fig. 6). The silica sinter deposits represent the remains of fumaroles or geysers that fed the circulating meteoric waters to the surface. The Hilton Creek fault provided the main structural control for the vents. Gold was complexed as a bisulfide ($\text{Au}(\text{HS})_2^-$).

As the hydrothermal solutions rose along fumarolic vents they were oxidized and the bisulfide complex broke down becoming sulfate (SO_4^{2-}). The gold could no longer remain in solution once the complex broke down and it was deposited. The dissolved silica was also deposited as its solubility dropped exponentially in the now much cooler solution. The sulfate combined with free hydrogen generating H_2SO_4 (sulfuric acid), a powerful acid. The acidic water flowed laterally, and percolated downward into the subsurface. The acid leached the rock leaving only the most insoluble of compounds, kaolinite and alunite (as at the Huntley kaolinite mine). As the water circulated downward it was reheated and reduced. This recharged the aquifer system. The now mineralized and reduced water was recycled convectively upward through the breccia pipe conduit (fumarole). The characteristic elements of this model are a fossil hot springs system; i.e., chert/silica sinter, brecciation, argillic alteration (kaolinite), and alunite. The Blue Chert property was truly a potential “gold mine” waiting to be discovered by anyone utilizing this model.

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Tectonic implications of basaltic volcanism within the central Owens Valley, California

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Abstract: Basaltic volcanism occurs at four locations in the Owens Valley. We examine the Big Pine field south of Independence, CA and the Darwin field situated 45 miles to the southeast on the Darwin Plateau. Big Pine basalts were extruded between 1.2 Ma and 25 ka. They range in composition from *ne* normative alkali basalt to Q normative tholeiitic basalt. Basalts of intermediate composition (olivine tholeiites) are rare. Alignment of cones and flows suggests that both dip-slip and oblique slip faults have acted as conduits for the magma. Olivine and its alteration product, iddingsite, are common in some flows, but entirely absent in others. Trace element and isotopic analyses argue for crustal contamination of some basaltic magma prior to extrusion. Volcanism on the Darwin Plateau occurred from 8 to 4 Ma. Structural control of flows is more obscure. Basalts span a similar range in composition to the Big Pine field, but are characterized by a broader population of olivine tholeiites and not a clustering of alkali and tholeiitic basalts. Olivine is a common modal mineral, as is iddingsite. Evidence for crustal contamination is less persuasive.

Tectonic setting appears to have a significant influence on basalt composition. Basalts extruded in regions of transtension or oblique slip related to dextral shear are dominantly alkali basalts, while those that are the products of pure extension are tholeiites. This suggests the Big Pine volcanic field underwent a change in the pattern of stress from dip-slip (tholeiites) to oblique-slip (alkali basalts) and supports episodic or cyclic partitioning of stress across the Owens Valley. The absence of olivine and iddingsite in many flows may indicate that some fractionation of Big Pine magmas occurred, perhaps in response to a thickened continental crust. Darwin basalts do not fit the established model. The preponderance of olivine tholeiites is difficult to rationalize and trends observed in other fields cannot be documented due to paucity of geochronology and detailed stratigraphic control. Darwin basalts were extruded over a time period that encompasses both Basin and Range extension and dextral shear. Perhaps, the compositional variation of Darwin basalts reflects the evolving tectonic setting.

Introduction

The Big Pine volcanic field is one of the most recognizable features within the central Owens Valley (Fig.1). Basalt flows and scoria cones can be seen along both sides of US 395 from the base of the Sierra Nevada Mountains to the west, to the Inyo Mountains to the east. The field encompasses an area of approximately 150 square miles, from one mile south of the town of Big Pine to 8 miles north of Independence, CA. Volcanism dates from 1.2 Ma to as recently as 25 ka (Manley, et al., 2000; Moore and Dodge, 1980). Basaltic rocks dominate, but a small outcrop area of rhyolite is present to the west of the Poverty Hills.

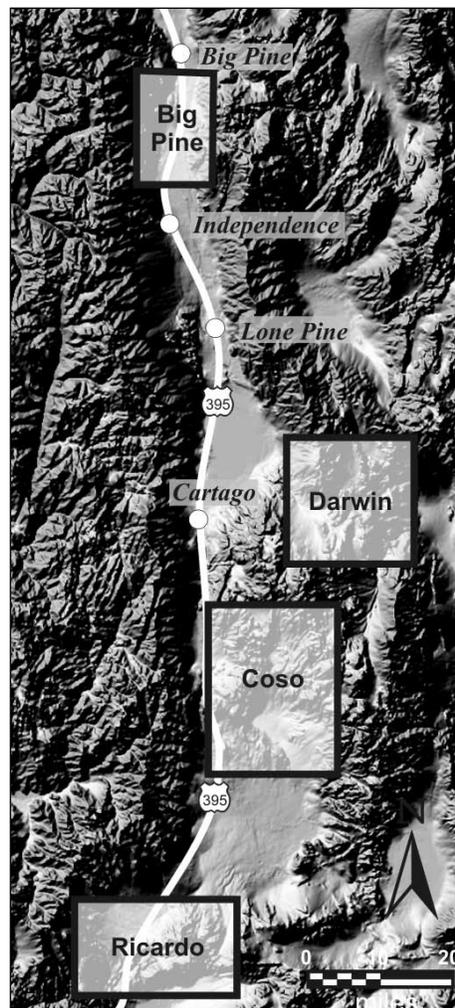


Figure 1. Index map showing the location of basaltic volcanic fields in the Owens Valley.

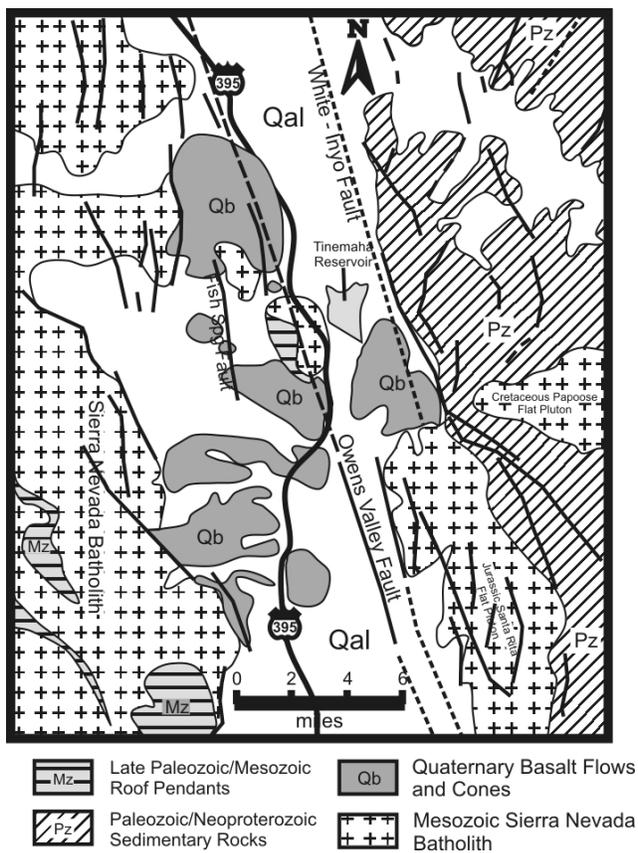


Figure 2. Generalized geologic map of the Big Pine volcanic field (after Varnell, 2006).

The Darwin volcanic field sits atop a horst block known as the Darwin Plateau and lies 45 miles to the southeast of the Big Pine field. It is situated 1000 feet above the Owens Valley to the west and as much as 4000 feet above the Panamint Valley to the east. The field covers an area of 300 square miles, much of it draped over the southern flank of the Inyo Mountains. Limited age dating suggests basaltic volcanism occurred between 8 and 4 Ma (Larsen, 1979; Schweig, 1989). Volcanics of felsic to intermediate composition are absent.

Although the Big Pine and Darwin fields currently lie more than 40 miles apart, this may not have always been the case. The younger Big Pine field straddles the Owens Valley fault, while the Darwin field lies to the east of this fault system. At the onset of Darwin volcanism (~8 Ma), the field may have been situated as much as 35 miles to the north; assuming a rate of motion of 6mm/yr across the Owens Valley fault system. A more conservative rate of 3 mm/yr places the two fields 20 miles closer together at the time of volcanism.

The two fields share many geologic, petrologic and geochemical characteristics. Given the span of volcanism from 25 ka to 8 Ma, an understanding of the

Big Pine and Darwin basalts may provide clues to the Neogene tectonic evolution of the Owens Valley and the Eastern California Shear Zone.

Geologic setting

Cenozoic sedimentary fill in the Owens Valley is quite thick, locally reaching three kilometers (Varnell, 2006). Exposures of pre-Cenozoic basement are unusual, occurring at only two localities; the Poverty Hills within the northern portion of the Big Pine volcanic field and 35 miles to the south in the Alabama Hills. The reasons for bedrock outcrops in the Big Pine field are unclear. Pakiser et al. (1964) suggested the Poverty Hills block had descended along a normal fault to its present location. This would be analogous to a model proposed for the Alabama Hills. Bishop (2000) indicated that Poverty Hills may represent a “long-runout avalanche”. Martel (1989) hypothesized the Poverty hills were the result of transpression caused by a left step in the dextral-shear Owens Valley fault system. Since basaltic volcanics locally overlie the exposed basement complex, all of the various models must assume that much of this fault movement occurred prior to the onset of volcanism at 1.2 Ma.

Figure 2 is a generalized geologic map for the Big Pine field. Much of the area surrounding the basalt outcrops is comprised of large alluvial fans than grade basinward into Quaternary lacustrine and fluvial sediments. The exposed basement, like the surrounding mountains, is primarily Mesozoic granodiorite and quartz monzonite. An isolated roof pendant of Paleozoic metasediments lies just to the west of Tinemaha Reservoir.

Numerous active faults have been mapped within the Big Pine field. The Independence fault lies to the west of the field along the Sierra Nevada front. Le, et.al. (2007) reported that movement along the Independence fault is dip-slip, east side down, at a rate of 0.2-0.3 mm/yr. The Fish Springs fault bisects the volcanic field offsetting several cones and flows. The alignment of flows and cones along the fault suggest it may have acted as an important conduit for ascending magmas. Beanland and Clark (1994) propose that motion along the Fish Springs fault is purely vertical, east side down. Zehfuss et.al. (2001) measured the slip rate at .24 mm/yr. The right-slip Owens Valley fault lies less than a kilometer to the east of the Fish Springs fault. Estimated rate of slip along the Owens Valley fault at this locality is 1.4 mm/yr (Zehfuss, et.al. 2001). Interestingly, the Fish Springs fault and Lone Pine fault, 30 miles to the south, share similar geographic relationships to the Owens Valley fault. The Lone Pine fault was long thought to be characterized largely by vertical slip. It wasn't until the 1960's that a significant strike-slip component was noted, suggesting the Fish Springs fault may also have a

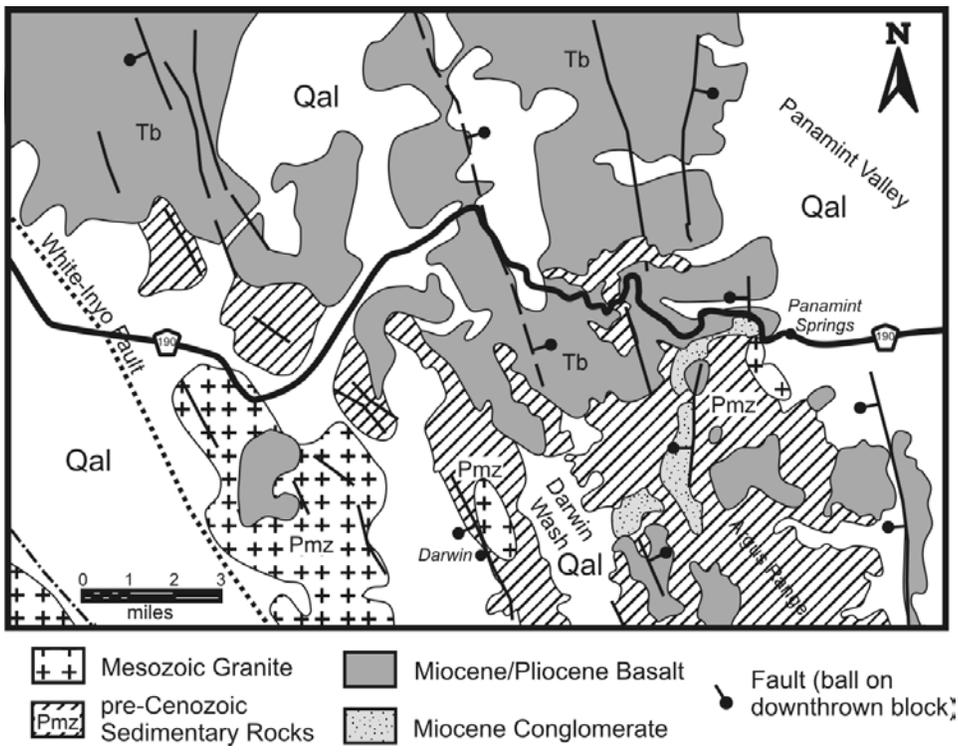


Figure 3. Simplified geologic map of the Darwin volcanic field (after Lusk, 2007).

strike-slip component not heretofore recognized. The White-Inyo fault lies along the east side of the Owens Valley. Bacon, et al. (2005) suggest that Holocene motion along this fault is predominantly right oblique-slip at a rate of 0.1-0.3 mm/yr. However, Bellier and Zoback, (1995) argue that motion along the White Mountains/Inyo Mountains fault system underwent a change from predominantly dip-slip (west side down) to oblique slip at 288 ka. Alignment

of cinder cones along the White-Inyo fault suggests it has acted as an important channel for basaltic lavas.

The Darwin volcanic field is situated on the Darwin Plateau, an elevated block of crust lying between the Owens Valley to the west and Panamint Valley to the east. The northern portion of the field is draped over the southern flank of the Inyo Mountains, while to the east volcanics spill into the Panamint Valley where they become the Nova or Pinto Peak volcanics of Coleman and Walker (1990).

Figure 3 is a generalized geologic map of the Darwin volcanic field. North of CA 190 thick flows of basaltic lava mantle the southern Inyo Mountains. Only scattered outcrops of Late Paleozoic to Triassic sedimentary and metasedimentary rocks are present. South of CA 190 pre-Cenozoic rocks are more common. Folded Paleozoic units have been intruded by post-kinematic Mesozoic granites. The Paleozoic/Mesozoic basement rocks are unconformably overlain by the Miocene-Pliocene Darwin volcanics (8-4 Ma) and contemporaneous

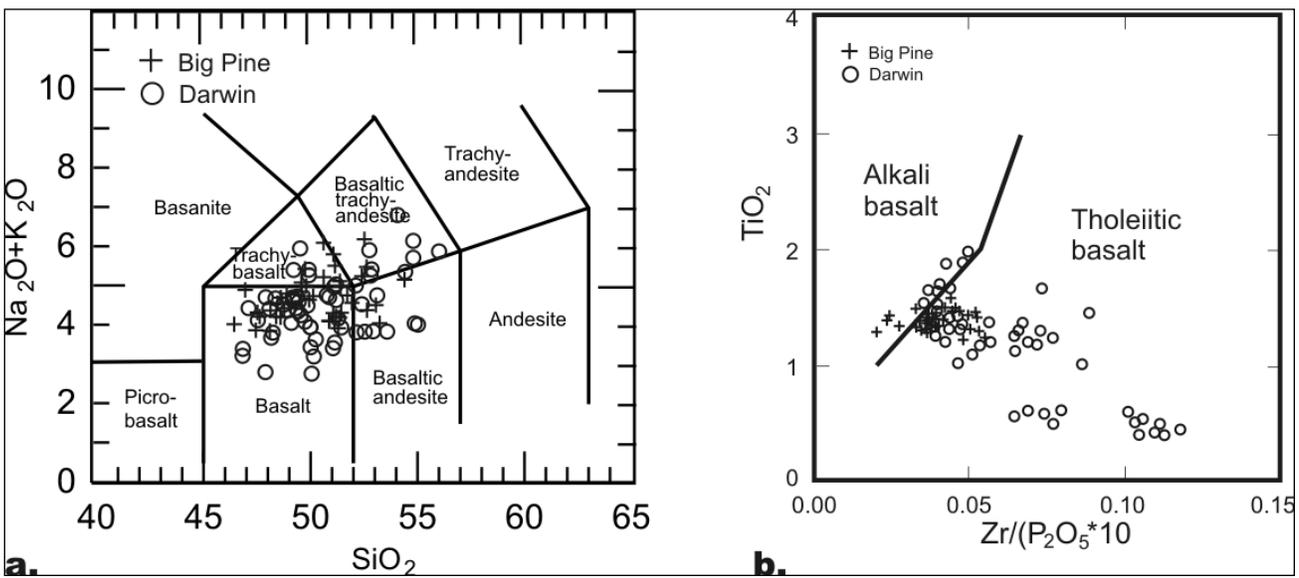
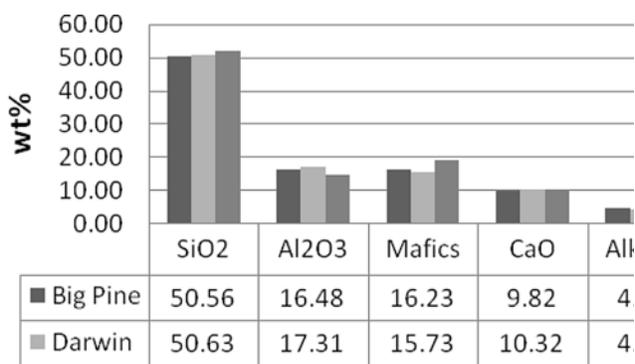


Figure 4 a. Total alkalis vs. silica diagram, and 4b. TiO_2 vs. Zr/P_2O_5 diagram for the Big Pine and Darwin volcanic fields.

Table 1. Average major element concentrations for the basalts of the Big Pine, Darwin and Ricardo (Dove Spring) basalts. Ricardo data from Anderson (2005).



fanglomerates and stream gravels (Schweig, 1989).

Numerous Holocene normal faults cut the flows. The dextral shear Owens Valley fault system and Hunter Mountain/Panamint Valley fault zone lie respectively to the west and east of the Darwin Plateau. The west edge of the field is marked by the projected trace of the White Mountain/Inyo fault. The role these faults have played in basaltic magmatism is uncertain. Schweig (1989) concludes that magma extrusion began shortly after the onset of Basin and Range extension (7-8 Ma). It most likely continued after the initiation of dextral shear at 5 Ma. Thus, it would appear that volcanism in the Darwin field spans the transition from east-west directed extension to north-south dextral shear.

Basalt geochemistry and petrology

Darrow (1972) published the first comprehensive study of Big Pine basalts. He noted that some samples were alkaline and attributed the alkalinity to thickened crust beneath the central Owens Valley. He also noted anomalous strontium content and concluded that the high strontium values were due to crustal contamination, or fractionation of early formed magmas. Gillespie (1982) completed a voluminous study of the northern Owens Valley. While his research focused mainly on glacial evolution, he compiled numerous K-Ar age dates for Big Pine basalts. Bierman et al. (1991) added Ar-Ar dates and the observation that the Big Pine field was characterized by episodic extrusion of both alkaline and siliceous basalts. Stone et al. (1993) published cosmogenic ages of the basalts. Waits (1995) studied peridotite inclusions in basalt and concluded that

crustal contamination seems to be limited to non-xenolithic cones. It should be noted that Waits' thesis has been the subject of some confusion. He sampled only a very limited population of inclusion-bearing flows stating they are highly alkaline. Later researchers, (Taylor, 2002; Dilek and Robinson, 2004) took this statement out of context and characterized the entire Big Pine field as "highly alkaline".

In contrast, the Darwin Plateau has received considerably less attention. Larsen (1979) published a small number of basalt analyses and two K-Ar dates. He characterized the basalts as high alumina to alkaline. Schweig (1989) added three additional age dates and stated the basalts were olivine basalt. The available ages suggest the majority of basalt was extruded between 7.7 and 4.3 Ma. Coleman and Walker (1990) examined all of the volcanic fields around the Panamint Valley. They published analyses of the Nova Basin/Pinto Peak basalts speculating that at one time they were a part of a larger Darwin basalt field that was tectonically dismembered. They also suggested the Darwin basalts had a calc-alkaline affinity.

Table 1 presents major element averages for the Big Pine and Darwin basalts. FeO, MnO and MgO have been combined and reported as mafics; K₂O and Na₂O as alkalis. The Ricardo (Dove Spring) volcanics in the El Paso Mountains are shown for comparison. Big Pine and Darwin share remarkably similar major element geochemistry. The Darwin field has a larger spread for individual analyses, but the overall averages do not differ significantly. Trace element averages (not shown) display somewhat greater variation, but remain quite similar. The chemical similarities of the older Darwin and the younger Big Pine fields suggest

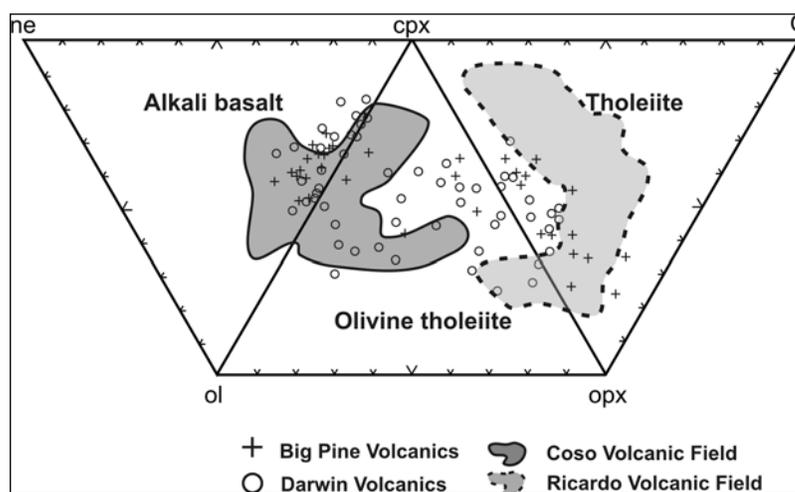


Figure 5. Basalt tetrahedron for the Big Pine and Darwin volcanic fields. Shaded areas represent the limits of compositional variation for the Coso and Ricardo fields. Data from Anderson (2005) for the Ricardo field and Groves (1996) for the Coso volcanics.

that uplift of the Darwin Plateau and subsequent erosion has done little to alter bulk rock chemistry.

Figure 4a is a total alkalis vs. silica (TAS) or LeBas diagram (Le Bas et al., 1986). The majority of samples from both fields classify as basalt or trachybasalt (hawaiite), in agreement with Dilek and Robinson (2004). The TAS diagram, however, is a generic classification diagram not meant to be utilized for the discrimination of basalt types.

Figure 4b employs minor and trace elements to differentiate alkali and tholeiitic basalt (Winchester and Floyd, 1977). The overwhelming majority of samples from both fields lie within the tholeiitic basalt field. (Note: the spread of values is significantly greater for the Darwin samples.) Dilek and Robinson (2004) acknowledge the unusually high zirconium values of Big Pine basalts, but choose not to consider the genetic implications. A high ratio of Zr to P₂O₅ does not unequivocally prove that basalts are tholeiitic. Zirconium is a crustal element and any magma equilibrating with crustal rocks (magma mixing or assimilation) will inherit an anomalous concentration of Zr. Therefore, the Big Pine and Darwin samples may not represent true tholeiites, but rather basalt that have incorporated crustal contaminants.

Figure 5 is a collapsed basalt tetrahedron after Yoder and Tilly (1962). It has distinct advantages over other geochemical diagrams. It relies on normative mineralogy which, although not an exact substitute for modal mineralogy, allows for some degree of comparison between hand sample petrography and geochemical analysis. It also incorporates "intermediate" basalt compositions, e.g. olivine basalt, permitting better recognition of fractionation trends in crystallizing basaltic liquids.

Samples from both fields span the compositional range from alkali basalt to tholeiite. They do not show the restricted range of either the Coso or Ricardo fields, or volcanic fields of the Mojave. Big Pine samples are spread almost equally between alkali basalt and tholeiite with only a small number lying in the olivine tholeiite subtriangle. Ringwood (1976) demonstrated that a thermal divide separated fractionating basaltic liquids and at low pressure it was not possible for an undersaturated *ne* normative magma to evolve into a *Q* normative tholeiite. This implies that the alkali basalts and tholeiites of the Big Pine field must have evolved separately. The Darwin volcanics show a similar range in composition, but the clustering seen for the Big Pine field is absent. The large population of olivine tholeiites would be the "olivine basalts" of Coleman and Walker (1990). The reason for the pre-

ponderance of olivine tholeiites is uncertain.

Darrow (1972) and Varnell (2006) examined Big Pine basalts in thin section. Both noted that olivine was present in some samples and absent in others. Nepheline was an occasional trace constituent. Darrow was the first to note the presence of iddingsite, but unlike its ubiquitous occurrence in the Dove Spring (Ricardo) basalts, it is a rare alteration product in Big Pine basalts. Most flows contain neither olivine, nor iddingsite. Darrow (1972) also noted the presence of peridotite inclusions, later sampled and analyzed by Waits (1995). Lusk (2007) examined the basalts of the Darwin Plateau. Olivine was a more common constituent, as was iddingsite. Nepheline was absent. Furthermore, samples taken from the same flow at different locations showed little variation in either modal mineralogy or normative mineralogy. In contrast, variation between flows was significant. Limited geochronology and the lack of detailed geologic maps made it impossible to recognize temporal trends with any certainty. Lusk (2007), also noted peridotite inclusions similar to those described for the Big Pine field.

Discussion

The basalts of the Big Pine and Darwin fields will be discussed in the larger context of the Owens Valley. Groves (1996) and Anderson (2005) published geochemical, and in the latter case, petrographic studies of the Coso and Dove Spring (Ricardo) volcanics. Analyses from both fields plot within restricted compositional ranges on a basalt tetrahedron (Fig. 5).

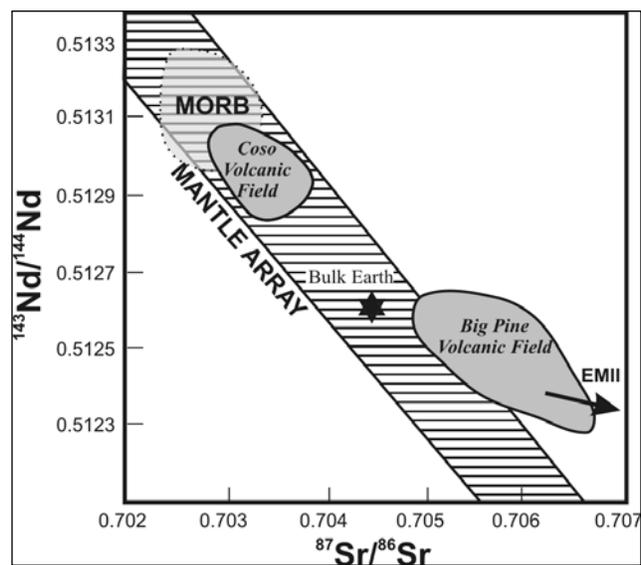


Figure 6. Epsilon neodymium diagram for xenoliths from the Big Pine and Coso volcanic fields. Data from the NAVDAT database.

Kushiro (1968) studied the effect of partial melting as a function of depth (pressure). He demonstrated that shallow partial melts yield tholeiitic magmas, whereas melts from greater depth become progressively more alkaline. Anderson and Jessey (2005) believed, therefore, that compositional difference between the two fields could be explained by a greater melting depth for the Coso magmas. This agreed with Wang et al. (2002) who attributed the increasing alkalinity of basalts along a transect from the Owens Valley to the Colorado Plateau to progressively greater melting depth.

Iddingsite replacement of olivine was first reported from the Big Pine field by Darrow (1972) (for a discussion of iddingsite see Brun, et al., this volume). Anderson (2005) noted that iddingsite was such a common phase in the Ricardo volcanics that only scattered remnants of unreplaced olivine remain. Darrow did not comment on the origin of iddingsite and Anderson felt it was an artifact of weathering. Lusk (2007) also reported the presence of iddingsite in Darwin basalts. He noted textures similar to those first reported by Edwards (1938) and attributed to deuteric alteration of olivine, not weathering. Edwards concluded that alteration occurred when oxygen fugacity increased within the magma chamber, perhaps during magma mixing. Baker and Haggerty (1967) favored circulating, geothermal water as the oxidizing agent. This led Lusk and Jessey (2007) to propose that the differing basalt types from the Coso, Ricardo and Darwin fields did not result from differing depths of melting. The presence of iddingsite required olivine as a precursor and hence the Ricardo/Big Pine/Darwin magmas had to have been as undersaturated as those of the Coso field. This required similar melting depths. Subsequently, the oxygen fugacity of the Ricardo/Big Pine/Darwin magmas increased, while Coso magmas were unaffected.

The reasons for the fugacity change are uncertain. Magma mixing is an unlikely hypothesis. Iddingsite is common in the Darwin field, however, felsic volcanics of equivalent age are unknown. The Coso field has nearly equal percentages of felsic and mafic volcanics, but iddingsite has not been reported. So while there were available reservoirs of felsic magma in the Coso volcanic field there is no evidence of mixing. In contrast, the oxidized Darwin basalts could not have interacted with felsic magmas as felsic volcanics are absent.

Circulating groundwater may also account for differences in basalt composition. This would be a reflection of large-scale fluctuation in groundwater

levels during glacial and interglacial periods. During glacial periods the abundant meltwater would elevate groundwater levels. The circulating groundwater would oxidize the basaltic magmas. In contrast, dry interglacial periods would lower groundwater levels decreasing the possibility of groundwater/magma interaction leaving oxygen fugacity unaffected.

Lusk and Jessey (2007) concluded that while magma oxidation by circulating groundwater would result in instability of olivine and the formation of iddingsite it could not account for the great differences in bulk rock chemistry, especially between the Coso and Ricardo volcanics. The large differences in both chemistry and normative mineralogy must reflect the addition and/or subtraction of major ions. This could occur during weathering, but if weathering is an important factor, it would not explain why the chemistries of the older Darwin basalts are nearly identical to the much younger Big Pine basalts. Nor do the chemical differences reflect the relative mobility of ions during weathering. For example, Ricardo basalts contain 2-3% less Al_2O_3 than Coso or Big Pine basalts. Brown et al. (2008) believed a better explanation for the change in oxygen fugacity and magma composition would involve magma assimilation. Basaltic magma would rise to shallow depth where it would become gravitationally stable. Crustal rocks would be melted and assimilated into the ponded magma. This would increase oxygen fugacity, destabilizing olivine while at the same time changing the composition of the magma.

Evidence in support of this hypothesis comes from isotopes. Figure 6 is an ϵNd diagram for the Coso and Big Pine fields. It is based upon xenoliths in the basalt flows. Coso basalts lie wholly within the mantle array (patterned area), between Bulk Earth (primitive mantle) and MORBs (depleted mantle). Therefore, Coso basalts represent partial melts extracted from the mantle and extruded following limited or no interaction with crustal rocks. Big Pine basalts, however, lie largely outside the mantle array and along the EMII evolutionary trend. Zindler and Hart (1986) propose the EMII reservoir is well above any reasonable indigenous mantle source requiring crustal contamination of mantle-derived magma. We propose this contamination occurred as ponded basaltic magma assimilated crustal rocks.

Monastero et al. (2005) conclude the Coso volcanic field lies at a right step or releasing bend in a dextral shear system that extends from the Indian Wells Valley northward into the Owens Valley. This results in northwest-directed transtension, which is accom-

modated by normal and strike-slip faulting. Nearby, Ridgecrest, CA bills itself as the “Earthquake Capital of the U.S.” due to the numerous seismic events. This implies that faults within the Coso field experience frequent movement and are most likely open conduits for magmas. Therefore, mantle derived magmas would be extruded quickly with little opportunity to interact with the shallow crust.

The tectonic setting of the Dove Spring basalts is less clear. Geochronology indicates they predate the onset of dextral shear and are most likely a product of Basin and Range extension (Loomis and Burbank, 1989). Current rates of movement along the Sierra Nevada Frontal faults (.1-.2 mm/yr) are generally an order of magnitude less than those for right-slip faults (2-4 mm/yr). Assuming these rates can be extrapolated back to the Miocene, pure dip-slip faults would be less readily available conduits than those of the current transtensional strike-slip regime. Rising basaltic magmas would pond until presented with an open conduit. During this ponding the magmas would assimilate country rock changing bulk chemistry and destabilizing olivine. When the dip-slip faults were breached, perhaps by a seismic event, the newly equilibrated magmas would rise to the surface yielding the tholeiites of the Dove Spring Formation.

The Big Pine field has a more complex history. Its tectonic setting is open to question. Martel (1989) believed the Poverty Hills are the result of transpression created by a left-step in the Owens Valley fault. Taylor (2002) states that volcanism is coincident with uplift of the Poverty Hills. This seems unlikely as transpressive stress would not provide the open conduit necessary for basaltic magmas. What is clear, is that the Big Pine basalts are the product of two distinct events, one pulse of tholeiitic basalt and a second of alkali basalt. Although the exact geochronology is uncertain, the tholeiites appear to be generally older. Bellier and Zoback, (1995) argue that motion along the White Mountains/Inyo Mountains fault system underwent a change from dip-slip (west side down) to oblique slip at 288 ka and that additional episodic fluctuations may have occurred in the past. If normal faulting generates tholeiitic basalt and dextral shear alkali basalt, this would explain the compositional differences of the Big Pine basalts. It also adds credence to the hypothesis that extension and dextral shear within the Owens Valley are accommodated largely by periodic changes in the regional stress field. However, this does not explain the relative scarcity of olivine and its alteration product, iddingsite, in many Big Pine basalt flows. Perhaps the thickened crust

of the central Owens Valley impedes magma ascent resulting in fractionation of olivine at depth. The fractionated magma then rises to shallower depth where it assimilates crust acquiring both the trace element and isotopic signature of crustal contamination.

If the genesis of Big Pine basalts is complex, the Darwin basalts present an enigma. They appear to represent a continuous evolutionary sequence and not the separate, discrete pulses seen at Big Pine. In the absence of detailed geochronology or stratigraphic control the direction of this trend is uncertain, although we speculate from limited field observation that tholeiitic basalts have evolved to alkali basalts. Nor is there any obvious mechanism to explain how basaltic magmas can evolve across a thermal divide, or why there is such a preponderance of samples of intermediate (olivine tholeiite) composition. Darwin basalts were extruded over a time period that encompasses both Basin and Range extension and dextral shear. Perhaps the tholeiites represent Basin and Range extension and the alkali basalts transtension, but there is little evidence to support this hypothesis. Finally, any theory must take into account that when the Darwin basalts were emplaced the field was situated at some undetermined distance to the north and, hence, may share a genetic relationship with the Big Pine field.

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The Goler Club: how one man's passion opened a rare window into California's geologic past

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Introduction

The Garlock Fault is a right lateral fault zone that trends northeast-east from the San Andreas Fault Zone in the northern Mojave Desert. Located just north of the Garlock Fault and southwest of Ridgecrest are the El Paso Mountains (study area in Figure 1). The present climate of the El Paso Mountains is extreme. Rainfall averages less than four inches a year, so water is only available during and shortly after rain events or from small springs that occur in a few dry canyons. The El Paso Mountains are about one hundred miles from the Pacific Ocean, with most of the area between 3,000-4,000 feet in elevation. These small mountains experience cold winters and hot summers as the average high temperature of the hottest month (July) is 100 degrees Fahrenheit, while the average low temperature of the coldest month (December) is 30 degrees Fahrenheit. The El Paso Mountains experience a desert climate and thus have a fauna and flora adapted to these extreme variations in environmental conditions.

However, the climate of the El Paso Mountains has not always been what it is today. Paleontological and geological research over the past few decades indicates that 60 million years ago in the Paleocene Epoch, the area was subtropical, at low elevation near the Pacific Ocean. This evidence comes from the Goler Formation which crops out within the El Paso Mountains. The Paleocene Epoch, which directly follows the last records of dinosaurs, was a time of great diversification of plant and animal life, especially mammals. The record of this diversification for the West Coast of North America comes entirely from the Goler Formation as it is the only Paleocene aged rock unit west of central Utah that has yielded a diverse fauna of non-marine vertebrates. Petrified logs two to three feet in diameter now found eroding from Goler sediments indicate that large trees once lined the banks of ancient rivers in this area. Evidence

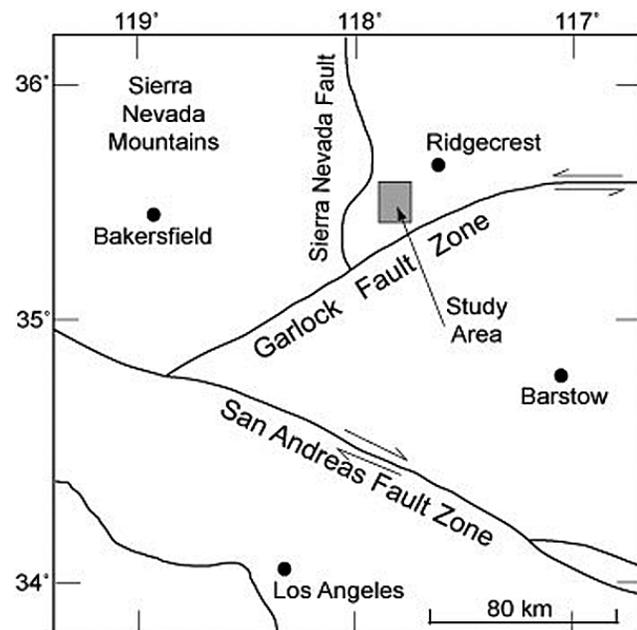


Figure 1. Map showing the location of the El Paso Mountains (shaded) north of the Garlock Fault. Outcrops of the Goler Formation are exposed within these mountains.

of a warm-moist climate comes from fossils leaves which have been identified as belonging to subtropical to warm-temperate plants (Axelrod 1949). Mammals, fish, sharks, crocodilians, lizards, and turtles once lived in or adjacent to the ancient Goler rivers, as their remains are found in these sediments. Mammal groups present include primates and marsupials which have modern relatives, as well as extinct groups such as multituberculates and condylarths (McKenna et al 2008; Lofgren et al. 2008, and references therein). The presence of primates and crocodilians in the Goler Formation indicate a warm-moist climate (Lofgren et al. 2008), in agreement with data from fossil plants.

Significant Goler Formation research efforts span six decades with collecting efforts over the last ten years having dramatically increased the sample of

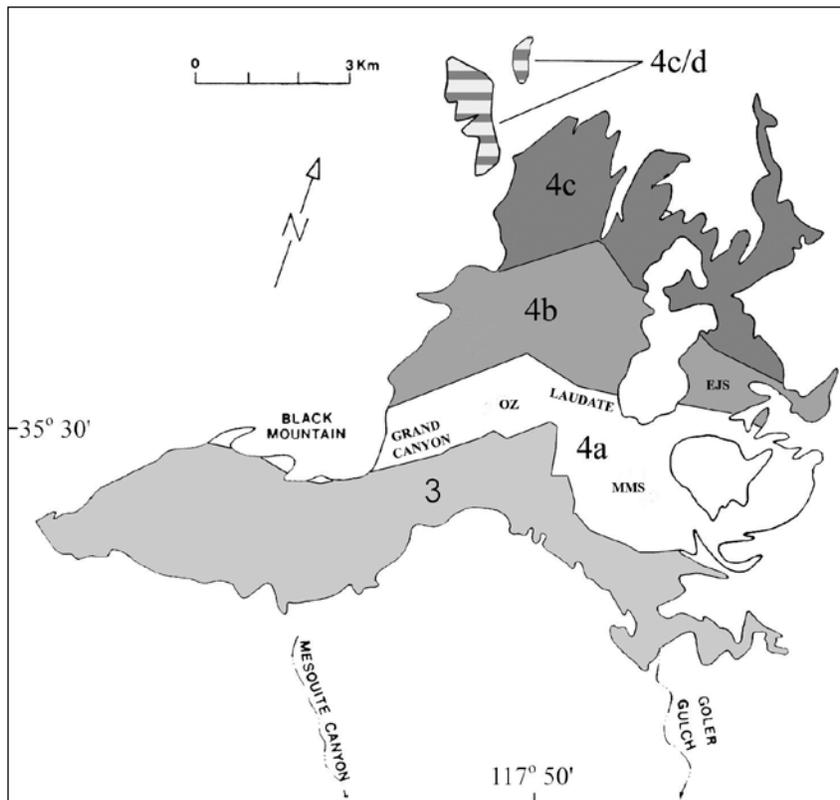


Figure 2. Outcrop map of Member 3 and 4 of the Goler Formation (adapted from Cox, 1982; Cox and Diggle, 1986) showing location of sites or collection areas discussed in text; Member 1 and 2 not shown.

Goler fossils. This recent success has significantly changed our perception of the Paleocene vertebrate fauna of California. Malcolm McKenna, Frick Curator of Fossil Mammals of the American Museum of Natural History and professor of geology at Columbia University, was the driving force behind this success. Unlike sedimentary rocks at many Paleocene sites in the Rocky Mountains, fossils are very rare in the Goler Formation. After dozens of trips to the Goler Formation to collect fossils, usually returning empty handed, McKenna never gave up his goal to see the Goler Formation yield a diverse collection of fossil vertebrates. McKenna had a long and productive career and became widely recognized as one of the most respected mammalian paleontologists of the 20th century. Here was a man with the knowledge and resources who could work anywhere in the world, but perhaps his favorite collecting area was the little known and fossil-poor Goler Formation. For many decades, the age of the Goler Formation was unknown and there was nothing McKenna liked more than trying to solve a geologic mystery. Those that searched with him in his nearly 60 year quest for Goler knowledge were referred to as members of the Goler Club. The ranks of the Goler Club have expanded over the last decade and the number verte-

brate fossils collected has increased tenfold since 1990. This paper is intended to honor the work of Malcolm McKenna and bring the results of Goler Formation research to a broader audience.

Early work and McKenna's introduction to the Goler

The first report of Goler fossils comes from the late 19th century when H. Fairbanks (1896) mentioned that fossil leaf impressions had been recovered from mudstones just above a coal bed in the El Paso Mountains. Fairbanks (1896) cited a personal communication from paleobotanist F. Knowlton that the plant fossils were probably Eocene in age. Early in the 20th Century, J. Merriam (1919) described Miocene mammals found in the Mojave Desert and Great Basin, but at that time fossil vertebrates were not known from the El Paso Mountains. Merriam did send J.

Buwalda to prospect for vertebrates near the area that yielded the Goler plant fossils, but without result. In the 1940s, paleobotanist D. Axelrod visited the site reported by Fairbanks (1896) and collected more plant fossils. Based on these collections, Axelrod (1949) described fossil leaves that indicated a warm-temperate to subtropical climate and interpreted the flora also as Eocene in age. Thus, by the end of the 1940s, the El Paso Mountains were known to have sedimentary rocks with plant fossils but the area was generally considered barren of vertebrate fossils.

Dibblee (1952) mapped the El Paso Mountains and named the 6,500 foot thick section of continental strata prospected by Axelrod, Buwalda, and others, the Goler Formation, which derived its name from Goler Gulch. Based on field relationship, lithologic correlation, and meager fossil evidence, Dibblee (1952) concluded that the Goler Formation was most likely Miocene, but that it probably contained some upper Eocene strata based on the earlier reports of Eocene plant fossils (Fairbanks 1896; Axelrod 1949). Later, Cox (1982; 1987) studied the geology of the Goler Formation and established that the formation is enormously thick, over 9,000 feet, and divided the formation into four members (Figure 2). Cox



Figure 3. Raymond Alf, Priscilla McKenna, and Malcolm McKenna (l-r) relax at a fossil site in Wyoming around 1954, about the time the first mammal specimen was found by M. McKenna in the Goler Formation.

(1982) also determined that the Goler Basin was once traversed by large streams as evidenced by the many thick beds of coarse sandstone and conglomerate in parts of the formation. In these stream deposits are many logs of fossil wood, indicating that trees were once abundant in this area.

Around the time that Dibblee was mapping the El Paso Mountains, vertebrate paleontologist Chester Stock from the California Institute of Technology (CIT) became interested in the Goler Formation. In 1950, Stock encouraged R. Tedford and R. Schultz to prospect in the Goler; subsequently they found a nearly complete turtle shell and a mammal tooth on the southeastern slope of Black Mountain (in Member 3). These specimens were taken to CIT for study. But later in 1950, Stock died and the turtle shell and mammal tooth in his laboratory were discarded when his lab was cleaned out in preparation for use by another professor. In 1948, Malcolm McKenna graduated from Webb School of California, a private high school in Claremont, California, that featured a paleontology program run by science teacher Raymond Alf who inspired McKenna to study paleontology in college. Thus, he enrolled as an undergraduate at CIT to study with Stock. Before Stock died, McKenna had visited Stock's lab and briefly viewed the Goler fossils that were later discarded. Because of Stock's death, McKenna transferred to the University of California at Berkeley to continue his study of paleontology. At Berkeley, McKenna worked on Eocene sites along the Wyoming-Colorado border for his dissertation research project. However, the lost Goler fossils had sparked McKenna's interest and he decided to pros-

pect for fossils in this little known but incredibly thick rock unit in his spare time. Little did he know that his quest for Goler knowledge would continue unabated for the next six decades.

Birth of McKenna's Goler Club

In 1952, R. Tedford and McKenna prospected in the Goler Formation and found a crocodilian tooth in Member 4, higher in the formation from where Tedford and Schultz had found the now lost turtle shell and mammal tooth in 1950. McKenna returned to the Member 4 site with his father Donald in 1954 and they found turtle bone fragments and a mammal jaw with a single tooth (McKenna, 1955). The place

where the jaw was found was named the Laudate Discovery Site (here-after referred to as Laudate; Figure 2).

"Laudate Deum" (praise God) was what McKenna's high school mentor, Raymond Alf (who later founded the Raymond Alf Museum of Paleontology at The Webb Schools), used to shout when an important fossil was found (Figure 3). The Laudate mammal jaw belonged to a member of the condylarth family Peripitychidae; condylarths were small herbivorous mammals that were very common in Paleocene rocks but are now extinct. The jaw represented a genus previously unknown, but its

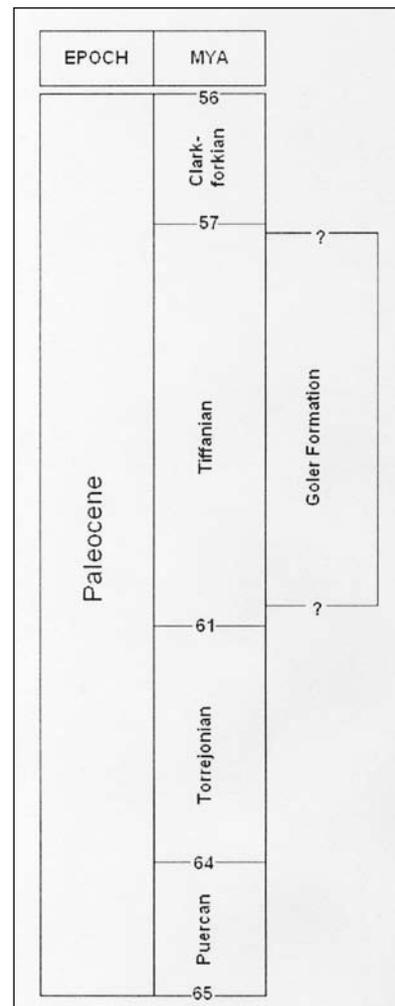


Figure 4. Mammal age subdivision of the Paleocene Epoch showing approximate time span of the Goler Formation.

closest relatives were early Paleocene in age. Because Laudate was higher in the sequence of Goler strata than the Member 3 site with “Eocene” plant fossils, the “Eocene” plants had to be Paleocene or older. This is not surprising as paleontologists have found that, in general, using fossil mammals to date rocks is more accurate than using fossil plants. Thus, in the late 1950s, part of Member 4 of the Goler Formation was regarded as Paleocene, but the ages of the top and bottom of the formation were unknown. Also in 1954, amateur collector F. Corwin found the partial jaws of a mammal somewhere in the upper part of the Goler sequence; the exact location remains unknown. West (1970; 1976) identified the specimen as *Tetraclaenodon puercensis*, a well-known early Paleocene condylarth.

Subsequently, a series attempts to find additional fossils in the Goler Formation were initiated by McKenna in the mid-late 1950s. Paleocene fossils are usually very small so if one is to find them you must often crawl on your hands and knees, scanning the outcrop surface with your eyes about a foot above the ground. With many square miles of Goler outcrop to prospect, knowing exactly which outcrop to crawl to have a good chance to find fossils was intimidating as it would take months to crawl every outcrop. But this did not discourage McKenna. However, after concerted prospecting efforts, success was modest as only four more identifiable mammal teeth were found, all from Member 4 (McKenna 1960). These fossils did corroborate the Paleocene age for Member 4. Also, a tibia and a phalanx of two types of turtles were collected. Interestingly, the tail section of a fish, a type usually found in rocks deposited in oceans was collected in Member 3, suggesting the Goler Basin was once located adjacent to the Pacific Ocean (McKenna 1960).

From 1960 to 1986, McKenna and colleagues continued to scour the Goler Formation, but these efforts yielded few identifiable fossils, all from Member 4. Only two mammal teeth and a few turtle shell fragments were found (McKenna et al. 1987), but these fossils indicated that Member 4 was probably deposited in the Torrejonian age of the Paleocene epoch (Torrejonian is the second oldest of the four mammal ages that comprise the Paleocene; Figure 4). A few land snails were found at a site named Malcolm’s Mollusk Site (Figure 2) but they were too poorly preserved to be identified as to what type of land snail they represented. Interestingly enough, in 1981, a fragment of turtle shell was found that represented a group usually associated with ocean deposits, an ad-

ditional clue that the Goler Basin was near an ocean. Soon thereafter, this was confirmed as marine fossils were found near the top of the formation, in Member 4d (Cox and Edwards 1984), high above the sites with Paleocene mammals. These deposits yielded marine mollusks (clams and snails), foraminifera (tiny shelled marine organisms), and other invertebrates dated as early Eocene (Cox and Diggles 1986; McDougall 1987) or late Paleocene (Cox and Edwards 1984; Reid and Cox 1989) or both (Squires et al. 1988).

By the late 1980s, discovery of new material in the Goler Formation had almost stopped and it was thought the area had been thoroughly prospected. Also, although recognized as the only formation on the West Coast of North America that might yield a diverse assemblage of Paleocene non-marine vertebrates, the Goler developed a reputation of being so sparsely fossiliferous that undertaking a long-term research project was unlikely to yield productive results. With only eight identifiable mammal fossils in hand by 1989, after thirty five years of searching, McKenna joked that those who prospected with him in the Goler Formation were members of the Goler Club, “a very small club of us that actually like the Goler Area!” (from a letter by McKenna to Joe Hartman dated February 1984). Club meetings were held in the field and a member was anyone who searched with McKenna for specimens. McKenna (1960) estimated that it took one person a week of intensive prospecting to locate one identifiable mammal specimen. Up to this point, all vertebrate fossils found in the Goler Formation were deposited in the collections at the University of California Museum of Paleontology.

Alf Museum of Paleontology joins the Goler Club

In 1991, Don Lofgren joined the staff of the Raymond Alf Museum of Paleontology where McKenna served on the Board of Trustees. Lofgren had completed a project on Paleocene mammals from Montana as part of his dissertation research while a graduate student at the University of California-Berkeley and was aware that the Goler Formation had yielded Paleocene mammals. In the winter of 1993, Lofgren decided to take a trip to the Goler Formation to prospect for fossils. Knowing that Alf Museum trustee Malcolm McKenna had published 3 reports describing Goler vertebrate fossils (i.e. McKenna 1955, 1960, McKenna et al. 1987), Lofgren called McKenna and left a message asking for assistance on where fossils



Figure 5. Malcolm McKenna crawls the Laudate site in 1994 in search of *Goler* fossils.

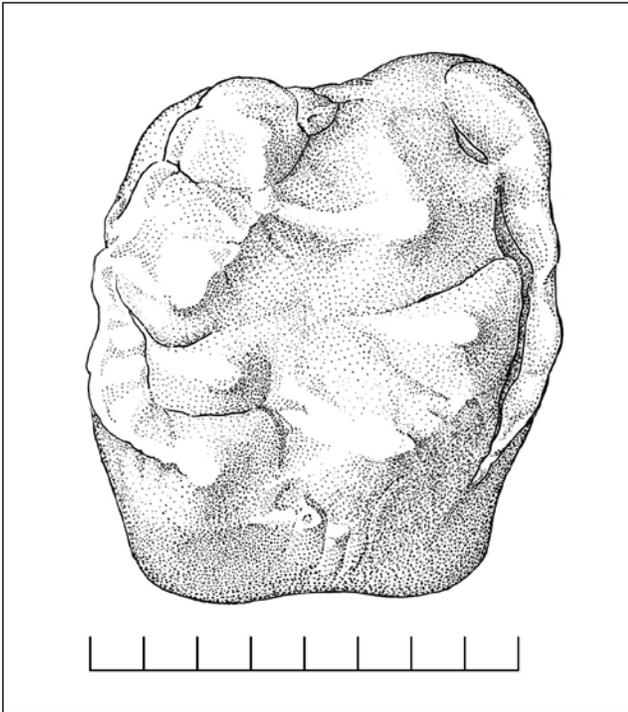


Figure 6. Upper molar of *Phenacodus* (RAM 6722) found by Priscilla McKenna on the day the Oz site was discovered. Scale bar in millimeters.

might be found. Lofgren was unaware of the “Goler Club” and that McKenna was hopelessly obsessed with the Goler Formation. McKenna responded immediately and two weeks later, he took Lofgren on a tour of the Goler Formation. They found a tooth of a peripitychid mammal at Laudate, the same species first found there by McKenna in 1954. Further visits

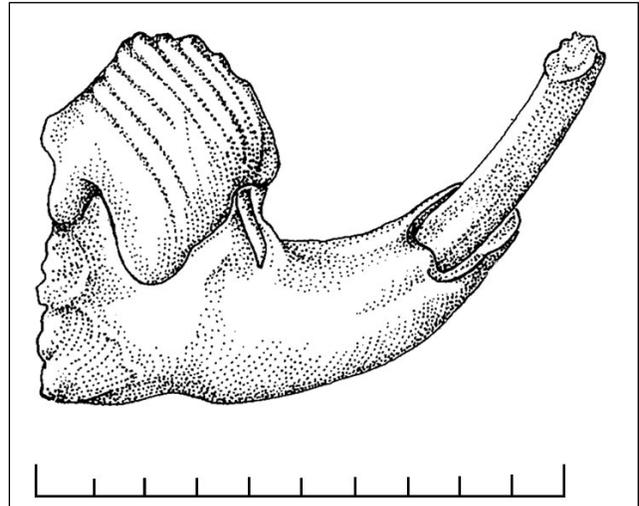


Figure 7. Partial jaw of *Parectypodus* cf. *laytoni* (RAM 7209) found in a concretion by Malcolm McKenna at the Oz site in 2002. Scale bar in millimeters.



Figure 8. Goler Club members on a trip in 2006; (l-r) Randy Nydam, Malcolm McKenna, Jay Greening, Dick Lynas, Don Lofgren, Jim Honey; not pictured, Mike Stokes, Bob Baum, Richard Clark.

in 1994 and 1995 to Laudate (Figure 5) produced a tooth of the multituberculate *Ptilodus* (multituberculates are extinct rodent-like mammals) and a broken molar of the mesonychid *Dissacus*, an odd mammal interpreted to be related to whales. In all, from 1993 to 1995, five identifiable mammal specimens were found which were deposited in the collections at the Raymond Alf Museum of Paleontology. For the Goler, these were good results, but Lofgren was used to finding Paleocene mammals in large numbers in Montana, so he was not hopeful that the Goler Formation was worth pursuing as a major research project. Lofgren became even more discouraged after two unproductive prospecting trips in 1996. But McKenna was not. He continued to invite others to partake in Goler prospecting efforts, most notably Howard Hutchison from the University of California-Berkeley.

The first of two breakthroughs came in spring 1997 when McKenna organized another Goler trip with about twelve participants. On that trip,

Hutchison found a toothless mammal jaw on a mudstone outcrop near where Lofgren had found some turtle shell fragments in 1993. This locality in Member 4b, named the Edentulous Jaw Site (here-after referred to as EJS; Figure 2), looked very promising because the mudstone containing the fossils could be screen-washed. Screen-washing is the most effective method used to separate tiny fossils from rock, as most sediment will pass through the screen mesh, but bones and teeth will not. McKenna contacted Steve Walsh from the San Diego Natural History Museum, a local expert on screen-washing sediment for fossil mammals.

In April of 1998, small screen-washing samples were processed

from both Laudate and EJS by Walsh and each sample yielded a few mammal teeth. Spurred on by this initial success, the scope of screen-washing efforts was expanded. Nine tons of sediment from EJS and three tons from Laudate were processed between 1998 and 2000. This was very labor-intensive work as sediment was first screen-washed in the field, using water from a small local reservoir, and then dried. The dried matrix was transported to Walsh's lab in San Diego

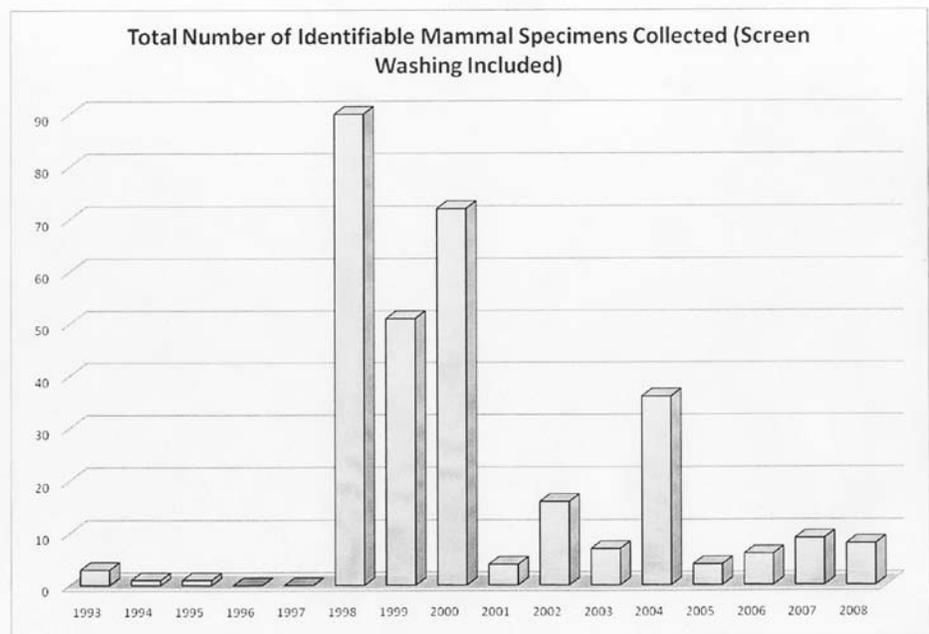


Figure 9. Annual totals for Goler mammals found by surface collection and screen-washing between 1993 and 2008.

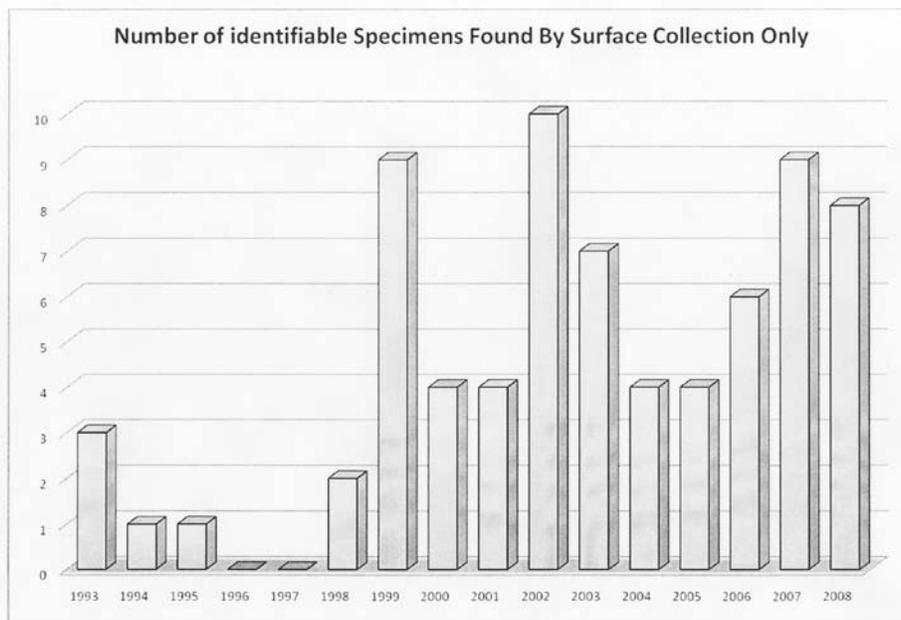


Figure 10. Annual totals for Goler mammals found by surface collection between 1993 and 2008.

for further processing. The matrix was screen-washed again, then soaked in limonene and acetic acid to break up any remaining clumps of sediment, and then screen-washed a third time. The small percentage of remaining matrix was floated in a high-density liquid whose density was adjusted to be between that of the fossils and the remaining sediment. The fossils and heavy minerals sank while the rest of the sediment floated and was discarded. The matrix, now highly concentrated with fossils, was delivered to the Alf Museum where students at The Webb Schools scanned the matrix through microscopes and removed any bones or teeth. Each specimen was then catalogued and studied.

About two hundred tiny mammal teeth were recovered from EJS and these fossils were exciting because of the diversity of mammals represented, notably California's oldest known primates, carnivores, ungulates, marsupials, and multituberculates, as well as lizards and crocodylians (Lofgren et al. 1999; 2002; Nydam and Lofgren 2008). Primates and marsupials were known from Eocene rocks in southern California, but those from

the Goler were much older and included the primates *Plesiadapis* and *Ignacius* and the marsupial *Peradectes*. The results of screen-washing from Laudate were also good. The tiny primate *Paromomys* was found, as well as a few teeth of multituberculates. Thus, from screen-washing the two main sites, Laudate and EJS, the number of specimens of identifiable mammals increased from less than twenty (the yield between 1954 and 1996) to over two hundred, a relatively miraculous development.

The second breakthrough came in November 1999.

The day after attending a memorial service for Alf Museum founder Raymond Alf, Malcolm McKenna and his wife Priscilla, Don Lofgren, Bob Baum, and Dick Lynas headed to the El Paso Mountains to prospect. The spirit of Ray Alf perhaps was with them on this trip, because Lofgren found an upper jaw of a new species of condylarth, *Mimotricentes tedfordi* (McKenna and Lofgren 2003) at Laudate. It was the first mammalian upper jaw ever found in the Goler Formation. The next morning, Bob Baum found a new site, lower in Member 4a, about two miles southwest of Laudate. While



Figure 11. The present day Goler Club; (l-r) R. Nydam, D. Lofgren, B. Baum, R. Clark, M. Stokes, D. Lynas, J. Greening; not pictured J. Honey.

prospecting, Baum wandered over to a flat area where there were hundreds of concretions (very hard nodules that form within sedimentary rock after the rock has been deposited and deeply buried) lying on the ground that had eroded from the underlying sedimentary rock. It is well known that concretions at some sites in North America are the source of fossils. Crawling across the litter of concretions, Baum found a mammal tooth and returned to camp to show the others. Excitement was high after the discovery and the whole group returned to the site to prospect further. After about an hour of walking the flat outcrop and finding nothing, the group gathered together to discuss what to do. To everyone's surprise, Malcolm McKenna looked down and picked up a mammal jaw lying next to his feet. Everyone was down on their knees immediately, and soon, a jaw and two isolated teeth were discovered, one still partially encased in a concretion. Also, an upper tooth of the mammal *Phenacodus* (Figure 6) was found by Priscilla McKenna, a specimen that would eventually provide important information about the age of Member 4a of the Goler Formation. This new site was named the "Land of Oz" (here-after referred to as Oz; Figure 2) in Baum's honor (Baum is the great grandson of L. Frank Baum who wrote *The Wizard of Oz*). In one visit the site had produced five mammalian specimens, representing two types of condylarths, including a *Phenacodus* upper molar.

What that first trip to the Oz site showed the Goler Club is that concretions were the source of many of the Oz specimens. This suggested that the many concretions present in other areas of Goler might also contain fossils. But for now, over the next two years, Goler prospecting trips were mainly concentrated on Oz and many excellent specimens were found, most in concretions. To remove fossils from concretions is very laborious, as the rock is harder than fossil teeth and bones. An experienced fossil preparator, Michael Stokes, was hired to prepare the Goler concretions and McKenna funded this work. It was well worth the expense, as for example, Stokes prepared one concretion that contained the skull with jaws of the multituberculate *Neoplagiaulax* cf. *hazeni* and another that contained the jaw of *Parectypodus* cf. *laytoni* (Figure 7). When McKenna found both of these concretions, just a little bone was evident. But after a hundred or more hours of work by Stokes, where he carefully scraped the rock away, these exceptional multituberculate specimens were revealed.

During this time, visits to Laudate in 2000 and 2001 continued to yield important specimens, as a

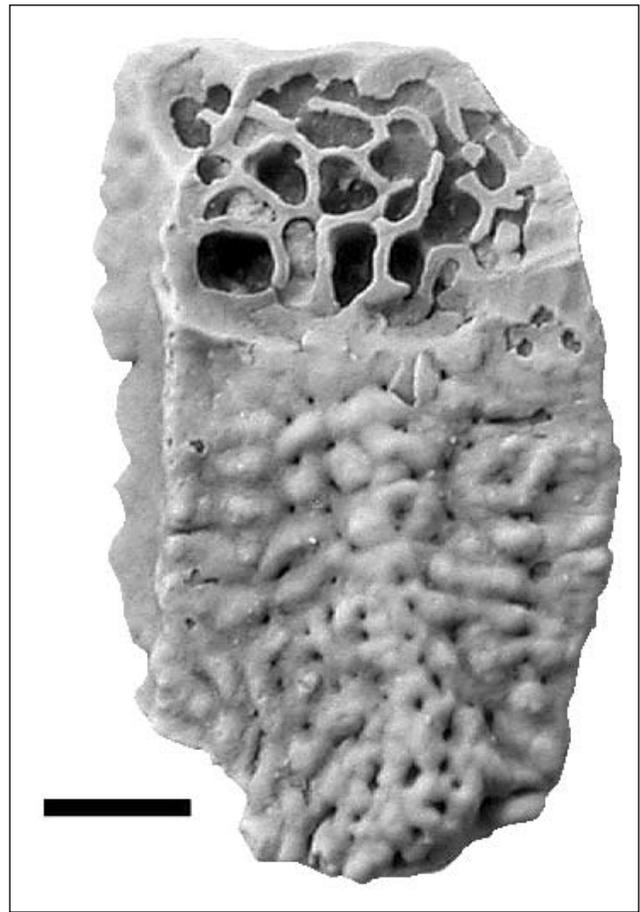


Figure 12. Lizard osteoderm (RAM 10021) from EJS. Scale bar equals one millimeter

tooth of the primate *Plesiadapis* was found, as was a tooth of *Phenacodus*. (Lofgren et al. 2002; Lofgren et al. 2008). In 2002 a jaw fragment of a second species of *Plesiadapis* was found at a new site in Member 4b, three hundred yards north of EJS (Lofgren et al. 2008). Then in 2003, a *Phenacodus* tooth, three shark teeth and numerous well preserved oysters and other mollusks were found in the marine unit of Member 4d near the top of the Goler Formation (Lofgren et al. 2004a). Positive momentum was building for the Goler Club. In contrast to early 1990s, now Goler Club members fully expected to find at least two excellent specimens per trip and often found more.

The new confidence was mostly based on the realization that concretions are a major source of mammal fossils. This meant a change in focus to prospecting outcrops where concretions littered the ground, knowing fossils must be present in some of them. Also, the addition of James Honey and Randy Nydam to the Goler Club (Figure 8) was critical as the two were very experienced collectors who rarely came back from a day of Goler prospecting empty-handed. On his first trip in 2001, Honey found a concretion

with a jaw enclosed that later became the type specimen of the new genus and species, *Goleroconus alfi* (McKenna et al. 2008). Nydam found many important specimens, and discovered the first mammal jaw in a new and very productive area called Grand Canyon (Figure 2).

Between 2004 and 2008, many sites were found in Member 4a; a stunning success. Four new sites yielding mammal teeth and jaws were found south of Malcolm's Mollusk Site. Also, the prospecting area called Grand Canyon (Figure 2) produced five new sites, including three nearly complete primate jaws. Most specimens from these new sites were found encased in concretions. There are now about four hundred cataloged mammal specimens from the Goler Formation in the collections at the Raymond Alf Museum of Paleontology; note that less than a dozen identifiable mammal specimens were collected between 1950 and 1990. On reason for this sharp increase in the number of identifiable mammal specimens was the success of screen-washing efforts (Figure 9; after a three year hiatus, another EJS screen-wash sample was processed in 2004). Also, the discovery of the Oz site combined with a systematic searching of outcrops laden with concretions increased the yield of surface collections (Figure 10). All this success brought everyone involved great joy, but there was great sadness as well, because the leader of the Goler Club, Malcolm McKenna, passed away in March 2008. Although current Goler Club members (Figure 11) will enthusiastically carry forth the research project that McKenna started so long ago, they will greatly miss their founder and inspirational leader.

Significance and age of Goler Formation fossils

The significance of the Goler Formation lies in its uniqueness; it is the only Paleocene-aged rock unit on the West Coast of North America that has yielded a diverse collection of fossil mammals and other vertebrates. Also, specimens of primates, marsupials, multituberculates, condylarths, and other mammalian groups from the Goler Formation constitute the oldest dental records of Mammalia from California; the nearest sites of comparable age are located in Wyoming and Colorado. Thus, the Goler Formation provides a very important window of what coastal California was like 60 million years ago. Based on the presence of petrified logs, fossils of crocodylians, primates, sub-tropical plants, and marine fossils, it is known that the El Paso Mountains were once close to,

and then part of, the Pacific Ocean and experienced a warm, wet climate. The Goler vertebrate fauna in many ways is similar to mid-late Paleocene faunas from the Rocky Mountain States. But even at this early stage of study of the Goler collections, a number of new species are present (McKenna and Lofgren 2003; Hutchison 2004; Lofgren et al. 2008; McKenna et al. 2008), evidence that the coastal Paleocene fauna differed from inland faunas to a significant degree.

Lizard remains from the Goler Formation were only found by screen-washing. Tiny lizard scales or osteoderms (Figure 12) were the most common type of vertebrate fossil recovered from screen-wash samples. Preliminary identification of the Goler lizard fossils indicates that three types were present, angnids, xantusiids, and scincomorphans (Nydam and Lofgren 2008). Angnids are alligator lizards, scincomorphans are relatives of skinks (leaf litter dwellers), and xantusiids are relatives of night lizards (rock and tree bark crevice dwellers). The Goler specimens represent the western most Paleocene-aged lizard remains in North America.

The age of the Goler Formation is now better known based on recently collected fossils. The mammals from Member 4a and 4b provide the best method for dating that part of the Goler Formation. In the 20th century, mammalian paleontologists from North America developed a series of mammal ages that subdivided the Paleocene Epoch. These units were based on assemblages of fossil mammals, each interpreted to denote a short interval of geologic time. Paleocene mammal ages are, in ascending chronologic order, Puercan, Torrejonian, Tiffanian, and Clarkforkian (Figure 4). In 1954, a set of lower jaws were found in Member 4 (probably 4b) and were described by West (1970, 1976) as *Tetraclaenodon*, a taxon that indicated a Torrejonian age for that part of the Goler Formation. However, in recent years, thirty specimens of the closely related taxon, *Phenacodus*, have been recovered by the Goler Club from Members 4a and 4b. Lower teeth of *Phenacodus* and *Tetraclaenodon* can be extremely difficult to distinguish, but upper teeth have a clear difference in the development of cusps used in chewing food; a well-developed mesostyle cusp on the outer edge of the upper molars is present in *Phenacodus*, but not in *Tetraclaenodon* (Thewissen 1990). Upper teeth of *Phenacodus* in the Alf Museum's collections from Member 4a and 4b have well-developed mesostyle cusps, strongly suggesting that the specimen once identified as *Tetraclaenodon* (West 1970) is *Phenacodus*. *Phenacodus* is common in Tiffanian aged rocks but is not known from rocks of Tor-

rejonian age (Thewissen 1990; Lofgren et al. 2004b). Also, there are now thirty primate specimens in the Alf Museum collections from Member 4a and 4b. Most of these specimens, including three nearly complete jaws, represent the genus *Plesiadapis*. *Plesiadapis* is another taxon not found in Torrejonian rocks but is common in Tiffanian ones (Gingerich 1976; Lofgren et al. 2004b). These specimens and others indicate that Member 4a and 4b of the Goler Formation are Tiffanian in age, about three to five million years younger than interpretations made many years ago based on very few fossils (e.g. McKenna 1955; 1960; West 1970).

A study that further refines the age of Member 4 of the Goler Formation will soon be published (Albright et al. in press), but there is still much work to do. About 25 years ago, marine fossils were first reported from Member 4d (Cox and Edwards 1984; Cox and Diggles 1986), which were dated as early Eocene (McDougall 1987) or late Paleocene (Reid and Cox 1989) or both (Squires et al. 1988). The one Alf Museum mammal tooth from Member 4d represents a species of *Phenacodus* well known from Late Paleocene rocks from the Rocky Mountain States (Thewissen 1990). Member 4d may indeed be late Paleocene but more data is needed to further refine its age. In 1950, Tedford and Schultz found a mammal tooth in Member 3. If that tooth had not been discarded, it might have helped constrain the age of Member 3. Unfortunately, after decades of prospecting, no other mammalian fossils have been found in Member 3 and its age remains uncertain. Also, Member 1, Member 2, and Member 4c have not yielded fossils of any type. Thus, the ages of these parts of the Goler Formation are also unclear. Continued exploration and collecting efforts by the Goler Club will surely expand the Goler fauna list for Member 4 and also hopefully provide new information that could be used to determine the age of Members 1, 2, and 3.

In closing, the Goler Formation is a unique paleontological resource as it provides an important window into California's geologic past. Its fossil resources should be managed in consultation with paleontologists from museums who have worked in the area and are the custodians of Goler specimens collected thus far (vertebrate fossils from public lands collected under permit remain the property of the US Government but are generally considered to be on permanent loan to the institutions that collected the specimens). Goler research will continue to be a high priority project for staff at the Alf Museum of Paleontology as well as other members of the Goler Club.

Acknowledgements

This paper is dedicated in memory of Malcolm McKenna who funded the Alf Museum's Goler research efforts from 1993 to 2008 and whose enthusiasm and passion for Goler research never wavered over six decades. Blessed with a lively and dry sense of humor, Malcolm considered the eventual success of his Goler project "the world's foremost example of deferred gratification!" We thank: Bob Baum, Dick Lynas, Richard Clark, Jay Greening, Mike Stokes, Bob Reynolds, Priscilla McKenna, Howard Hutchison, Mike Woodburne, the late Steve Walsh, students of The Webb Schools, and many others for help in Goler activities; David Lawler, Donald Storm, Judyth Reed and Russell Kaldenburg of the Bureau of Land Management for their assistance in the permitting process; National Geographic Society (grant #6736-00), the Mary Stuart Rogers Foundation, and the Goler Research Fund of the Raymond Alf Museum of Paleontology for financial support.

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Fish Creek rhythmites and aperiodic temporal sedimentation

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Abstract: We present an analysis of rhythmites in the Pliocene Deguynos Formation of Fish Creek Wash, Anza Borrego Desert State Park, CA. These rhythmites occur as repeating triplets or quadruplets of silty arkosic sandstone and claystone. Measurements of the rhythmite spatial intervals and simulation based on Gaussian statistics reveals that the Fish Creek rhythmites display a narrowband spatial periodicity with a mean interval of about 19 cm and a standard deviation of about 0.18 cm. Rhythmites are widely thought to have been produced by temporally-periodic deposition. It is also possible that aperiodic processes can produce rhythmic sedimentation. We suggest two conceptual mechanisms by which this can happen: aperiodic limited-velocity overflow of normally-closed basins, and threshold angle-of-repose sedimentation in granular flow.

1. Introduction

Rhythmites (Figure 1) are spatially periodic, graded, sediment beds or lithified beds in which two (couplet) or more visually contrasting layered sequences repeat many times with nearly the same thickness (e.g. Reading 1996). Most are formed in the ocean near continental shelves or in lakes and rivers. The term *rhythmite* implies no limit as to thickness or complexity of bedding and it carries no time or seasonal connotation (Jackson and Bates 1997). Most rhythmites consist of couplets of upward-fining, alternating light and dark sediments. The light-colored layers are flood deposits and are usually thicker than the dark, intraflood layers. Couplet components can be clastic,

organic or biogenic and in lacustrine environments can be mediated by biologic activity. Many rhythmites are turbidites, showing predictable changes in bedding from coarse layers at the bottom to finer laminations at the top, known as Bouma sequences (Bouma 1962). Particle size gradients results when larger or denser particles settle faster and sooner than smaller particles.

Rhythmites produced by annual forcing are called varves. The term is also used to describe semidiurnal, daily and bi-monthly tidal deposits that are mediated by the Moon. In broad geologic usage, a *varve* is any regularly banded sedimentary rock whose depositional time intervals were truly periodic, usually a year or less.

Although most rhythmites are probably deposited in a temporally periodic manner, the depositional driving forces need not be temporally periodic. Indeed, there is evidence that some rhythmites form sporadically (Lambert and Hsu 1979). To our knowledge, however, no one has ever clearly set forth a mechanism for aperiodic rhythmite production. In this paper we (1) analyze the composition and particle size distribution of the Fish Creek rhythmites, (2) present a spatial analysis of them and (3) suggest a class of mechanisms by which tempo-

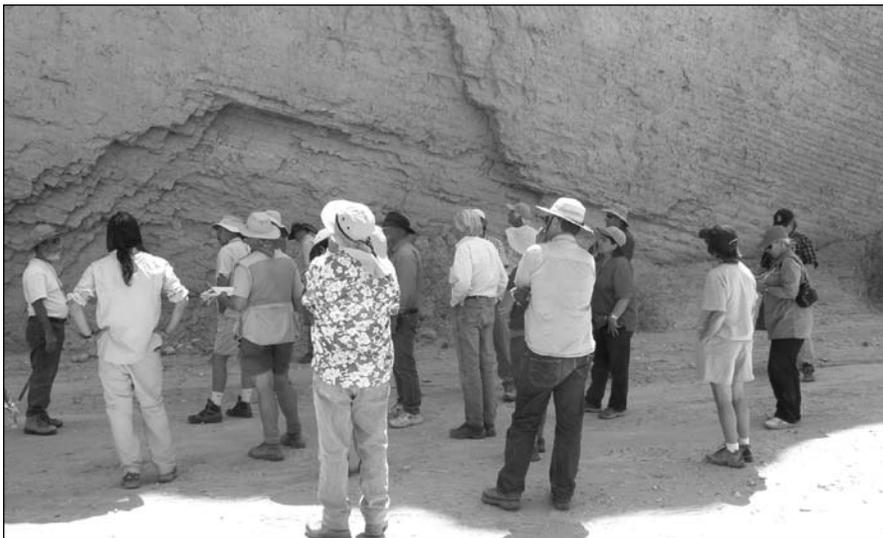


Figure 1. Rhythmites in the Pliocene Deguynos Formation of Fish Creek Wash, San Diego County, CA. 32 58 48N 116 08 45W, California, San Bernardino Meridian T14S,R8E, sec4 UTM zone 11 (X,Y) 579826, 3649339).

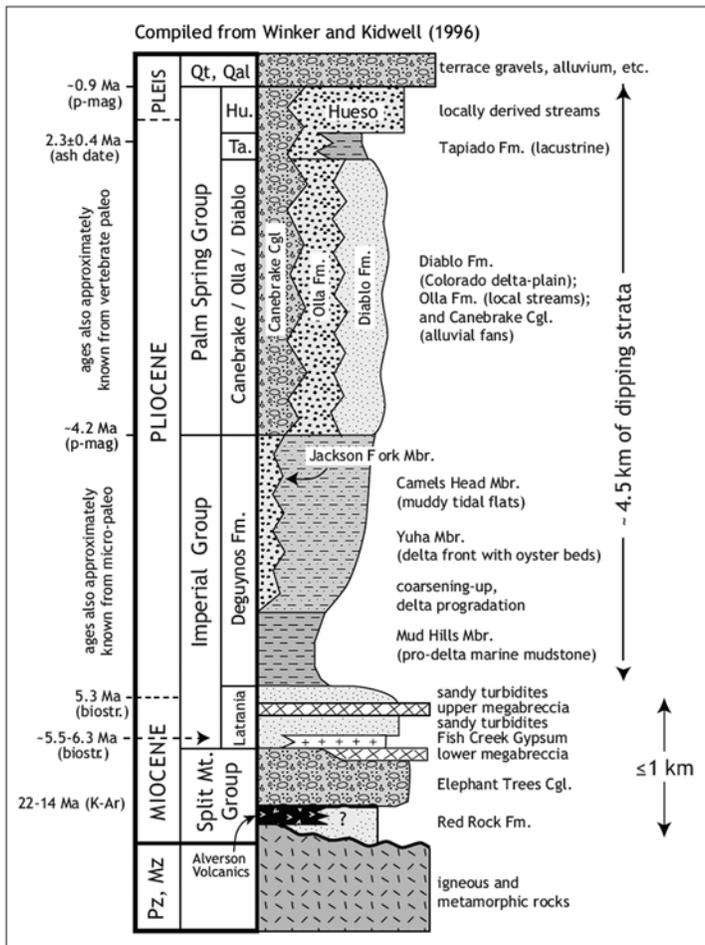


Figure 2. Winker and Kidder's (1996) generalized stratigraphic column of the Vallecito Creek-Fish Creek stratigraphic section, adapted from Dorsey (2006). from Dorsey (2006).

rally random processes can result in spatially periodic sedimentation. We do not know the deposition time intervals of Fish Creek rhythmites nor are we suggesting that their depositional intervals were aperiodic.

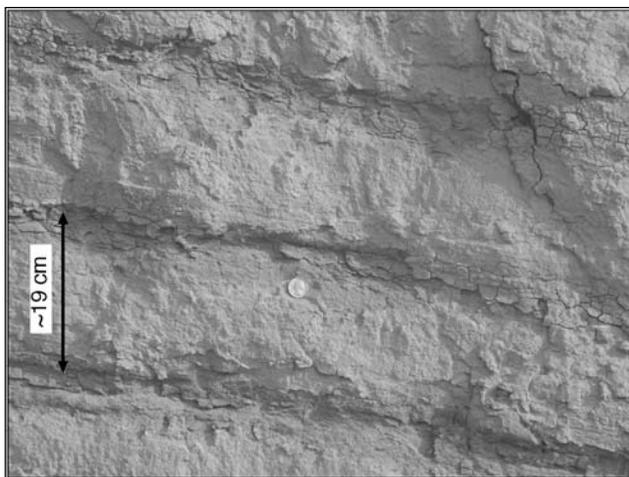


Figure 3. Fish Creek rhythmites consist of repeating triplets or quadruplets. A broad, tan layer with harder, darker brown laminae above and below bound the thin, soft gray layer.

2. Fish Creek Rythmites

Along Fish Creek Canyon, Anza Borrego Desert State Park, San Diego County, CA, the rhythmites are exposed in a near-vertical cliff that extends approximately 20 meters above grade. According to Reynolds, Jefferson and Lynch (2008), "These primarily yellowish and medium brown cyclic silty sandstones occur in the top of the Coyote Mountain Clays or Mud Hills member of the Deguyunos Formation. They are ancestral Colorado River delta front sediments, about 4.8 to 4.5 Ma, and crop out for 1.5 km along Fish Creek Canyon. The physical drivers for this depositional phenomenon (climatic, tidal or other) and the duration of a single cycle are presently undetermined." Figure 2 shows a generalized stratigraphic column of the Vallecito Creek-Fish Creek stratigraphic section from Winker and Kidwell (1996), adapted from Dorsey (2006).

Figure 3 shows a small section of the Fish Creek rhythmite. In most places it consisted of a repeating triplet, though in some places it appeared as a quadruplet (Figure 4). Within a few tens of meters in any direction the mean spatial interval varied considerably, and there was visual evidence that the higher rhythmites had smaller spatial periods than the lower ones. The vertical cliff face showed a scalloped pattern (Figure 4) where the soft gray layers were more deeply incised than the harder tan and brown members.

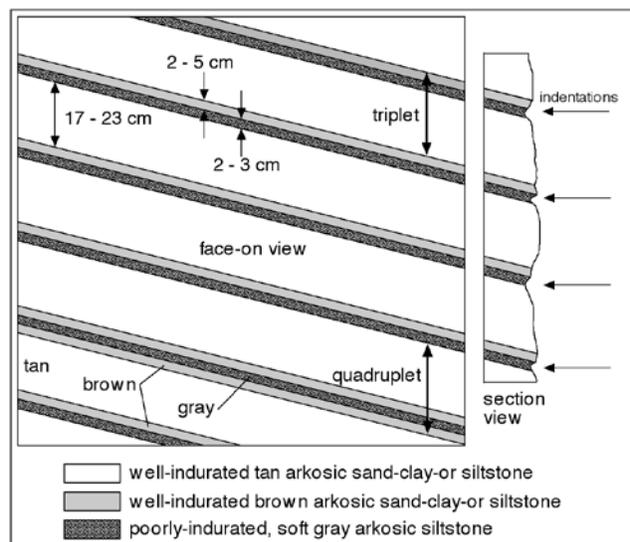


Figure 4. Description of the exposed Fish Creek rhythmite.

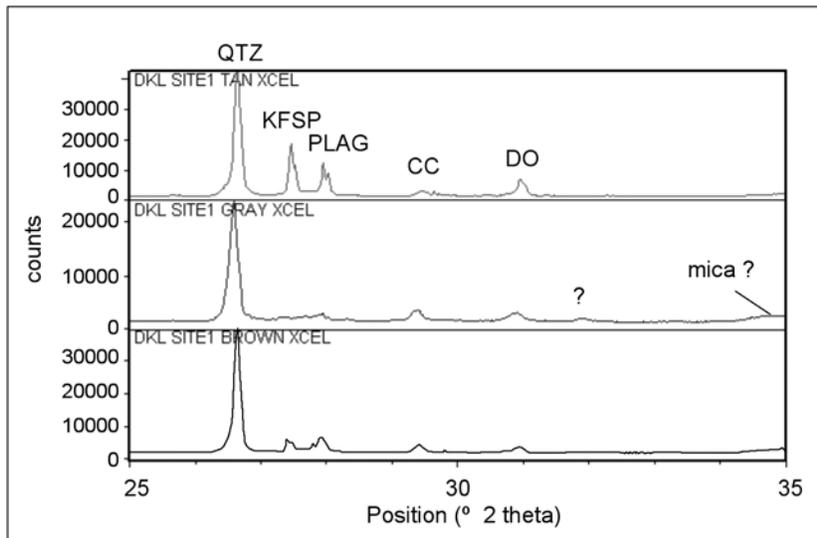


Figure 5. X-Ray Diffraction (XRD) of the three components - tan (upper), gray (middle), and brown (lower). Quartz (QTZ) is the dominant component, followed by feldspar (KFSP & PLAG) with minor amounts of calcite (CC), dolomite (DO) and mica.

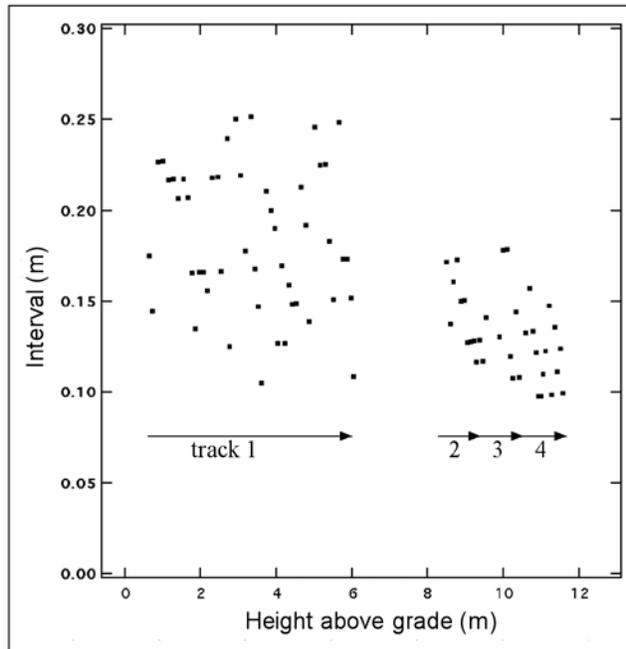


Figure 6. Rhythmite interval vs height above grade along four tracks. Note that the intervals fall into a fairly narrow range, the higher ones being smaller than the lower ones.

We analyzed each of the three rhythmite components using SEM imagery of dispersed samples and X-Ray diffraction (XRD). Based on SEM imagery, we found that the gray particles were mostly angular, <20 μm in diameter and had relatively few smaller particles. The tan and brown material had slightly smaller sizes (<10 μm) but many more smaller particles than the gray did. With no particles larger than about 20 μm , the rocks appear to be siltstone and claystone. XRD revealed compositions and structures of quartz (QTZ), K-feldspar (KFSP), plagioclase (PLAG), calcite (CC), Dolomite (DO), mica and clay (Figure

Table 1

	mean (μ) (m)	std. dev (σ) (m)	β (σ/μ)
track 1	0.192	0.0398	0.207
track 2	0.141	0.0191	0.135
track 3	0.138	0.0303	0.220
track 4	0.112	0.0192	0.173

5). The assemblages might be termed arkosic claystone, fine-grained arkosic sandstone or silty sandstone.

3. Statistical Analysis of the Fish Creek Rhythmites.

Using a laser rangefinder, inclinometer and digital camera mounted on a tripod, we obtained calibrated digital photographs of the cliff face.

These were digitally stitched together and then rotated by 10 degrees so that the rhythmite planes were horizontal. The rhythmite intervals were

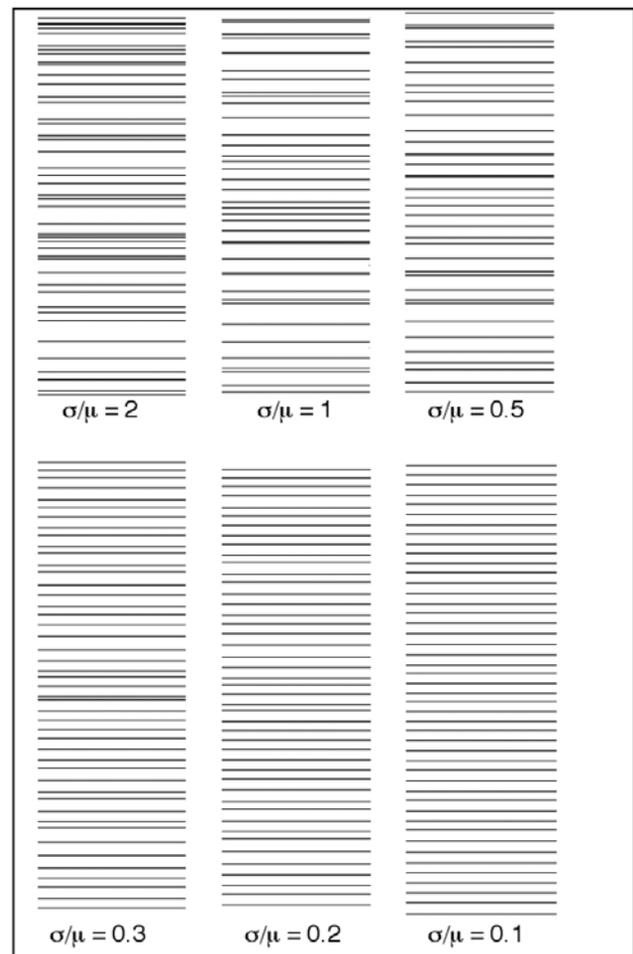


Figure 7. Rhythmite simulations using randomly selected intervals from a Gaussian distribution. The spatial bandwidth $\beta = \sigma/\mu$ is shown below each simulation.

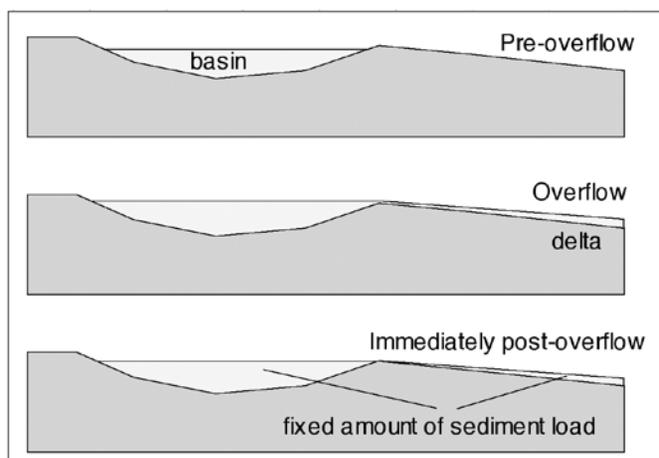


Figure 8. One way to make rhythmites with aperiodic deposition events: a normally-undrained basin (top) occasionally over flows (middle) creating a delta. When the flow stops (bottom) there is a fixed amount of water and sediment flowing onto the delta. Providing that no deposition has taken place before the flow from the basin stops, the fixed amount of sediment downstream from the basin will result in a fixed thickness of deposited sediment. Note also that the amount of water in the basin (and presumably the suspended sediment load) remains fixed throughout.

measured from the photographs along four different vertical tracks and corrected for foreshortening with angle above horizontal. Four tracks (rather than a single one from cliff base to top) were necessary because in many locations a loose layer of dried, tan mud obscured the rhythmite, making measurements impossible. We did not know the true dip of the sediments, so our measured intervals represent upper limits.

The intervals were not strictly uniform but had a relatively narrow range of thicknesses (Figure 6). For each track the mean μ and standard deviation σ were computed (Table 1). As we will see in analyzing rhythmite properties, the value σ/μ is important in characterizing the sediment's spatial bandwidth in the context of normal (Gaussian) distributions. For the Fish Creek rhythmites that we measured, $\sigma/\mu = 0.18 \pm 0.038$.

There is a clear trend towards small intervals higher in the cliff. This is evident in both Figure 6 and table 1. Indeed, as noted earlier, there is considerable variation in spatial period over distances of tens of meters. The bandwidth, however, remains relatively small throughout the formation.

A potentially useful spatial metric for characterizing rhythmites and varves is the quantity $\beta = \sigma/\mu$, which we call the *spatial bandwidth*. Clearly, a small value of β - much less than unity - would be considered "narrowband" and indicate relatively regular spacing. A larger value would be considered "broad-

band" and would indicate significant spacing variation. The significance of β will be developed in the next section.

4. Modeling Spatial Periodicity.

We previously mentioned that $\sigma/\mu = \beta$ could be viewed as a "bandwidth" and used to characterize the statistical properties of the rhythmite intervals. In this section we will put this significance of $\beta = \sigma/\mu$ on a firmer mathematical footing by simulating couplets based on Gaussian statistics and using the formal definitions of μ and σ from normal distributions. Using a pseudorandom number generator, we computed a number of rhythmite sequences derived from Gaussian distributions of intervals with various values of β . Representative results are shown in Figure 7.

As defined, β has two useful properties: (1) it is unitless and is therefore independent of the actual couplet spacing. This allows quick characterization of the rhythmite without *in situ* measurements: a photograph can be analyzed to determine β , and (2) both terms in the ratio σ/μ can be easily measured and understood: mean and standard deviation.

5. Temporally aperiodic, spatially periodic sedimentation

Here we present some concepts for producing rhythmites by temporally random deposition events. Although the processes described may not actually work in a real geological setting, the goal here is to stimu-

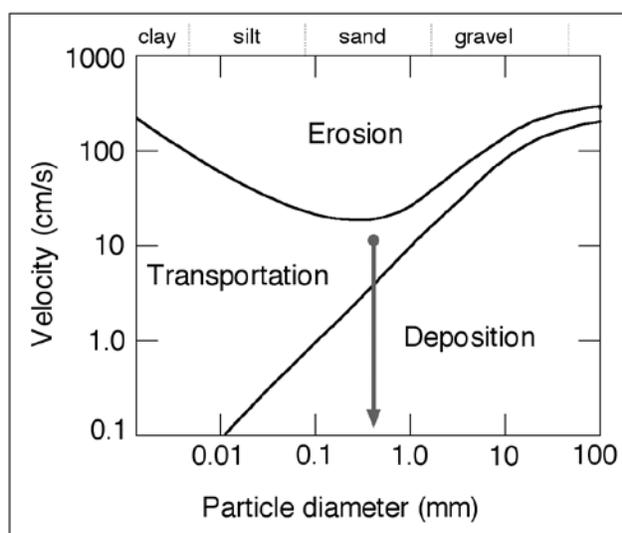


Figure 9. Hjulström diagram. If the overflow velocity from the basin begins in transportation and then diminishes into the deposition (never reaching the erosion regime), it will deposit a series of layers with equal thicknesses, regardless of the temporal deposition interval.

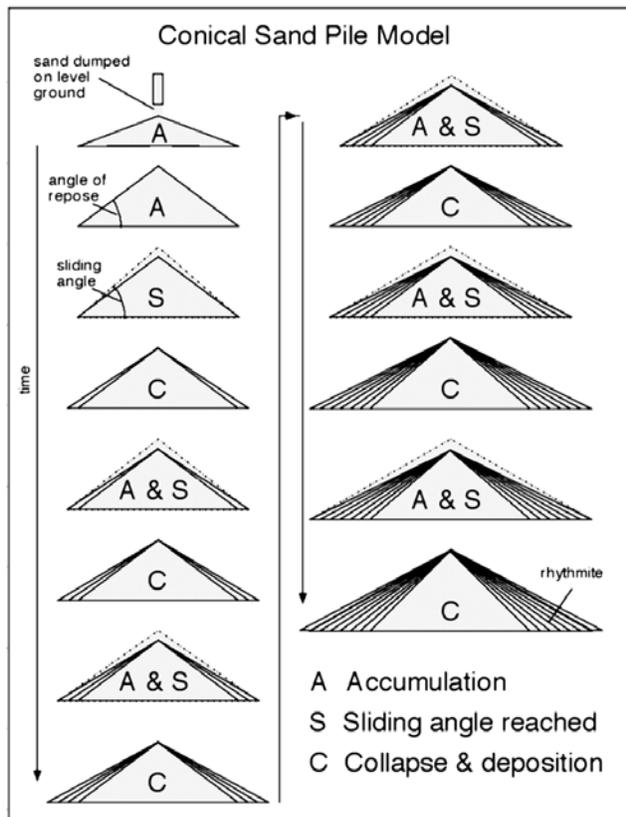


Figure 10. Threshold rhythmite production: Sand flowing episodically from a point source onto a flat surface. Between flows, the sand surface lies at the angle of repose AR. When new sand flows, it accumulates (A) at a steeper angle until this angle reaches the sliding angle AS where $AS > AR$. At this point (S) the sand collapses (C) and returns to the previous angle of repose. The process then repeats and with each collapse, a fixed amount of new material is deposited in the form a conical rhythmite.

late thinking and explore notions of how random or chaotic processes can produce regular, periodic structures. Some of these may involve nonlinear dynamics and recursive-defined systems.

One model for aperiodic rhythmite production was suggested by Miller (2008) in the form of a tipping bucket rain gauge. A vessel collects water until it reaches a preset value, at which time the bucket tips, discharges the water (and suspended sediments), then resets to begin collecting again. The process repeats. The amount of water discharged is the same in every tipping event, but bears no relation to the frequency of discharge because the flow rate of water into the bucket may vary. Can analogous processes occur naturally?

If so, we must search for mechanisms that (1) quantize sediment, (2) deposit these sediments at random intervals, and (3) repeat in a manner that is independent of the previous deposition.

With these conditions in mind, imagine a normally-undrained depression that receives rain and surface

runoff (Figure 8). Most of the time the basin does not overflow, but occasionally - after heavy rains - it does. When this happens, sediment in the overflow will be deposited in a deltaic structure. The time intervals for such overflows can vary widely as the hydrologic budget changes: it could be several times a year after each heavy rain, or only every few years when seasonal accumulation or unusually heavy rains result in overflow.

To understand how such events can lead to spatially rhythmic sedimentation, consider the following scenario (Figure 8). When the basin overflows, water and suspended sediments pour into an overflow delta. This can continue for a long time (weeks, months) or a relatively short period (days, hours). The actual duration of overflow is not relevant. If we assume that the flow velocity is high enough to prevent deposition but not high enough to cause erosion of the surface bed, then the water flows until the basin water level drops to the level of the dam. At this time, a quantized amount of water remains on the overflow fan, the same amount each time. With no water behind it, this fixed amount of water and sediment slows down as the water depth decreases. Eventually the flow velocity will decrease until it crosses the threshold between transportation and deposition (Figure 9). With a fixed amount of sediment in the now pinched-off flow, the deposited sediment thickness is also fixed.

This model assumes that the flow is not fast enough to cause erosion of the previously deposited bed. This can happen if the flow is just right, but can be further enhanced if the previously deposited bed is somewhat consolidated or cemented. If the interval between overflows is many years, the exposed bed may very well become consolidated and perhaps even cemented by calcification from the previous flow. This will raise the erosion-transportation threshold (dotted curve in Figure 8), perhaps significantly, thereby allow a larger range of flow velocities without inducing erosion than might otherwise occur.

Such a model can produce rhythmites from temporally random deposition events. Each layer will have the same thickness. But without at least two visually contrasting layers to form a couplet, the boundaries between each deposition event would not be discernable. Therefore, we need only suppose that the sediments contained in the run-off were of a binary nature, one component being distinct from the other in both color and settling rate (density or particle size). Such a situation is not difficult to imagine, i.e. sand and clay, silt and organic matter, etc.

At the time the overflow ceases (and before), the amount of water in the basin is also quantized, i.e., it is the same for each flood even. Providing that the basin dries out between each flood and that little or no deposition takes place in the basin during the flood, each deposit will also have the same thickness with each episode. Slight variations from year to year such as incomplete desiccation of the basin between floods can cause slight variations in the rhythmite intervals.

Another example of quantized deposition involves granular flow. We are all familiar with a conical sand pile produced in gravel quarries where the cone angle is the angle of repose (Figure 10). As new sand is deposited at the apex, it builds up an angle greater than the angle of repose until the surface inclination of the new material reaches the critical slipping angle at which time it slides down and deposits on the previous surface. The slumped material then resides at the angle of repose and the sequence repeats. The final configuration of the sediments will not depend on the input rate of the sand at the top of the pile. Although Figure 10 illustrates the process using conical geometry, the same thing can happen if the sediments flow over a cliff and build up along a linear front. In this case, successive layers would not be deposited horizontally as in most sedimentary bedding. A key diagnostic would involve vertical cuts through the sediment at different azimuth angles.

6. Summary and Conclusions

We have analyzed the composition, particle size and spatial intervals of Fish Creek rhythmites. They are found as triplets and quadruplets of distinct layers of silty, arkosic sandstone, claystone and siltstone. A unitless, easily-measured spatial bandwidth parameter is defined that characterizes the distribution of repeating rhythmite intervals. We have argued that rhythmites can be produced by temporally random processes and suggested three possible mechanisms. Each involves quantized sediments, independent deposition events and threshold triggering of the process.

7. Acknowledgements

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Holocene slip rate and tectonic role of the western Garlock fault

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Tectonic models

Three primary models have been proposed for the tectonic role of the Garlock fault (Figure 1). It has been interpreted (1) as a conjugate shear to the San Andreas fault (Hill and Dibblee, 1953) that helps to accommodate convergence at the major restraining bend in the San Andreas fault in southern California (Stuart, 1991), (2) as a transform fault accommodating extension in the Basin and Range (Davis and Burchfiel, 1973), and (3) as a structure accommodat-

ing block rotation in the northeastern Mojave (Humphreys and Weldon, 1994; Guest and others, 2003). Contrary to all three geologic models, geodetic data suggest that very little left-lateral strain is accumulating on the Garlock fault (Savage and others, 1981, 1990, 2001; Gan and others, 2000; Miller and others, 2001; Peltzer and others, 2001; McClusky and others, 2001). Rather, the region surrounding the Garlock fault has dominated by northwest-oriented right lateral shear for the past couple of decades.

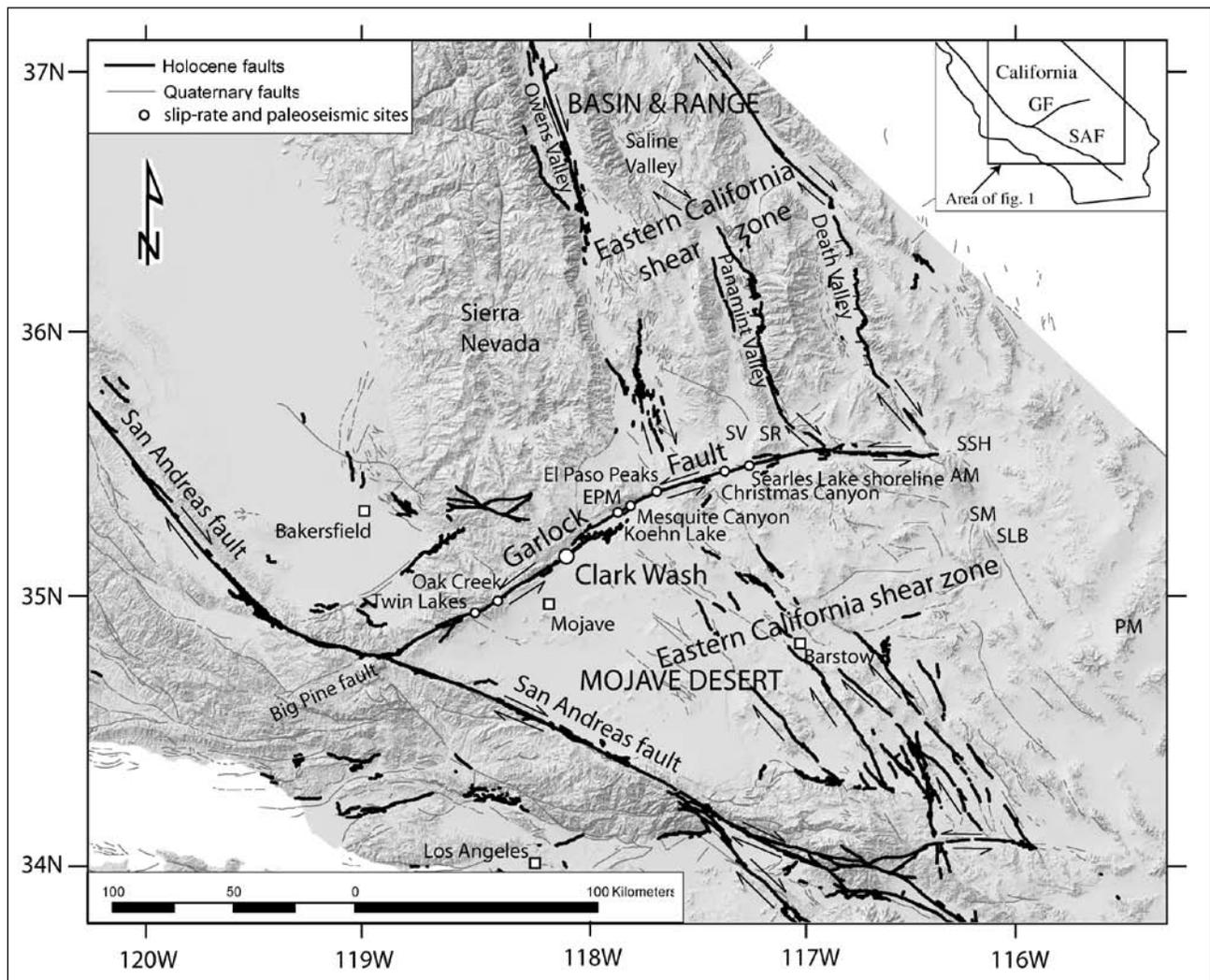


Figure 1: Location of the Clark wash site (large white circle) as well as other slip-rate and paleoseismic sites (small white circles) along the Garlock fault. AM, Avawatz Mountains; EPM, El Paso Mountains; PM, Providence Mountains; SLB, Soda Lake Basin; SM, Soda Mountains; SR, Slate Range; SSH, Salt Spring Hills; SV, Searles Valley.

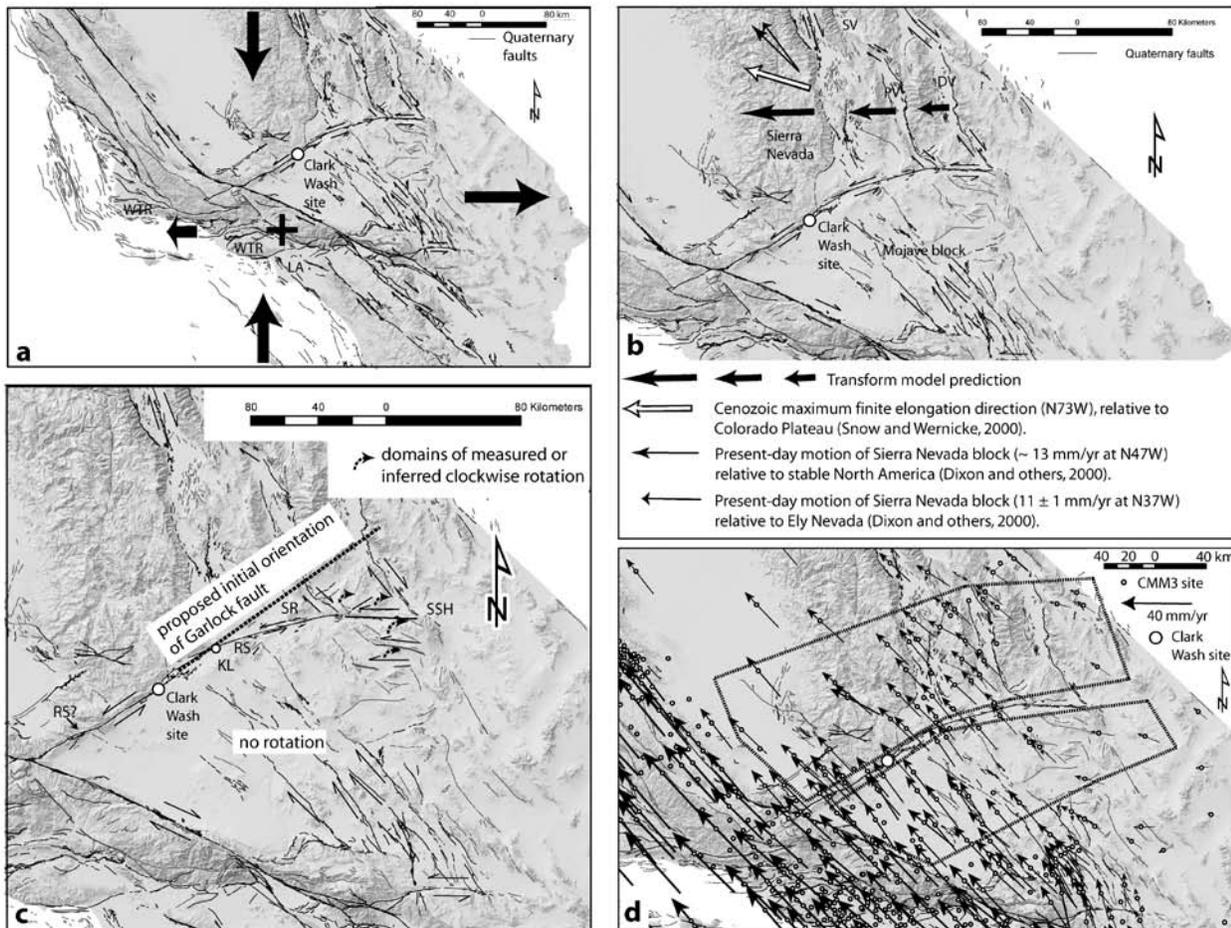


Figure 2: Proposed tectonic models for the Garlock fault. (a) Conjugate faulting (Hill and Dibblee, 1953). Convergence at the large restraining bend in the San Andreas fault is accommodated by a combination of eastward extrusion of the Mojave Desert and westward extrusion (Walls and others, 1998) and crustal thickening (indicated by large “+” symbol) (Argus and others, 1999) of the Transverse Ranges. LA, Los Angeles; WTR, western Transverse Ranges province. (b) Transform fault model (Davis and Burchfiel, 1973). DV, Death Valley; PV, Panamint Valley; SV, Saline Valley. (c) Model in which the eastern Garlock fault accommodates clockwise block rotation in the northeastern Mojave Desert (Guest and others, 2003). KL, Koehn Lake slip rate site (Clark and Lajoie, 1974); RS, Rand schist; RS?, schist between stands of Garlock fault zone that may correlate with Rand schist; SSH, Salt Spring Hills; SR, Slate Range. (d) Geodetic velocity vectors from the Southern California Earthquake Center’s Crustal Motion Model, version 3 (CMM3) (<http://epicenter.usc.edu/cmm3/>) relative to a reference frame defined by twelve stations on the North American plate. Vectors shown have been arbitrarily selected from the complete data set to reduce clutter.

(1) Conjugate fault model

Hill and Dibblee (1953) viewed the left-lateral Garlock and Big Pine faults and the right-lateral San Andreas fault as conjugate shears defining a regional strain pattern of north-south compression and east-west extension (Figure 2a). In this view the Garlock and San Andreas faults both accommodate eastward motion of the Mojave block as it extrudes from Transverse Ranges restraining bend along the San Andreas fault in southern California. Consistent with this model, left-lateral faulting of Quaternary age in California is largely confined to the vicinity of this regional-scale restraining bend in the San Andreas fault (Figures 1 and 2a).

(2) Transform fault model

Other investigators proposed a second model in which the Garlock fault is a transform fault (Figure 2b), with left-slip on the Garlock fault accommodating differential extension in the Basin and Range province (between the Sierra Nevada and Death Valley) relative to the Mojave block (Hamilton and Myers, 1966; Troxel and others, 1972; Davis and Burchfiel, 1973). The location of the Garlock fault at the southern margin of the western Basin and Range province is consistent with the transform model, as is the eastward termination of the Garlock fault at the eastern limit of significant Quaternary extension in the western Basin and Range province (Figures 1 and 2b). However, the present-day extension direction

in the Basin and Range province is northwestward (Dixon and others, 2000; Minster and Jordan, 1987), nearly perpendicular to the northeast- to east-striking Garlock fault (Figure 2b). This is difficult to reconcile with a transform fault model. Modern deformation in the portion of the Basin and Range province north of the Garlock fault is largely northwest-oriented dextral shear (Figure 2d). Late Quaternary extension north of the Garlock fault appears to be largely concentrated within pull-apart basins (Death Valley, Panamint Valley, Saline Valley) between northwest-striking, right-lateral faults. It is these right-lateral faults, rather than the Garlock fault, that are parallel to the present-day extension direction and appear to be serving as transform faults for the Late Quaternary extension north of the Garlock fault (Figure 2b).

The Cenozoic extension direction in the Basin and Range province is west-northwestward (Wernicke and others, 1988; Snow and Wernicke, 2000; Stewart, 1983; Burchfiel and others, 1987; Jones, 1987). This orientation more closely approaches the strike of the central and eastern Garlock fault, but is still at a 45-degree angle to the western Garlock fault. The transform fault model may thus be a partially viable model for the initiation of left-slip on the central and eastern Garlock fault (if that portion of the fault has not been rotated—see third model below), but it is unable to explain the orientation of the western Garlock fault, nor does present-day extension seem capable of driving a significant amount of left-slip on any part of the Garlock fault (Figure 2b,d).

Since the recognition of the Eastern California shear zone (Dokka and Travis, 1990), with through-going northwest-oriented right-lateral shear both north and south of the Garlock fault (e.g., Savage and others, 1990) (see figure 2d), it is no longer easy to view the Garlock fault as an accommodation structure separating two provinces with radically different tectonic regimes. Instead, the Garlock fault appears as an enigmatic anomaly within a single large province, the Eastern California Shear Zone (ECSZ). Both north and south of the Garlock fault, the ECSZ (Figure

1) has accommodated tens of kilometers of northwest-oriented right lateral shear in the late Tertiary (Dokka and Travis, 1990; Stewart, 1983; Burchfiel and others, 1987), at rates of 5-10 mm/yr across the zone since the late Pleistocene (Oskin and others, 2006; Lee and others, 2001; Oswald and Wesnousky, 2002). Geodetic data (Figure 2d) indicate that the Garlock region is currently dominated by northwest-trending right-lateral strain rather than by northeast-trending left-lateral strain (Savage and others, 1981, 1990, 2001; Gan and others, 2000; Miller and others, 2001; Peltzer and others, 2001). Present-day, northwest oriented right lateral shear has been documented at rates of 9-11 mm/yr north of the Garlock fault (Meade and Hager, 2005; Dixon and others, 2000) and ~ 7-15 mm/yr south of the Garlock fault (Saubert and others, 1986; 1994; Meade and Hager, 2005).

Despite the prominence of the ECSZ, the Garlock fault arcs northeasterly to easterly through this zone (Figure 1), with undeniable left-lateral bedrock offsets of 48-64 km (Smith, 1962; Smith and Ketner, 1970; Davis and Burchfiel, 1973; Carr and others, 1993; Monastero and others, 1997; Jachens and Calzia,

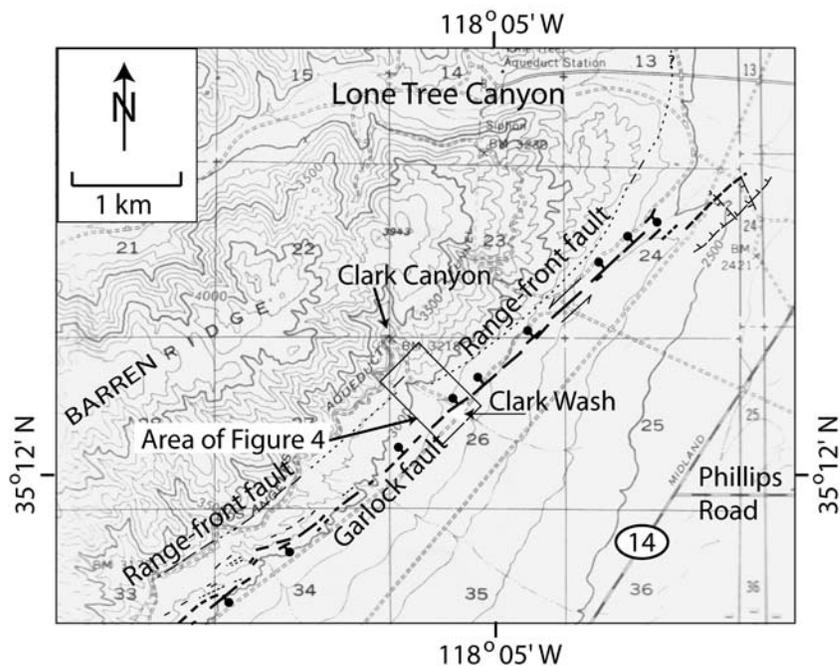


Figure 3: Location of offset channel of Clark Wash. Base map is the Mojave NE, 15-minute topographic quadrangle published by the U.S. Geological Survey. The site is located near the center of Section 26, T. 31 S., R. 36 E. Garlock fault location is from Clark (1973). Heavy lines mark the youngest fault traces; thinner lines mark older fault traces. Bar and ball on lower side of fault scarps with predominately horizontal movement. Hachures on relatively downthrown side of fault scarps with primarily vertical movement. Approximate location of range-front fault is from Dibblee (1959), dotted where buried. Note that the canyon labeled "Lone Tree Canyon" in this figure and on the Mojave 15-minute quadrangle is labeled as "Pine Tree Canyon" on the Mojave NE 7.5 minute quadrangle.

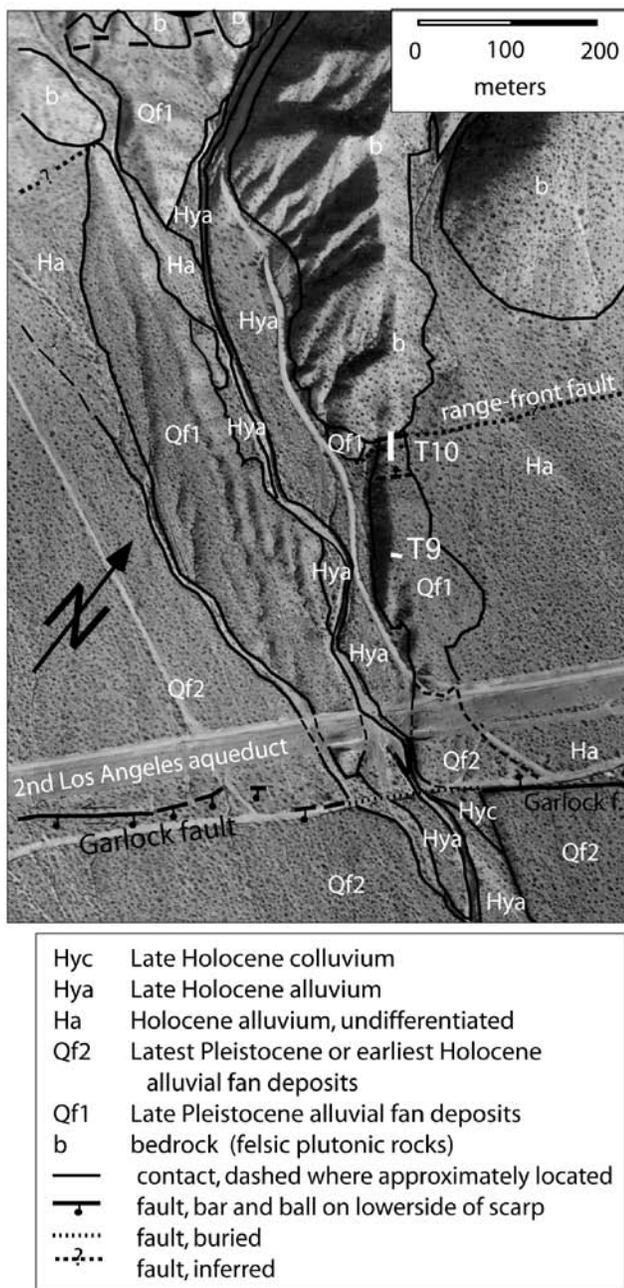


Figure 4: Simplified geologic map of Clark wash. Base is a U.S. Geological Survey aerial photograph, courtesy of Malcolm Clark.

1998) and a Holocene slip rate between 4 and 9 mm/yr (McGill and Sieh, 1993; Clark and LaJoie, 1974). Dextral shear in the ECSZ may have rotated the central and eastern sections of the Garlock fault from an original northeastward strike to their current east-northeast and eastward strikes (Figure 2c) (Jones, 1987), but the Garlock fault has remained an important, through-going fault that has produced several large earthquakes in the past few thousand years (Dawson and others, 2003; McGill and Sieh, 1991; McGill, 1992) and is still seismically active (Astiz and Allen, 1983). Evidence for Holocene left-lateral slip

is abundant and nearly continuous along the entire ~250-km length of the fault (Clark, 1970; Clark, 1973; McGill and Sieh, 1991), whereas the right-lateral faults to the north and south die out (e.g., Oskin and Iriondo, 2004) or exhibit possible left-lateral drag folding as in the case of the southern Death Valley and Panamint Valley faults, as they approach the Garlock fault (Figure 1).

(3) Accommodation structure for block rotation

In light of recent understanding of dextral shear and vertical-axis block rotation in the ECSZ (Dokka and Travis, 1990; Luyendyk and others, 1985; Ross and others, 1989; Carter and others, 1987; Schermer and others, 1996), Humphreys and Weldon (1994) and Guest and others (2003) have suggested a third model for the Garlock fault in which clockwise rotation of blocks in the northeastern Mojave Desert contributes left slip to the eastern part of the Garlock fault (Figure 2c). Guest and others (2003) have shown that a combination of large magnitude (~35° or more), domino-style, clockwise rotation of blocks (and of their left-lateral bounding faults) both south and north of the Garlock fault and a moderate amount of extension north of the Garlock fault between the Slate Range and the Salt Spring Hills (Figure 2c) can accommodate the full 48-64 km of left-slip on the Garlock fault without requiring a search for an elusive eastward extension of the Garlock fault (e.g., Plescia and Henyey, 1982) that was previously thought necessary (Davis and Burchfiel, 1973).

Predictions of the three tectonic models

These three different models for the tectonic role of the Garlock fault each result in different predictions for the slip rate of the fault as a function of position along strike. The conjugate fault model (Figure 2a) predicts a slip rate that would gradually decrease eastward, away from the San Andreas fault. The transform model predicts an incrementally westward-increasing slip rate, as each normal fault north of the Garlock fault contributes additional left slip to the fault (Figure 2b). The rotating block model (Figure 2c) primarily explains slip on the eastern Garlock fault and predicts little or no slip on the western Garlock fault (Humphreys and Weldon, 1994).

Slip rate of the western Garlock fault

Recent results from the Clark Wash site, near Lone Tree Canyon on the western Garlock fault, indicate that the Holocene slip rate of the western strand

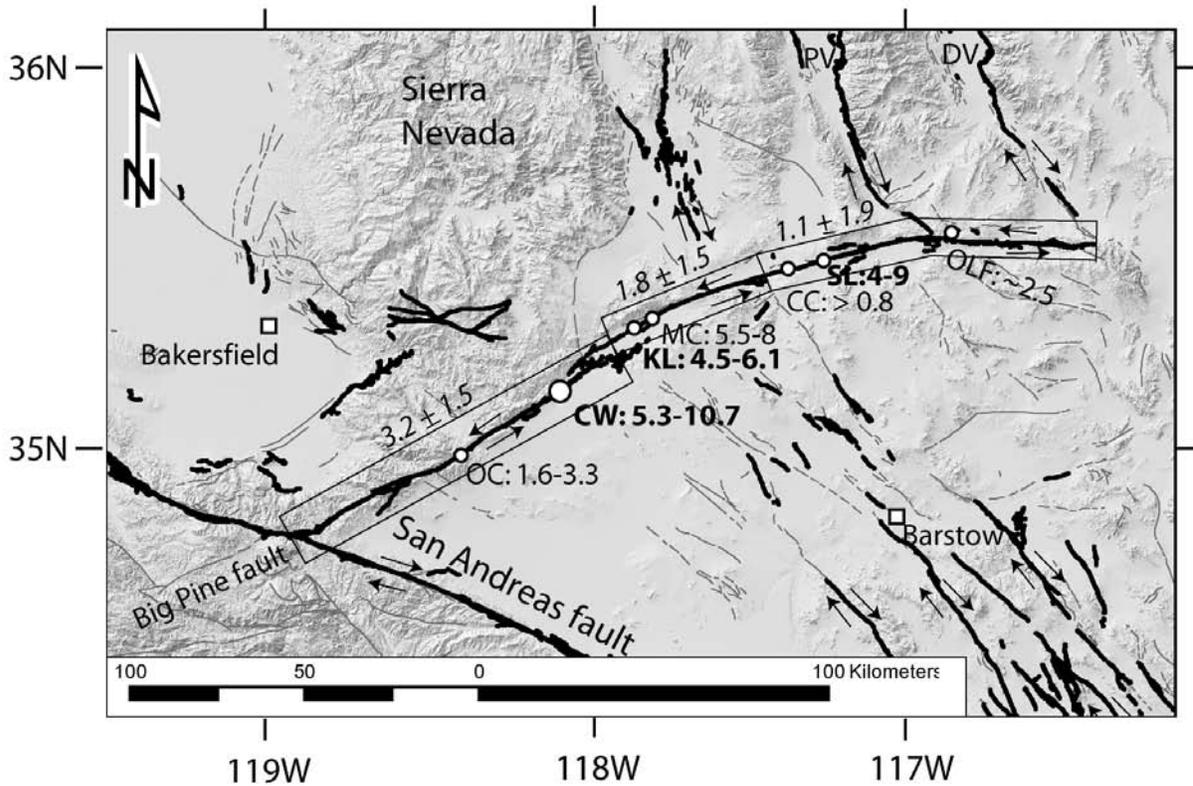


Figure 5: Comparison of slip rate estimates for the Garlock fault. The three values in italics, associated with boxes that outline sections of the fault, are the slip rates and formal uncertainties from Meade and Hager's (2005) best-fitting elastic block model of available geodetic data. They report, however, that experience with a range of models suggests that true uncertainties are ~ 3 mm/yr. White-filled circles mark the locations of Holocene and Late Quaternary geologic slip rate estimates. The Holocene rates that are constrained by radiocarbon dates and are thus considered most reliable are shown in bold (CW, Clark Wash, this study; KL, Koehn Lake, Clark and Lajoie, 1974; SL, Searles Lake, McGill and Sieh, 1993). Other Holocene and Late Quaternary rates are shown in plain text (CC, Christmas Canyon, Smith [1975], minimum rate based on an offset channel incised into a dated shoreline that may be much older than the channel; MC, Mesquite Canyon, Carter and others [1994], age control from fossil identification; OC, Oak Creek, La Violette and others [1980], age control from degree of soil development; OLF, Owl Lake fault, McGill, [1998b], age control from radiocarbon dating of rock varnish).

of the Garlock fault is at least as fast as previously published rates along the central section of the fault (McGill and others, 2009; see reprint in pocket of printed guidebook). The Clark wash site (informally named for Malcolm Clark, who first noted the left-lateral offset of this drainage [Clark, 1973]) is located on an alluvial fan complex that drains southeastward from the southern Sierra Nevada (Figs. 3 and 4). The intersection of Clark wash with the Garlock fault is located near the center of section 26, T. 31 S., R. 36 E. (N35.205°, W118.087°), about 3.5 km southwest of Lone Tree Canyon (Fig. 3) and about 18 km northeast of the town of Mojave. The site is located about 3 km west of the western end of the 3.5-km-wide left step in the Garlock fault that separates the western section of the fault from the central section (Fig. 1). Koehn Lake occupies the center of a large closed depression that has formed within this 3.5-km-wide dilational step-over in the fault. Within the study area the dominant strand of the Garlock fault is the segment

that enters the Koehn Lake step-over from the west. However, the central segment of the fault, which enters the Koehn Lake step-over from the east, may continue westward into the study area as a buried fault along the southern range front of the Sierra Nevada (Smith, 1964). This fault strand is here referred to as the range-front fault.

Clark wash has been left-laterally offset at least 66 ± 6 m and no more than 100 meters across the western Garlock fault, indicating a left-lateral slip rate of 7.6 mm/yr (95% confidence interval of 5.3-10.7 mm/yr) using dendrochronologically calibrated radiocarbon dates. The timing of aggradational events on the Clark Wash fan corresponds closely to what has been documented elsewhere in the Mojave Desert, suggesting that much of this activity has been climatically controlled. The range-front fault, located a few hundred meters northwest of the Garlock fault, has probably acted primarily as a normal fault, with a Holocene rate of dip-slip of 0.4 to 0.7 mm/yr. The

record of prehistoric earthquakes on the Garlock fault at this site, though quite possibly incomplete, suggests a longer interseismic interval (1200-2700 years) for the western Garlock fault than for the central Garlock fault.

Discussion

The new Holocene slip rate measurement for the western Garlock fault at Clark Wash is at least as fast as previously published Holocene and Late Quaternary rates along the central strand of the fault (Figure 5). The high rate of motion on the western Garlock fault is most consistent with a model in which the western Garlock fault acts as a conjugate shear to the San Andreas fault. Other mechanisms, involving extension north of the Garlock fault and block rotation at the eastern end of the fault may be relevant to the central and eastern sections of the fault, but they cannot explain a high rate of slip on the western Garlock fault.

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New Pleistocene faunas from Kern River terrace deposits, Ming Lake, Kern County, California

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Abstract: Excavation monitoring east of Bakersfield recovered vertebrate fossils from both middle and late Pleistocene sediments. Vertebrate fossils, which have been deposited in the collections of the Buena Vista Museum of Natural History (BVMNH) and referred to as the Upper Eagle Crest Fauna and the Lower Eagle Crest Fauna, are described. This report describes vertebrate fossils in the collections of the Buena Vista Museum of Natural History (BVMNH) referred to as the Upper Eagle Crest Fauna and the Lower Eagle Crest Fauna. These faunas occur in Pleistocene sediments overlying Miocene sediments at elevations of 750 feet (229 m) and 580 (177 m) feet, respectively. The Upper Eagle Crest Fauna produced a camel metapodial attributed to *Camelops* sp. The Lower Eagle Crest Fauna was located on a Pleistocene river terrace approximately 40 feet (12 m) feet above the current Kern River channel. This terrace deposit produced 23 taxa of invertebrates and vertebrates which include a small late Pleistocene horse (*Equus* sp.) and an extinct diminutive antelope (*Capromeryx* sp.). A bat and mole represent the first late Pleistocene records of insectivores in the Kern County portion of the southern Central Valley. The muskrat (*Ondatra zibethica*) is the first record of this large aquatic rodent from the Kern County portion of the southern Central Valley. Bat, mole, and muskrat represent the first late Pleistocene records of insectivores and the large aquatic rodent in the Kern County portion of the southern Central Valley. The fauna includes fish, frog, pond turtle, and the first Pleistocene record of the extinct freshwater ridged mussel (*Gonidea angulata*) from the Kern River.

Location and Geology

The Eagle Crest upper and lower faunas were recovered south of Ming Lake, on the south side of the Kern River in eastern Bakersfield. Ridges exposed on the south side of the Kern River are supported by the Round Mountain Silt (Bartow, 1981). As the Kern River incised downward through the continental late Miocene Kern River Formation, the marine late Miocene Santa Margarita Formation, and the marine Middle Miocene Round Mountain silt, its meanders created terraces along the river gorge. These terraces contain deposits of Pleistocene sediments that are younger at lower elevations. Geologic mapping refers to these sediments as the younger and lower Qoa1 (approximately 10ka–100ka) and the older and higher Qoa2 (approximately 1 Ma; Bartow, 1981). The stratigraphic relationship of Qoa2 and Qoa1 on

the Miocene Round Mountain Silt is shown in Figure 1 (from Bartow, 1981).

Middle Pleistocene Qoa2 sediments on mid-elevation ridges occur between elevations 720–780 ft (approx. 220–238 m). The Qoa2 sediments are easily recognized by the large (12+ inch; 30 cm) round

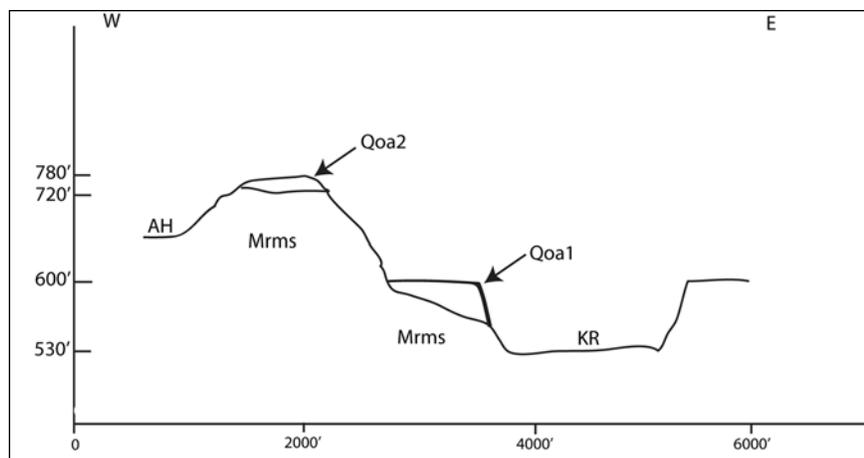


Figure 1. Geologic cross section south of Ming Lake, view north. Mrms—Miocene Round Mountain Silt; Qoa2—Middle Pleistocene sediments; Qoa1—Late Pleistocene sediments; AH—Alfred Harrell Highway; KR—Kern River. [Geology from Bartow, 1981].

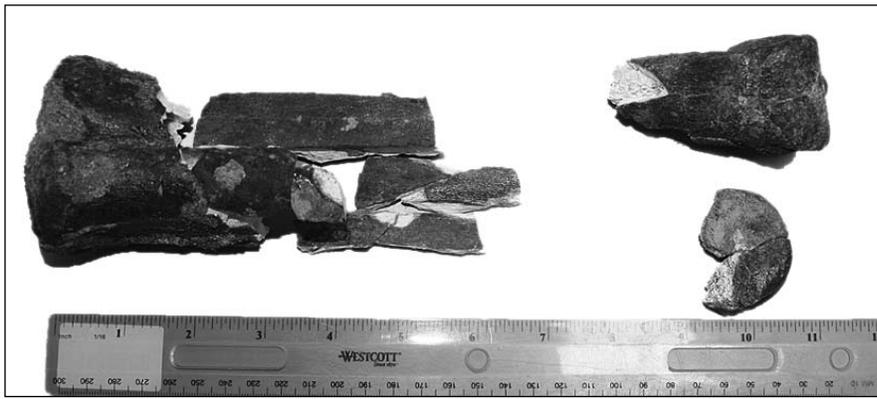


Figure 2. *Camelops* sp. (camel) metacarpal. BVMNH nK127.



Figure 3. *Gonidea angulata* (ridged mussel) valve pair. BVMNH nK001.

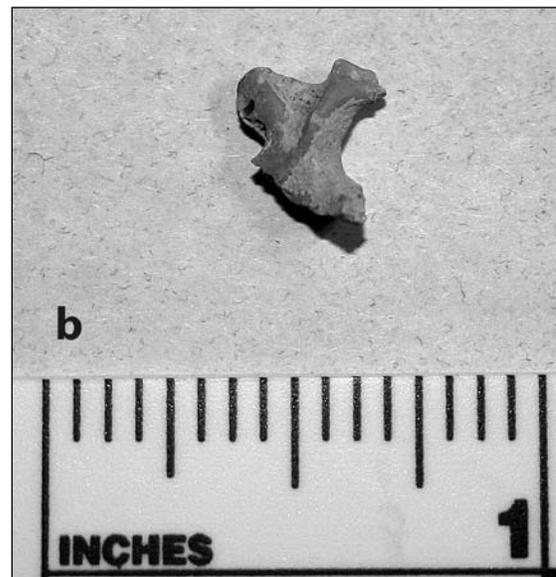
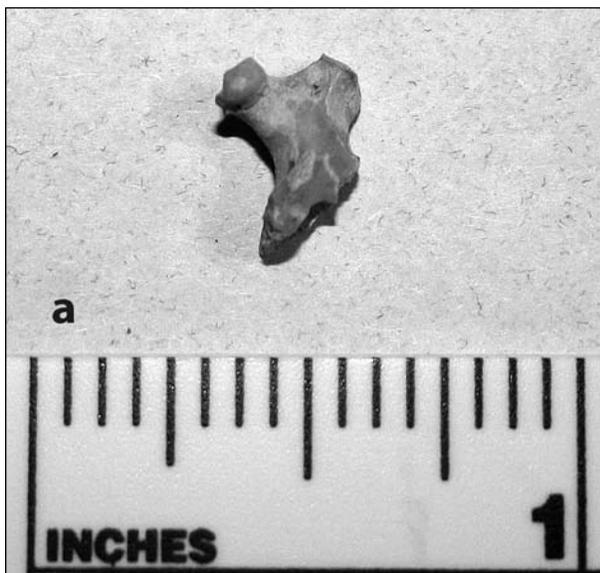


Figure 4a, b. *Scapanus* sp. (mole) left humerus. a: anterior view; b: posterior view. BVMNH nK120.

cobbles that they contain, probably representing deposits by a river with capable of significant flows. Brown interstitial sands fill spaces between the cobbles.

The late Pleistocene fossils on the lower terrace (Qoa1) approximately 40 feet (12 m) above the current Kern River bed between elevations 560 and 600 feet (approx. 170–183 m). These terrace sediments are medium-coarse, well sorted sands suggesting deposition by a broad river meander followed by a covering of alluvial slope wash and formation of soils. Fossiliferous sands have black stains and encrustations of pyrolusite, and contain lenses and stringers of gravel and cobbles.

Large (2 cm) pieces of charcoal in the sands suggests forest fires upstream, with deposition in downstream eddies. Charcoal samples preserved in the collections of the BVMNH are available for radiometric dating.

Middle Pleistocene *Camelops*

The upper and lower faunas are apparently the first records fossils from the Kern River terraces east of central Bakersfield. The upper and older deposits of Qoa2 produced the fossil metatarsal of an Ice Age camel (*Camelops* sp.). If the age inference of 1 Ma is correct (Bartow, 1981) it would place the fossil in the Irvingtonian North American Land Mammal Age (NALMA). The proximal and distal articula-



Figure 5. *Ondatra zibethesis* (muskrat) distal humerus. BVMNH nK123.



Figure 6. *Capromeryx* sp. (small antelope) rear ungual phalanx. BVMNH nK126.



Figure 7. *Equus* sp. (horse) right lower second molar. BVMNH nK130.

tions of the metapodial compare in dimensions to *Camelops* (Webb, 1965). The shaft of the metapodial is broken, preventing accurate measurement of length. This metapodial represents the first middle Pleistocene record of a camel from Kern County (Jefferson, 2008a).

Late Pleistocene Fauna

The younger Pleistocene deposits on the lower terrace (Qoa1) contain a diverse fauna of mollusks, lower vertebrates, and small and large mammals. Late Pleistocene vertebrates from the Kern River terraces include fresh water fish, frogs or toads, and pond turtles. Small birds, lizards, and snakes were present on adjacent river banks. The assemblage includes insectivores—bats and moles—that are rare in any fossil fauna. Additional small mammals include rabbits,

rodents, and muskrat. Large herbivorous mammals include a small Ice Age horse, diminutive pronghorn antelope, and deer. The horse and small antelope strongly indicate a late Pleistocene fauna, which is consistent with the low position of the terrace above the current river channel.

Late Pleistocene Habitats

Certain fossil invertebrate and vertebrate fossil taxa help provide proxy habitat data for the area during late Pleistocene time. The large mammals suggest a parkland or grassland with discrete clumps of brush and trees. Gophers and moles would have inhabited

Table A: Late Pleistocene fauna from Kern River Qoa1

Family	Scientific Name	Common Name
Pelecypoda	<i>Gonidea angulata</i>	freshwater ridged mussel
Pisces	Osteichthyes	freshwater bony fish
Amphibia	Urodela	frog or toad
Testudinoidea	<i>Actinemys</i> (formerly <i>Clemmys</i>)? sp.	freshwater turtle
Emydidae		
Reptilia	Lacertilia	lizards
	Colubridae	non-venomous snakes
Aves	Fringillidae	small birds
Mammalia		
Insectivora	Chiroptera	bat
	<i>Scapanus</i> sp. cf. <i>S. latimanus</i>	mole
Lagomorpha	<i>Sylvilagus</i> sp.	cottontail
	<i>Lepus</i> sp.	jack rabbit
Rodentia	<i>Eutamias</i> sp.	chipmunk
	<i>Peromyscus</i> sp.	deer mouse
	<i>Neotoma</i> sp.	wood rat
	<i>Microtus</i> sp.	vole – meadow mouse
	<i>Perognathus</i> sp.	pocket mouse
	<i>Dipodomys</i> sp.	kangaroo rat
	<i>Dipodomys</i> sp. (lg)	large kangaroo rat
	<i>Thomomys</i> sp.	pocket gopher
Artiodactyla	<i>Ondatra zibethica</i>	muskrat
	<i>Capromeryx</i> sp.	small antelope
	<i>Odocoileus</i> sp.	deer
Perissodactyla	<i>Equus</i> sp. (sm.)	small Ice Age horse

moist, brush- and grass-covered, friable soil. Small mammals (rodents and rabbits) and birds, snakes, and lizards have less specific habitat requirements, but certain species require open terrain while others require brushy terrain or trees. Mussels, frogs or toads, turtle, and fish require fresh water. As a whole, the Pleistocene fauna may have been equivalent to what is present today, which is described as lower to upper Sonoran life zone with a dense riparian (streamside) habitat surrounded by grassy parklands bordering a hilltop woodland (Schoenherr, 1992).

Summary

These fossils include a new geographic range extension for the large middle Pleistocene camel, *Camelops* sp. (Jefferson, 2008a) The presence of fresh water mussel *Gonidea angulata* is a new southern geographical record for the state of California; the closest living occurrences of *Gonidea angulata* are north of Bakersfield at Clear Lake and Los Banos (Nedeau and others, 2008). If the mussel previously lived in the Pleistocene Kern River, it has apparently been extirpated. The bat and mole represent the first late Pleistocene records of insectivores in the Central Valley portion of Kern County (Jefferson, 1991; Jefferson, 2008b; UCMP, 2008). The muskrat (*Ondatra zibethica*) is the first record of this large rodent from the Central Valley portion of Kern County (Jefferson, 1991; Jefferson, 2008b; UCMP, 2008). The fossiliferous Pleistocene terrace deposits appear to be new sources of faunas that may provide additional age constraints regarding the down cutting of the Kern River, and the maximum age of the older Kern River Formation.

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Owens pupfish: a brief overview

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The Owens pupfish (*Cyprinodon radiosus*) in the early 1960s was thought to be extinct due to the dewatering of Fish Slough, north of Bishop, caused by activities of the Los Angeles Department of Water and Power. The fish were rediscovered in 1964 by C.L. Hubbs, R.R. Miller, and E.P. Pister and plans were made to create a refuge. In 1970, by the time the refuge was ready to be stocked, the marshy pools of Fish Slough were nearly dry. Pister and his crew at CDFG rescued about 800 fish that were temporarily placed in artificial cages. Unfortunately, poor water quality nearly extirpated the stock. At one point Phil Pister had the entire population in two buckets which held the total remnants of the once wide-spread species. At first, the species was protected in seven different refuges. Today the fish are abundant in two. One is in the Fish Slough area and the other near Lone Pine is called Well 368. Unfortunately, without constant monitoring the refuges become threatened by encroachment of aquatic plants such as tules and cattails, or the introduction of non-native fish species. Local fishermen seem intent on introducing the predatory largemouth bass to the refuges.

Pupfish are in the tooth-carp family, Cyprinodontidae. In California, this includes killifish, which are mostly estuarine. Counting adjacent Nevada there are seven species and six subspecies of pupfish and killifish. Most of them occur in isolated springs in the Death Valley system. The Devils Hole pupfish, *Cyprinodon diabolis*, inhabits a limestone fissure in the Nevada portion of Death Valley National Park. Its total habitat of 180 square feet (20 m²) of surface is the smallest known distribution of any vertebrate species. The pupfish at the Desert Studies Center at Zzyzx are the Amargosa pupfish, *Cyprinodon nevadensis*, whose native range is in the Amargosa River.

Pupfish are noted for their great tolerances for environmental extremes. For thousands of years these species have existed in isolated habitats subject to environmental perturbations, but no other fishes. The Owens pupfish has been observed under ice in Fish Slough, and the Desert pupfish, *Cyprinodon macularius*, at 114EF (44.6EC) holds the record for the highest recorded body temperature for a fish. The Cottonball Marsh pupfish, *Cyprinodon salinus milleri*, has been observed in saline water five times the concentration

of sea water, and all species are able to move from freshwater to sea water and back with no apparent ill effects. Also recorded for the Desert pupfish is the lowest tolerated minimum for dissolved oxygen at 0.13 mg/l. In spite of these great tolerances, most of these species are classified as threatened or endangered, their greatest enemy being habitat destruction or introduction of non-native species.

Present distribution of these fishes in isolated spring fed habitats is enigmatic. Ancestors of these little fish may have been associated with a Miocene marine embayment known as the Bouse Embayment. Retreat of this water may have left the remnant populations stranded in isolated bodies of water where they evolved in association with local conditions, i.e. carbonate springs, saline springs, hot springs, and ephemeral streams. For many years, proposed connections between Pleistocene lakes in the Death Valley system and the Colorado River were deemed responsible for dispersal of these species from Owens Valley to the mouth of the Colorado River in Mexico. Today, geologists are not sure the connections existed, and new evidence indicates the Bouse Embayment was freshwater. Relationships among the species are further complicated by DNA analysis indicating that the southernmost form, present in the Salton Sea and lower Colorado River, is most closely related to the Owens pupfish, the northernmost form.

Recent records of fossil fish from eastern Owens Lake, Inyo County, California

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Introduction

The Owens River drainage separates the Sierra Nevada geologic province from the Basin and Range province. The river runs from northwestern Long Valley south to Owens Lake, its current terminus. In Pleistocene times, the river flowed south to China Lake and east to Searles Lake and Panamint Valley. Lacustrine sediments of Lake Waucobi to the north, dated at 3–2.3 Ma (Bachman, 1974, 1978) and the Coso Formation (containing fossil bony fish; Schultz, 1937; Jefferson, 2008) south of Owens Lake suggest that the development of Owens Valley and the internal drainage has probably been ongoing since the late Pliocene.

Owens Lake basin is located east of the Sierra Nevada range, southwest of the Inyo Mountains and northwest of the Coso Range. High stand lake shorelines from the Tahoe glacial (65 ka) at elevation 3,881 ft (Bates, 2009) reach frontal faults of all three ranges. Shoreline elevations project a lake 250 feet deep.

The Owens Lake Project involves grading to level the current surface of Owens Lake to allow Owens River water to cover the desiccated surface and to re-establish salt grass vegetation, thus preventing wind erosion. Environmental monitors are present to observe potential impacts to resources, and the paleontological monitor (R. Serrano), recovered fossil fish remains in February 2009.

Stratigraphy

Partially articulated and disarticulated fish skeletal material, teeth, and scales were recovered on berms south of Keeler and west of the junction of highways 190 and 136. The area producing fossils is referred to herein as “Keeler South.” Two facies produce fish remains; the lower (e. 3564 ft) is referred to as “Keeler South: Owens Lake Clay,” a gray-green silty clay with abundant mica (L1 Lawrence Clay; Soil and Water West, 2002). Fish skeletons are partially articulated suggesting that oxygen was at a minimum and currents did not move them far after death. This silty clay

has dried, and is cut by vertical desiccation cracks.

The source of the second facies is 20 feet higher (el.3580–3590 ft) and is referred to as “Keeler South: Owens Lake Sand,” an iron oxide-stained, brown, loamy sand that covers the silty clay and fills desiccation cracks. Fish elements in the sand are disarticulated, abraded, and altered to hydroxyl apatite.

The position of the Owens Lake Sand fossils in eastern Owens Lake suggests that current drainage patterns could not have carried them onto the lake from the late Pliocene Coso Formation (3.0–2.5 Ma; Lundelius, et al., 1987; Jefferson, 2008) that crops out south of the lake in the northern Coso Mountains (Jennings, 1958; Duffield and Bacon, 1981). The source of the fossils higher in the Owens Lake section and their alteration to hydroxyl apatite suggest that they are older than fossil fish from Owens Lake Clay, and are perhaps of middle Pleistocene age.

Owens Lake fossil fish

The Owens River drainage is currently inhabited by four species of native fishes: *Catostomus fumeiventris* Miller (Owens sucker), *Siphateles bicolor snyderi* (Miller) (Owens tui chub), *Rhinichthys osculus* (Girard) (speckled dace), and *Cyprinodon radiosus* Miller (Owens pupfish). Robert R. Miller contributed five papers and three of the four descriptions of the fishes of the Owens Basin (Hubbs and Miller, 1948; Miller, 1973). Fossils of two of these, *Catostomus fumeiventris* and *Siphateles b. snyderi*, as well as two additional species, *Prosopium williamsoni* (Girard) (mountain whitefish) and *Oncorhynchus clarki* (Williamson) (cutthroat trout), were recovered in the Owens Lake core (Firby et al., 1997). Remains of *Catostomus fumeiventris* and *Siphateles b. snyderi* were also recovered from Native American middens near the Alabama Hills (Smith et al., 2002). Other faunas important to understanding of Owens River fishes include the Lahontan and Mohave Basin faunas, recent and fossil, the Pliocene Lake Russell fish fauna east of Mono Lake, Pliocene Honey Lake, and Pleistocene

Mopung Hills, China Lake, Lake Manix, and Fort Rock Basin Oregon (Fig. 1).

This report is based on fossil fish bones with sufficient morphological details to provide clues concerning the relationships of the fishes and evidence for past hydrographic connections of the Owens Basin. Two common forms are represented in the Owens Lake clay: *Catostomus fumeiventris* and *Siphateles bicolor*; forms similar to each of these are represented in the Keeler South clay and sand deposits on the east side of the lake. The bones are compared to relevant recent species as well as to Pliocene and Pleistocene fossil species with which they might have had hydrographic and genetic connections (Fig. 1).

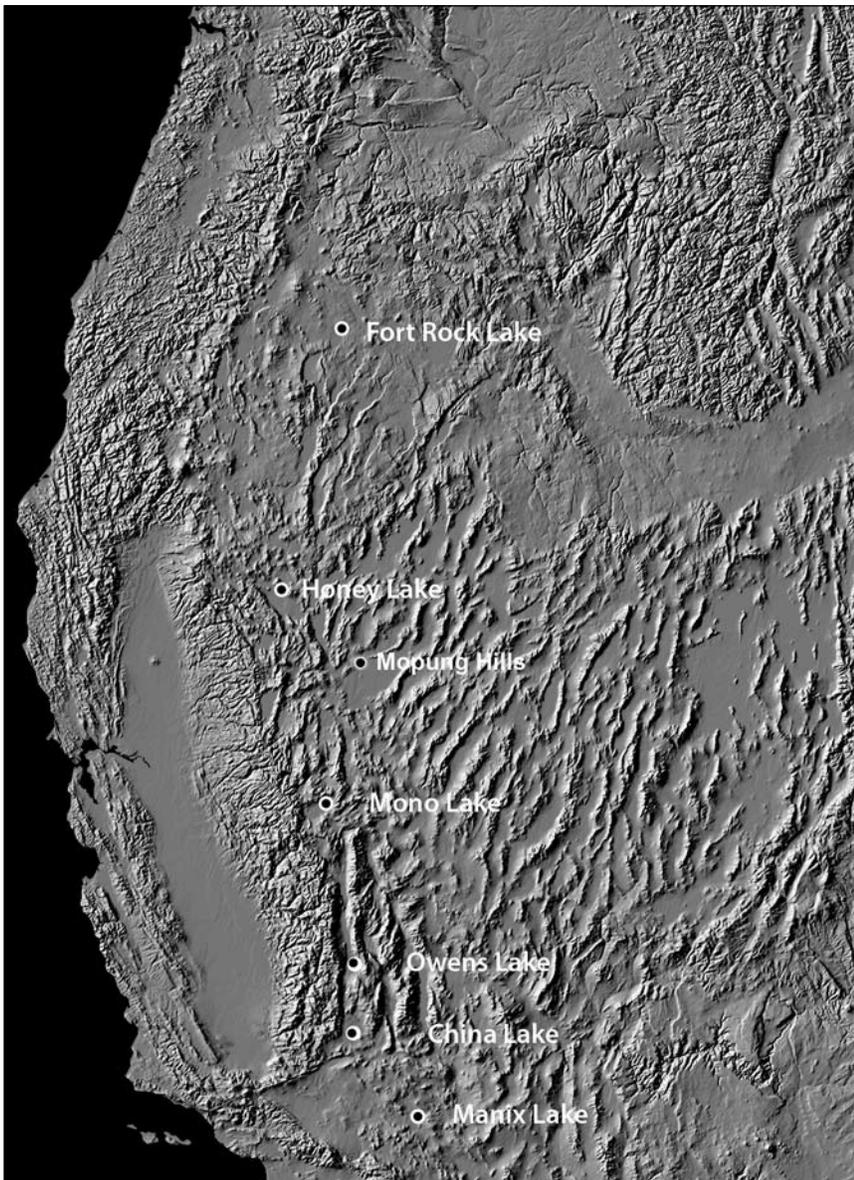


Figure 1. Distribution of Plio-Pleistocene records of *Siphateles* spp. The genus now extends to the eastern margin of the Lahontan Basin. Pliocene records are near the western edge of the Basin and Range Province.

Fish Species Accounts

Keeler South: Owens Lake Clay

Family Catostomidae, Suckers, UMMP

50905-50906.—*Catostomus tahoensis* Gill and Jordan (Tahoe sucker) may be a sister species to *Catostomus fumeiventris*. The fossils consist of a partial Weberian apparatus including a second rib and a complete pharyngeal arch with teeth. The Weberian second rib is narrow in posterior aspect, with a porous median plate (Fig. 2j), unlike the recent species. The pharyngeal arch and teeth (Fig. 2i) are similar to *C. fumeiventris*, especially in the somewhat molariform lower teeth.

Family Cyprinidae, Minnows, UMMP 50907-50908.—

Siphateles bicolor is represented in the clay by several opercles, hyomandibulae, vertebrae, scales, and miscellaneous non-diagnostic fragments. Opercles (Fig. 3) are similar, in general, to recent specimens of *Siphateles bicolor snyderi*, but differ in three features: the dorsal profile is less concave than in *S. b. obesus* of the Lahontan Basin, the anterior edge is slightly concave, and the postero-ventral corner extends moderately posteriorly as in *S. bicolor obesus*. The hyomandibular is like that of *S. b. snyderi*.

Keeler South: Owens Lake Sand

Family Cyprinidae, Minnows, UMMP 50899-50903.—

A form similar to *Siphateles bicolor* is abundantly represented in the sediment sample by one complete and two partial pharyngeal bones and 40 isolated teeth, 25 partial dentaries, three hyomandibulae, eight ceratohyals, two partial urohyals, four partial maxillae, one palatine, nine opercle fragments, three preopercles, four post-temporals, a basioccipital, five frontals, eight parietals, additional skull bones, and Weberian ribs. The pharyn-

geal arches are similar to *S. b. snyderi* (Fig. 4a, b) in the lengths, robustness, and curvature of the anterior and dorsal processes (Fig. 4c, d). The teeth are mostly typical of *Siphateles*, they are laterally compressed, slightly hooked, and with well-developed grinding surfaces, but several are somewhat molariform, like *Myopharodon* (Figs 4e). The dentaries (Fig. 3a) are similar to *Siphateles b. snyderi*, *S. b. obesus*, and *S. mohavensis*, but variable, some resembling *Mylopharodon* and *Lavinia* in the lateral flare and flat anterior of the biting surface. The maxillae fragments are similar to *Siphateles*. Ceratohyals are slightly longer and slenderer than *S. b. snyderi*, resembling *S. mohavensis* and *S. b. pectinifer* from Pyramid Lake. The basioccipital differs slightly from *Siphateles bicolor*, *S. b. snyderi*, and *S. mohavensis*. The post-temporals, frontals, and parietals are variable but with features diagnostic of *S. bicolor*, generally, as well as *S. b. snyderi*. Two cyprinid dermethmoids, one complete (Fig. 3j) are the right size for *Siphateles* (Fig. 3i, k), but wider, especially anteriorly, with anterolateral pits, unlike any North American or Eurasian minnow seen. They are interpreted as divergent *Siphateles*.

Special attention was directed to searching for *Rhinichthys osculus* and *Cyprinodon radiosus*, but no bones of these species were seen, despite the abundance of small elements of appropriate size. The numerous cyprinid teeth are robust and laterally convex, unlike *Rhinichthys* teeth, which are laterally compressed with posterior concave profiles.

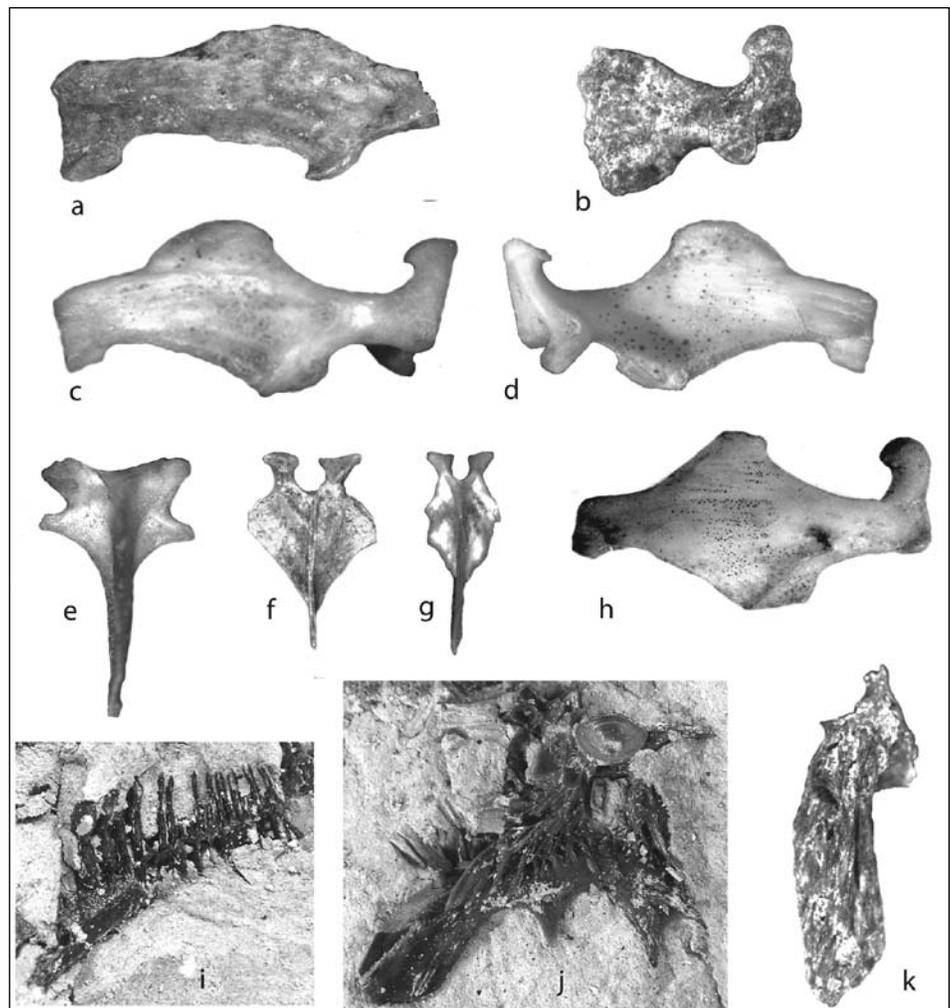


Figure 2. Fossil *Catostomus* spp from Owens Lake Sands and Owens Lake Clay, Owens Valley; *Catostomus fumeiventris*, UMMZ 181667, recent, Owens Valley; and *Catostomus tohoensis*, UMMZ 169022, Lahontan Basin. Width of bone(s) in each view given in millimeters unless otherwise indicated. a *Catostomus* sp., Owens Lake Sands, posterior part of maxilla, lateral view, 12 mm; b *Catostomus* sp., Owens Lake Sands, anterior part of maxilla, mesial view, 7 mm; c *Catostomus tohoensis*, Washoe Co., NV, lateral view of maxilla, 10 mm; d *Catostomus tohoensis*, Washoe Co., NV, mesial view of maxilla, 10 mm; e *Catostomus fumeiventris*, Mono Co., CA, dorsal view of urohyal, 8 mm; f *Catostomus* sp., Owens Lake Sands, dorsal view of urohyal, 4 mm; g *Catostomus tohoensis*, Washoe Co., NV, dorsal view of urohyal, 3 mm; h *Catostomus fumeiventris*, Mono Co., CA, maxilla, 21 mm; i *Catostomus* sp., Owens Lake clay, pharyngeal arch and teeth, 15 mm; j *Catostomus* sp., Owens Lake clay, Weberian apparatus (part), 26 mm; k *Catostomus* sp., Owens Lake Sands, Weberian rib, length 15 mm.

Family Catostomidae, Suckers, UMMP

50895-50898.—*Catostomus fumeiventris* and *C. tohoensis* bones include one left Weberian rib, two partial maxillae, a urohyal, a partial palatine, 44 vertebrae, about 100 scales, and a tooth. The Weberian rib resembles *C. tohoensis* in its elaborate lateral sculpting (Fig. 2k). The maxilla is dorso-ventrally slender with a prominent ventral process for the dorsal maxillary muscle attachment (Fig. 2a), most like *C. tohoensis* (Fig. 2c). The antero-dorsal process emerges at a right angle to the axis of the bone, as in *C. fumeiventris* (Fig. 2h) rather than *C. tohoensis* (Fig. 2d). The

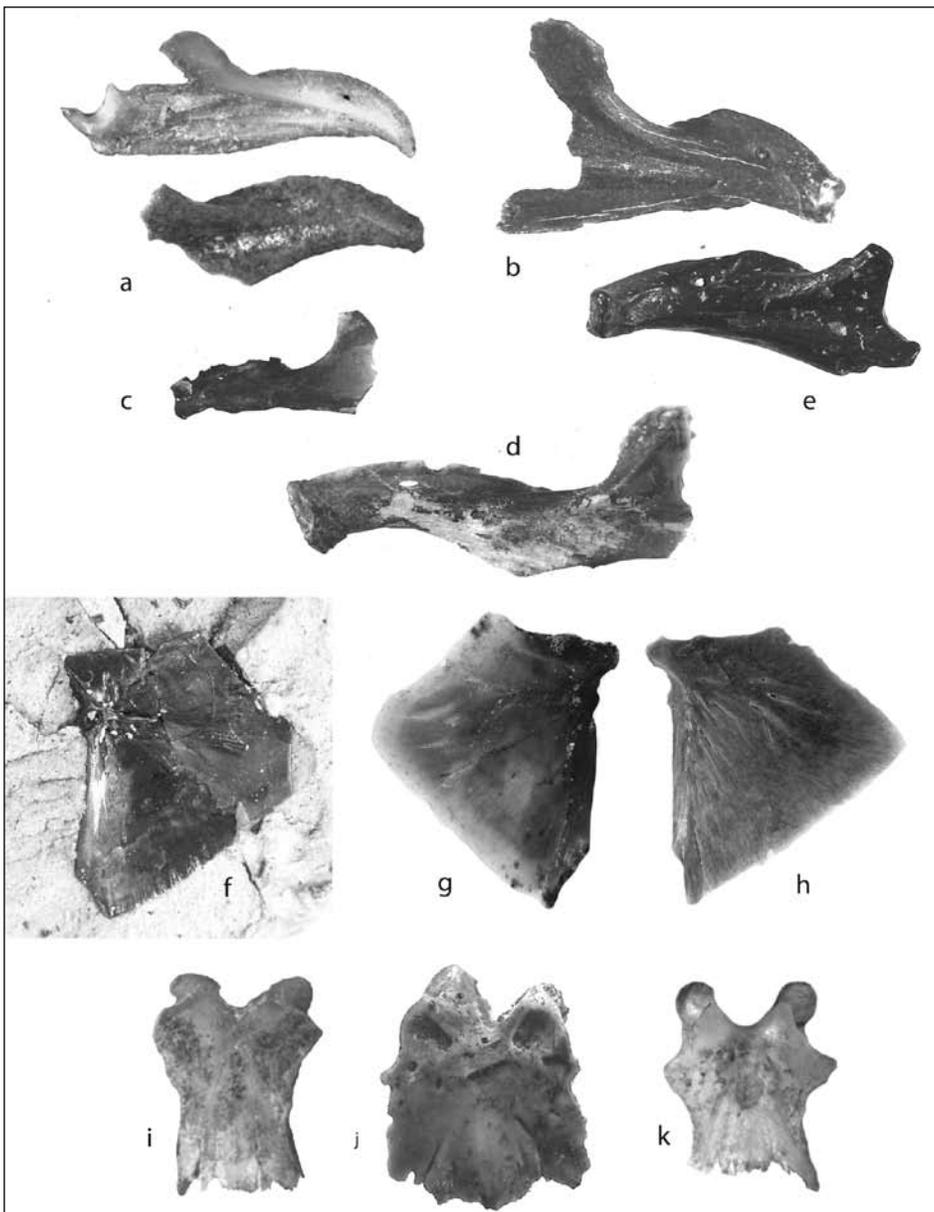


Figure 3. Fossil dentaries, opercles, and mesethmoids of *Siphateles* spp from Owens Lake Sands and Owens Lake clay, Owens Valley, and related drainages; *Siphateles bicolor* UMMZ 188955, Lahontan Basin. Width of bone(s) in each view given in millimeters. a *Siphateles bicolor*, Lahontan Basin left dentary (above), 11 mm; and Owens Lake Sands left dentary (below), 9 mm; b *Siphateles altarcus*, Fort Rock Basin, dentary, 17 mm; c *Siphateles* sp., Lake Russell, juvenile dentary, 9.5 mm; d *Siphateles* sp., Lake Russell, dentary, 28 mm; e *Siphateles* sp., Honey Lake, dentary, 11mm; f *Siphateles* sp., Owens Lake clay, opercle, 12 mm; g *Siphateles* sp., Lake Russell, opercle, 16 mm; h *Siphateles mohavensis*, Manix Lake, opercle, 23 mm; i *Siphateles bicolor*, Lahontan Basin, mesethmoid, 2.4 mm; j *Siphateles* sp., Owens Lake Sands, mesethmoid, 6 mm; k *Siphateles mohavensis*, Manix Lake, mesethmoid, 3 mm.

urohyal (Fig. 2f) is intermediate between *Catostomus fumeiventris* (Fig. 2e) and *Catostomus tahoensis* (Fig. 2g). The tooth is moderately compressed with a slight terminal, blunt point. Vertebrae are not diagnostic at this level. The scales are robust and large, 4-5 mm in diameter (too large for the *Siphateles* in the fauna), with 8-12 posterior radii, no anterior or lateral radii, and the focus anterior to the center, unlike *C.*

fumeiventris or *C. tahoensis*, which have anterior radii and central foci.

Discussion

Fish specimens from the Owens Lake clay show more differentiation from their recent counterparts than is usually observed in Pleistocene fish bones. The well-preserved opercle is clearly a *Siphateles*; its shape is different but with overlapping similarities to other forms of *Siphateles* in the Lahontan Basin and formerly connected drainages (Fig. 3 g, h).

The oldest *Siphateles* are from the Barstovian (ca 15 Ma) Buffalo Canyon beds of Nevada. Because the fossil record of *Siphateles* is nearly continuous (in 1 Ma intervals) since that time, 15 Ma is a reasonable estimate for origin of the genus. Because the genetic distance between *Siphateles* and its sister clade, *Mylopharodon* plus *Lavinia*, is 14.7% (Smith et al., 2002; Estabrook et al. 2007), a rate of about 1%–1.8% per million years is plausible. Modern *Siphateles* in the Owens Basin differ from their counterparts in the Lahontan, Mohave, and Mono Lake Basins by about 3.1% of their cytochrome C mitochondrial DNA sequence divergence

(Smith et al., 2002; Estabrook, 2007; Spencer et al., 2008) suggesting that these lineages split approximately 1.7–3 Ma.

The fossil *Siphateles* bones from Owens Lake sands have plesiomorphic (primitive) characters and are mineralized—replaced with hydroxyl apatite, abraded, probably reworked, and comparable to

relatives seen in Pliocene and early Pleistocene fish in the western U.S., suggesting that these fishes might be Pliocene or early Pleistocene. The molariform teeth (Fig. 4 d, e) are particularly plesiomorphic, similar to Mono Basin (Fig. 4 f, g, h), Carson Valley, Honey Lake (Fig. 4l), and Fort Rock Basin (Fig. 4m), as well as the sister genus *Mylopharodon* of the Pliocene Snake River Plain and Sacramento–San Joaquin Basin, in their blunt shape. (Pliocene *Siphateles* from Lake Russell were confused with *Mylopharodon* and *Lavinia* by Miller and Smith, 1981, based on the molariform teeth. K.W. Gobalet examined and described a similar *Siphateles* from Pliocene sediments in Carson Valley and corrected the error in Smith et al., 2002.) The molariform teeth in the Owens Lake sands are not as large or robust as those from the older lake basins listed above, but are consistent with hypothesized gene flow at the time of Lake Russell spillover into Owens Valley, sometime after 1.3 Ma (mid-Pleistocene; Reheis et al. 2002). Pleistocene *Siphateles* from China Lake (Fig. 4 j, k) and Lake Manix (Fig. 4 n, o) have slender teeth with grinding surfaces as in modern *S. bicolor* (Fig. 4 a, b) and *S. mohavensis*. Some individuals of *Siphateles bicolor pectinifer* of Pyramid Lake gain molariform replacement teeth when they grow up to about 250 mm, but in the Pliocene forms reported here, blunt teeth developed at smaller growth stages.

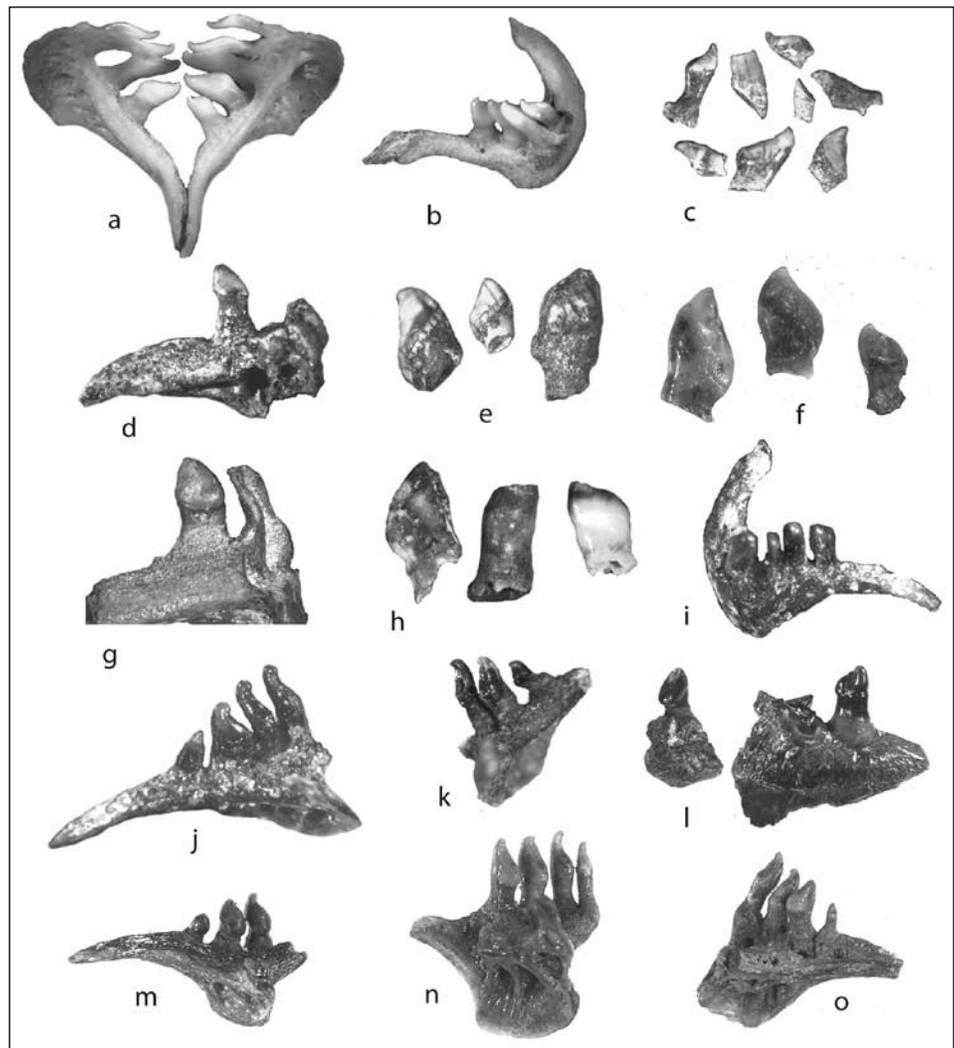


Figure 4. Fossil pharyngeal teeth and arches of *Siphateles* spp. from Owens Lake Sands, Owens Valley, Inyo Co., CA; Lake Russell beds, Mono Basin, Mono Co, CA; China Lake beds, Inyo Co., CA; Honey Lake beds, Lassen Co, CA; Fossil Lake beds, Fort Rock Basin, Lake Co., OR; and Manix Lake beds, San Bernardino Co., CA; *Siphateles bicolor*, UMMZ 188955, Lahontan Basin, Churchill Co., NV. Width of bone or lengths of teeth given in millimeters. a *Siphateles bicolor*, Lahontan Basin, paired pharyngeal arches with teeth, antero-dorsal view, 13 mm; b *Siphateles bicolor*, Lahontan Basin, pharyngeal arch shape, mesial view, 10 mm; c *Siphateles* sp., Owens Lake Sands, eight isolated teeth, 1.5–3 mm long; d *Siphateles* sp., Owens Lake Sands, partial pharyngeal arch and tooth, 9 mm; e *Siphateles* sp., Owens Lake Sands, three slightly blunt teeth, 4, 2.5, 5 mm long; f *Siphateles*, Lake Russell, three isolated pharyngeal teeth 4, 3.5, 3.5 mm long; g *Siphateles*, Lake Russell, partial pharyngeal arch with 1 ½ teeth, 13 mm; h *Siphateles*, Lake Russell, three isolated pharyngeal teeth 6, 5, 4 mm long; i *Siphateles* sp., China Lake, pharyngeal arch and teeth, 10 mm, see j for shapes of teeth; j *Siphateles* sp., China Lake, pharyngeal teeth 10 mm, see i for shape of arch of this specimen; k *Siphateles* sp., China Lake, pharyngeal arch and teeth, juvenile, 4 mm. l *Siphateles* sp., Honey Lake, two teeth, 6 mm, and 4.5 mm long (omitting arch fragments); m *Siphateles altarcus*, Fort Rock Basin, pharyngeal arch and three of five teeth, 17 mm. n *Siphateles mohavensis*, Manix Lake, partial pharyngeal arch and teeth, 8 mm wide; o *Siphateles mohavensis*, Manix Lake, partial pharyngeal arch and teeth Lake, 12 mm wide.

Catostomus fumeiventris, the Owens sucker, and *C. tahoensis* of the Lahontan Basin are distinct morphologically (Miller, 1973). The Owens Valley *Catostomus* fossils from the Owens Lake sands share traits with both species, or are intermediate. Surprisingly, *C. fumeiventris* shares mtDNA (at 5.3% sequence

divergence, Thomas Dowling, unpublished data) with *Chasmistes cujus* of Pyramid Lake, indicating hybridization about 5 Ma between these forms.

It is interesting that no *Rhinichthys* or *Cyprinodon* fossils have been identified from the Owens Valley samples, despite the presence of many bones and teeth in their size range. The *Rhinichthys osculus* in the Mono and Owens Basins and *Cyprinodon radiosus* from the Owens Basin are relatively old lineages, according to their mitochondrial DNA. However, new interpretations of the effect of body size and temperature on mtDNA rates in fishes require re-evaluation of the conclusions of Smith et al. (2002) and Echelle (2002). Mitochondrial DNAs of small fishes and warm-water fishes, such as those in the Death Valley–Amargosa River waters, evolve much more rapidly—as fast as 3% per million years (Estabrook et al., 2007; Smith and Dowling, 2008)—compared to mtDNA in larger, coldwater fishes. These results, coupled with the 3–3.5% sequence divergence in the Owens River populations of the small pupfish and dace (Smith et al., 2002) suggest that they could have colonized as recently as ca 1 Ma, as argued by Knott et al. (2008). The four Owens River fish species and their fossils do not provide evidence for through-flowing river segments to the lower Colorado River, except for the Owens pupfish and its connection to Death Valley. The other five species have connections to the north. The connections that enabled pupfish to disperse between the Owens River and Death Valley waters from the Lower Colorado River remain enigmatic.

The speckled dace, *Rhinichthys osculus*, clearly colonized the Owens Basin from the East Walker River, which was in turn the source of populations in the Amargosa River (Smith and Dowling, 2008, Chow et al., in revision) according to mtDNA haplotypes in the Owens, Mono, Amargosa, and East Walker basins. Long-term, though sporadic, connections between the Mono Basin and East Walker River in the late Pliocene and early Pleistocene were documented by Reheis et al. (2002). Genetic distances indicate that the split between the Owens River and Amargosa River populations date from the early Pleistocene (1.7 Ma).

Various basins along the east side of the Sierra Nevada were occupied by isolated fish populations in a north–south system (Fig. 1) for 104- to 106-year periods during the Late Pliocene and Pleistocene, as inferred from divergent fish populations, especially *Siphateles altarcus*, *Chasmistes* spp., and *Catostomus* spp. (Miller and Smith, 1981; Smith et al., 2002; and Reheis et al., 2002). *Siphateles altarcus* and other

blunt-toothed tui chub populations of the Pliocene Owens, Mono, Honey Lake basins diverged but went extinct as they gave way to Pleistocene-to-recent sharp-toothed tui chubs, *Siphateles bicolor* subspecies and *S. mohavensis*, of the Lahontan, Mohave, Klamath, and related basins. Occasional Pleistocene and Quaternary gene flow is indicated by the morphologically homogenized recent representatives of the occasionally connected basins.

Acknowledgments

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Competing mesquite and creosote dune systems near Stovepipe Wells, Death Valley National Park

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Abstract: Historical records of turnover of desert vegetation have focused on processes such as invasive expansion of shrubs (e.g. mesquite (*Prosopis glandulosa*) or creosote (*Larrea tridentata*)) on sandy desert soils, pointing to human disturbance or fire as major drivers (e.g. Bahre and Shelton, 1993). The underlying presumption in previous reports is that the pre-disturbance plant distribution was stable, and the disturbance (i.e. grazing, ground water extraction, fire) shifts the balance to favor one species. Observations of a collapsing portion of a mesquite nebkha field juxtaposed with an expanding creosote dune complex suggest an alternative pathway for vegetative turnover based on slow, possibly oscillatory shifts in edaphic controls. Preliminary field observations are presented, and a working hypothesis of a natural, possibly periodic process is proposed as a stimulus for further research.

Introduction

This note describes observations of portions of a mesquite nebkha field being replaced by a creosote dune system due to degradation of nebkhas by surface water flow. The rate of this process is unknown, but very slow, as all individual shrubs in degraded nebkhas are observed to have remained in place over at least

a dozen years. Dynamic change is inferred from the distribution of dune, nebkha, and channel features.

This process may be cyclic, assuming sand replenishment balances fluvial removal over long periods, leading to eventual nebkha stabilization and restora-

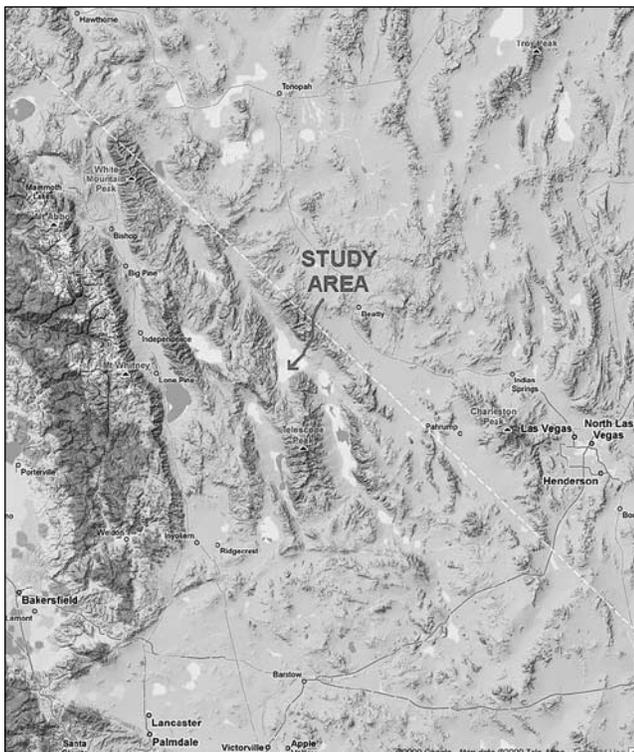


Figure 1. Location Map showing Death Valley and the study area within the Mojave Desert region (source: GoogleMaps).

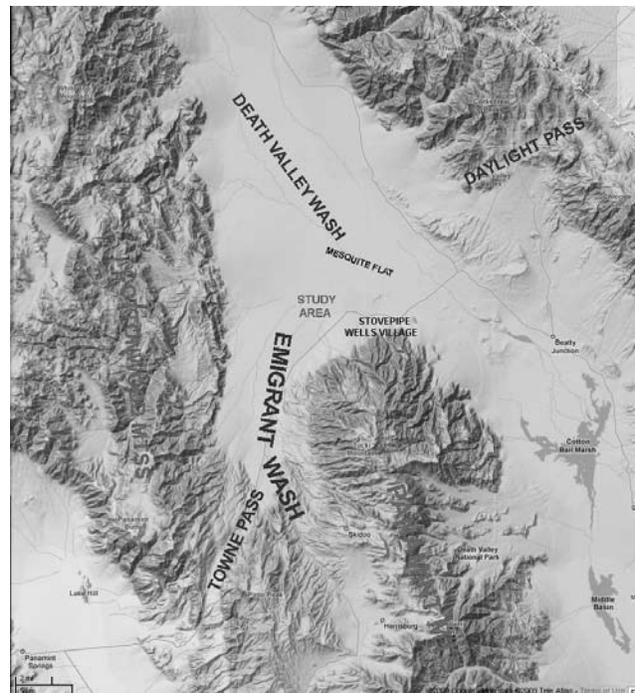


Figure 2. Topography and major drainages around Mesquite Flat. The large drainage areas of Emigrant Wash, Towne Pass, Cottonwood Mts. and Death Valley Wash capture rainfall from elevated areas with much higher precipitation than the valley floor. Winds are channeled by the mountains, with low passes on both sides of the Valley (source: GoogleMaps).

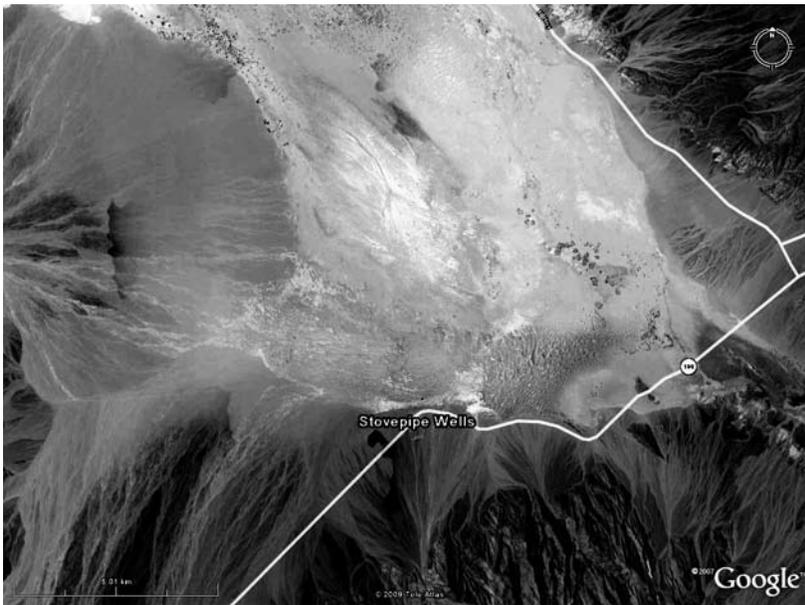


Figure 3. Color and contrast enhanced image of Mesquite Flat Dunes. The terminal playa of Death Valley Wash is at the center; Cottonwood and Emigrant washes enter from west and southwest, respectively. Tall dunes are NE of Stovepipe Wells; study area is NW of Stovepipe Wells. Scale bar is 5 km (source: GoogleEarth).



Figure 4. Eastern portion of Mesquite Flat dunes, viewed from the south looking toward the largest dunes. Note stable mesquite nebkhas in middle distance and creosote in foreground.



tion of the mesquite system, or it may represent a directional change in the whole system modulated by a change in sediment supply to the upwind dune area or increased rates of fluvial erosion along the flanks of the dunes. Resolving this will require further research.

Study Area

The Mesquite Flat sand dunes are located in the northern portion of Death Valley proper (Figures 1 and 2), at the junction of the north–south trending Death Valley trough with the southwest–northeast trending gap between the Panamint and Cottonwood Mountains. Abundant surface sediment is available to accumulate into dune formations because major drainages from the north and both sides of Death Valley converge here, and winds are topographically focused to collect this sediment into one of the most visited and photographed areas of Death Valley National Park.

The Mesquite Flat dunes consist of two sand masses. The frequently visited and photographed portion lying northeast of Stovepipe Wells (throughout this paper, this name denotes the present resort development, not the historical well site) is a shallow (1–3 m) sand sheet with scattered mesquite nebkhas and creosote shrubs grading into a compact core deep sand area of bare dunes tens of

Figure 5. Vegetation and surface features of the study area. Stable nebkhas are visible on both sides of the image. Flow from Emigrant wash enters the dune area from the southwest, crossing Cottonwood Rd. Flow from Cottonwood Canyon enters from the west along the bright channels at the edge of the dunes. Scale bar is 1100 m (source: GoogleEarth).



Figure 6. Partially exhumed mesquite nebkha with new active dune approaching from the northeast. Note the complete absence of surface sand to the right of the nebkha. Partial surface in front of the nebkha is a weakly cemented mudstone deposited by flooding in an interdune space (note concave shape), with confining dunes subsequently removed. Shrubs on the dune and in distance are creosote.



Figure 7. Degraded nebkha with dead vegetation and exposed roots. In background are old flood deposits and a cobble-strewn surface about 2-3 m lower than foreground surface.



Figure 8. Degraded nebkhas in the channel area on the western edge of the dunes. Exposed deadwood is visible on top of the remnants. Living shrubs in image are all creosote. This nebkha's demise predates the 1993 aerial photos.

meters high (Figure 5). Northwest of Stovepipe Wells is a lower sand sheet and deep sand complex consisting mostly of transverse dunes with scattered creosote, with the transverse dunes migrating southwestward from the terminal playa of Death Valley Wash. There is a substantial area of large, stable mesquite nebkhas on the western end of this sand body, and scattered nebkhas along the southwest and eastern portions.

The local drainage base level is the playa, which extends under the sand nearly to Stovepipe Wells (light areas north of Stovepipe Wells in Figures 3 and 5). The playa and its catchment are the presumed sources of shallow ground water that support the mesquite nebkhas.

Observations

The vegetation of the study area is shown in an annotated aerial image in Figure 5. Large, stable nebkhas are visible on the right and left portions of the image. Smaller shrubs, almost entirely creosote, are visible on the alluvial flats and active transverse dune areas.

Nebkha collapse is occurring where anastomosing channels and sheetflow enter the dunes,

Figure 9. Sheetwash area near the western edge of the dunes. Mudstone deposits and water-lain gravel with ripple marks are accumulating drift sand in low spots and in the lee of the mud deposit. Note deadwood cores of completely eroded mesquite nebkhas in background.



cutting away the sides of the nebkhas and exposing the roots of the mesquite plants. Figure 6 shows a partially exhumed nebkha surrounded by evidence of flood erosion, including a concave mud deposit such as would form by flooding between dunes, but the bordering dunes have been subsequently removed (see Langford, 1989). In the background the advancing transverse dune field is visible.

Figures 6–11 show the evidence of repeated flood activity along the interface between the confluence of Cottonwood and Emigrant Washes and the dune area. The landforms shown here are similar to those previously described for dune-flood interface areas along the Mojave River (Langford, 1989).

Comparative imagery from 1993 and 2005 (Figure 12) shows the recent reworking of the Cottonwood Wash channel system, presumably in the floods of August, 2004, and Figure 13 shows the 1993–2005 changes in fresh surface sediment at the low point of

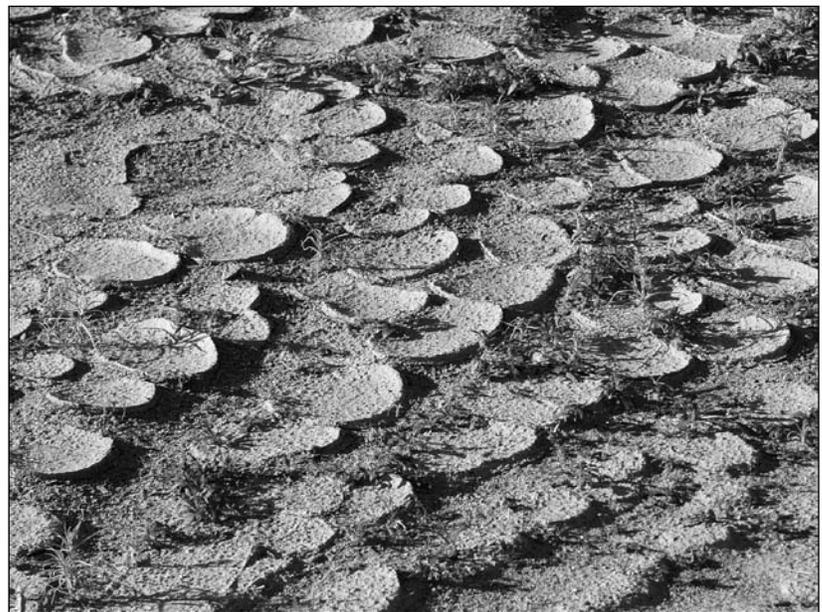


Figure 10. Relatively recent mud cracks in area between degraded nebkhas.



Figure 11. Old flood deposits underlying a partially degraded nebkha.

the system, near the Stovepipe Wells airport.

Figure 14 shows the sub-linear array of transverse dunes that, if it continues to advance, will replenish the flood-eroded area with fresh sand. These dunes are also visible in the background of Figure 6.

Discussion

Initial field observations were made in March, 2005. They were preceded by at least two events of unusually strong precipitation. Very strong monsoonal thundershowers occurred in the Death Valley region on August 15, 2004. The valley floor was estimated to have received about 5 cm of precipitation (equivalent to a typical yearly total) in a period of less than one hour (NWS 2004). Precipitation totals in the

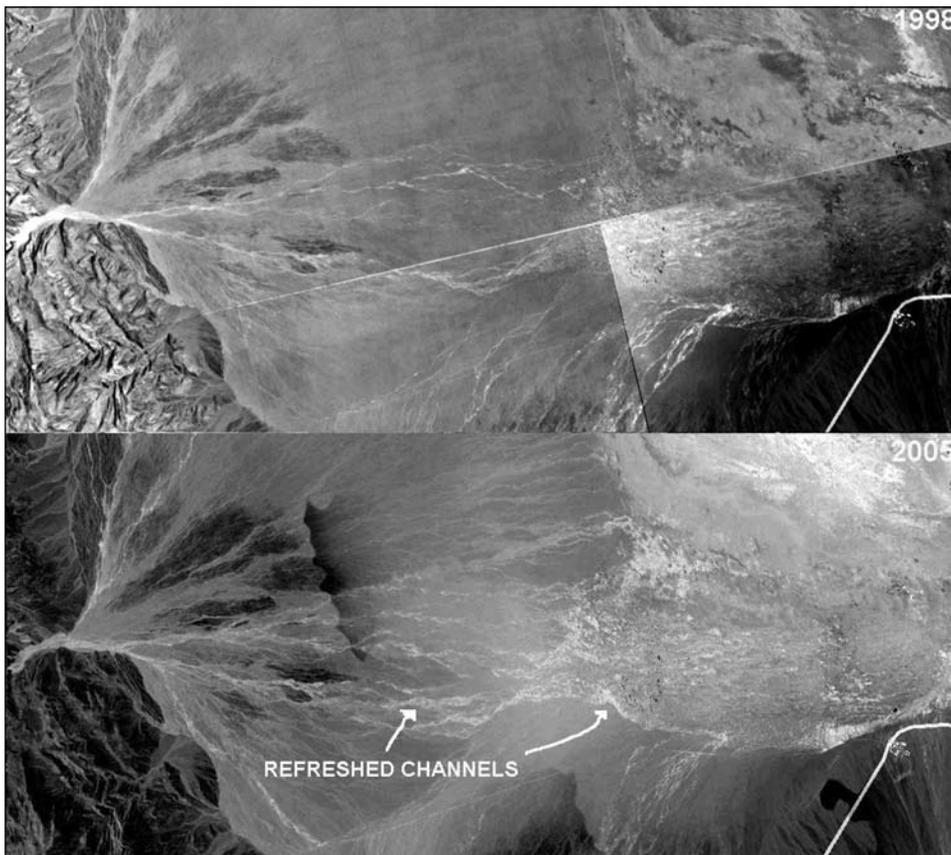


Figure 12. Comparison of Cottonwood Wash channel patterns between 1993 and 2005. Refreshed and expanded channels are visible along the lower edge of the dune area.

surrounding mountains are not known. Most of the news reports of this event focused on flooding along Furnace Creek and Mud Canyon, and the degree of surface flow into the northwestern flanks of the Mesquite Flat dunes is unknown, even qualitatively. The winter 2004–5 El Niño also contributed significant, unusual precipitation to the Death Valley region (16.5 cm, or about 3 times average annual supply) (DVNM, 2007). Either or both of these events may be responsible for the relatively fresh channel reworking and fresh mud crack surfaces observed the following March (Figure 10).

Evidence of previous channel activity is abundant in the area, undercutting mesquite nebkhas as large as 3 m high, leaving the roots of the plants exposed, (presumably) causing the observed above ground dieback, and removing sand from the spaces

between them. Obviously a large volume of sand has been removed in this area, but comparison of aerial photos from 1998 to 2005 show no loss of large plants, thus the collapsed nebkhas (e.g. Figure 8) were more than 12 years old at the time of the photos, and the degraded ones had thus survived a considerable period with exposed roots.

Evidence that this may be a cyclical process is drawn from the observation that some of the nebkhas are perched on old flood deposits (Figure 11), suggesting that the nebkha field can be reconstituted after flood removal. Again, no time frame is suggested.

The novel aspect of these observations is that the same system that provides the shallow ground water on which the nebkhas depend can destroy them with high surface flows in extreme events. Based on

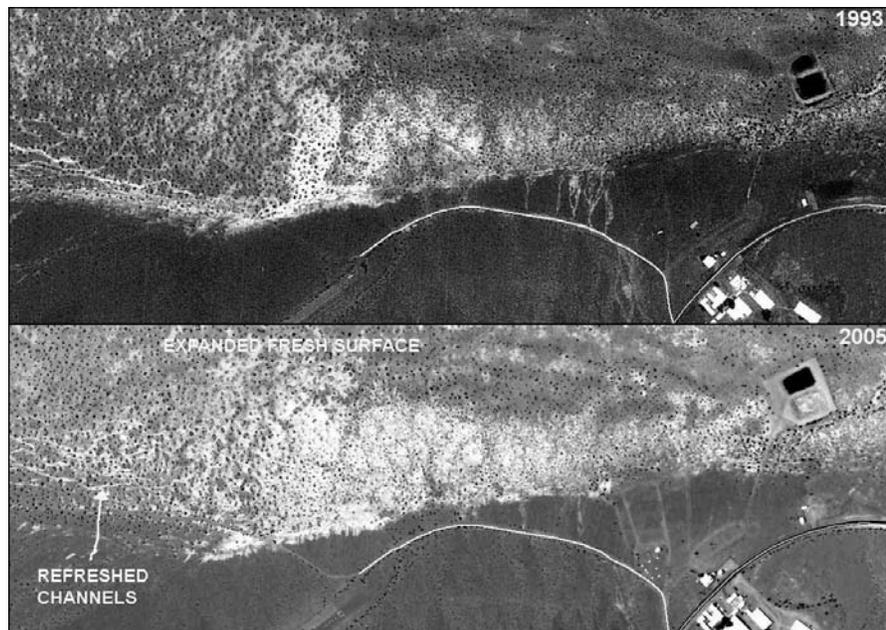


Figure 13. Active surface sediment accumulation at lower end of Cottonwood/Emigrant channels is evident as a lighter surface color. The growth of the new surface area is evident in comparing 1993 to 2005.



Figure 14. View looking southeast along the sub-parallel fronts of the transverse dunes advancing southwestward toward the channel area at the confluence of Cottonwood and Emigrant washes.

subjective experience of Death Valley residents, such large flash floods have a recurrence frequency of about 20 to 40 years (NWS 2004). How many floods are needed to remove the dunes, and how long the cycles of reestablishment may last are subjects for future research.

A brief literature review suggests that this phenomenon has not been previously reported, but this search is not definitive, and the author would appreciate any communication on this topic. A very short review of some relevant literature is provided here for context.

Vegetation

Observations of changing vegetation in the U.S. desert southwest have been reported since the early 20th century. Much of the original interest was driven by the devastation of rangeland from Texas to Arizona that followed the cattle boom of the 1870s and the decade of drought beginning in 1884 and the subsequent realization, with later droughts (not least the 1930s), that drought is a recurring problem in this region (Bark, 1978). Systematic range evaluations date to the middle of the last century, rooted in landmark papers discussing the balances between grass and shrub dominance and invasive spread of woody species (*e.g.* Brown, 1950, Bogush, 1952, and Branscomb, 1958). Over the next 40+ years, a general

understanding evolved that the relative distributions of grass and shrubs are strongly influenced by fire and grazing, and that the relative distribution of woody shrubs such as mesquite (*Prosopis* spp.), and creosote (*Larrea*) is driven by soil texture and soil water availability, with some (true xerophytes such as creosote) succeeding in drier sites or on thin, rocky soils, and others (especially phreatophytes such as mesquite) succeeding on sandy deep soils or where consistent soil moisture could be exploited by deeper root systems (Nilsen *et al.*, 1983; Bahre and Shelton, 1993).

Across the deserts of northern Mexico and the southwestern U.S., creosote and mesquite coexist on desert flats in the Chihuahuan and Sonoran Deserts, where summer rainfall from the North American monsoon occurs reliably. On the western edges of this region, from Baja California northward through the Mojave Desert, summer rains are unreliable and creosote dominates the open areas, while mesquite is restricted to riparian habitats and nebkhas which can trap and retain the sparse summer moisture. In many sites, nebkhas develop over surfaces with shallow ground water or over relatively impermeable layers like playa surfaces or clayey soils which facilitate moisture retention in the sand body (Langford, 2000; Khalaf *et al.*, 1995). The remarkable efficiency of mesquite nebkhas to capture and hold moisture is discussed by Parsons *et al.* (2003).

There is limited literature on mesquite–creosote dynamics in the Mojave, owing to the general separation of the populations. This separation is particularly well demonstrated by a study by Laity (2003) that tracked the replacement of mesquite by creosote along the Mojave River as a result of upstream ground water extraction (Laity, 2003).

In marginal grass–desert areas, nebkhas have been described as developing as a consequence of overgrazing or transient cultivation. An example of this process is evident in the agricultural areas of the

Antelope Valley west of Lancaster (personal observation). The conceptual issues of nebkha development in New Mexico have been presented by Kocurek and Nielson (1986), and the difficulty of identifying a single controlling process is made clear by a paper describing nebkhas in the deserts of China (Wang *et al.*, 2006).

Dunes

The dynamics of dunes and their fluvial sediment sources is a well-established field, with extensive literature on sediment movement and dune form evolution, a comprehensive discussion of which is beyond the scope of this paper. The general relationship for topographically enclosed areas is that movement of sediment by water exposes it in dry streambeds or along playa shores, whence it is transported by local winds to accumulate in dune fields which are topographically confined. Precipitation (climate) is frequently cited as a controlling factor, with increasing or decreasing sediment supply being reflected in changing dune structure. A good example of this is presented by Lancaster (1997) in his discussion of the dynamics of the Devil's Playground dune field near Kelso, CA, later expanded into a generalized theory of dune systems (Kocurek and Lancaster, 1999).

Research on the relationship between dunes and watercourses is more limited, with the most relevant work in regard to the interaction between flash flooding and dunes, such as discussed here, to be found in Langford's (1989) examination of perfluvial landforms and sedimentary facies in the Great Sand Dunes in Colorado and along the Mojave River. That paper is, however, reporting on overbank flooding due to seasonal or short period (about 2 years) pulses of water into the surrounding dunes, and thus does not address the long recurrence period or the high velocity nature of the flooding that occurred in 2004 (and presumably repeated on decade+ time scales) in Death Valley. In addition, it is silent about vegetation, focusing instead on developing modern analogs for fossil dune systems such as the Nebraska Sand Hills.

Conclusions

The landforms and plant life discussed here show that vegetative patterns in some dune systems are dynamic on periodic time scales of uncertain length (probably decades to centuries). The drivers of these changes may be independent of the direction of change (or stability) imposed by seasonal or inter-annual processes, or by gradual climate change. Such dynamics

deserve consideration in evaluating historical photographs, tracking long term vegetation plots, and in planning restoration of disturbed desert environments.

The evidence presented here is limited, and obtained more by luck than premeditated field work. It is likely that more extensive observations covering more of the dune fields could answer some of the outstanding questions noted here.

The author encourages investigators with greater resources to follow up on these observations, especially tracking the rebound of the dune field over ensuing years, or finding analogs in other stages of this (hypothesized) cycle.

Acknowledgement

The author owes a debt of gratitude to an anonymous reviewer, whose questions on the original draft led to a reexamination of the field evidence and a thorough rethinking of this presentation. This is how peer review is supposed to work, and it has been a gratifying experience to have the benefit of this reviewer's efforts.

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The Oak Creek mudflow of July 12, 2008, Independence, Inyo County, California: preliminary report

Dave Wagner, *California Geological Survey*; Margie DeRose, *Bureau of Land Management and U.S. Forest Service*; and Jim Stroh, *The Evergreen State College, Emeritus*

Introduction

On July 12, 2008, tropical moisture moved eastward from the Pacific Ocean across southern California causing thunderstorms and local pockets of intense rain. In the late afternoon, a cell of intense rain was centered on the Oak Creek drainage just north of Independence in Inyo County, California. This area had been burned in an intense, lightning-sparked range fire in July 2007. Around 5 PM, mud and debris flows began moving down tributaries in the Oak Creek drainage area, eventually concentrating in the north and south forks of Oak Creek.

At about 5:30 PM a hyperconcentrated mudflow destroyed a motor home parked at the Oak Creek campground on the north fork of Oak Creek. Its sole occupant was swept downstream in the mud but he managed to survive. By then, mud and boulders had severely damaged the Mt. Whitney Fish Hatchery and several Fish and Game employee residences. Farther downstream along Oak Creek, 25 homes were destroyed and another 25 were damaged to some extent. Relatively few people were home and amazingly there were no serious injuries or fatalities among the residents. At 5:30 PM, the Independence Volunteer Fire Department received reports of flooding and propane leaks and responded along with Cal Fire personnel to



Photo 1. Mud along north fork of Oak Creek. Mt. Whitney Fish Hatchery is in the background. Photo by Ken Babione.

find mud flowing across Highway 395. Mud flows and or debris flows also moved down the south fork of Oak Creek destroying the Bright Ranch. Fortunately, no one was there at the time. Mud and boulders from the south fork merged with mud from the north fork near the fish hatchery and filled the channel of Oak Creek, forcing it to incise a new channel south of the old one.

The south fork of Oak Creek changed to a more northerly channel west of the fish hatchery. Mud flowed east of Highway 395 through the Ft. Independence Indian Reservation eastward, almost reaching the Los Angeles Aqueduct. Highway 395 was closed during the evening and night of July 12 and was subject to restricted travel for over a week.

Extent of the Mudflow and Debris Flow

Rilling of slopes and scouring of stream channels extend up to 3000 m elevation on the escarpment of the Sierra. From the high country, mud moved through the Oak Creek drainage onto alluvial fans and out into Owens Valley, a distance of over 16 km with 1800 m of relief (Figure 1). Slopes are veneered with



Photo 2. Remains of the main house at the Bright Ranch on the south fork of Oak Creek. Photo by Dave Wagner.

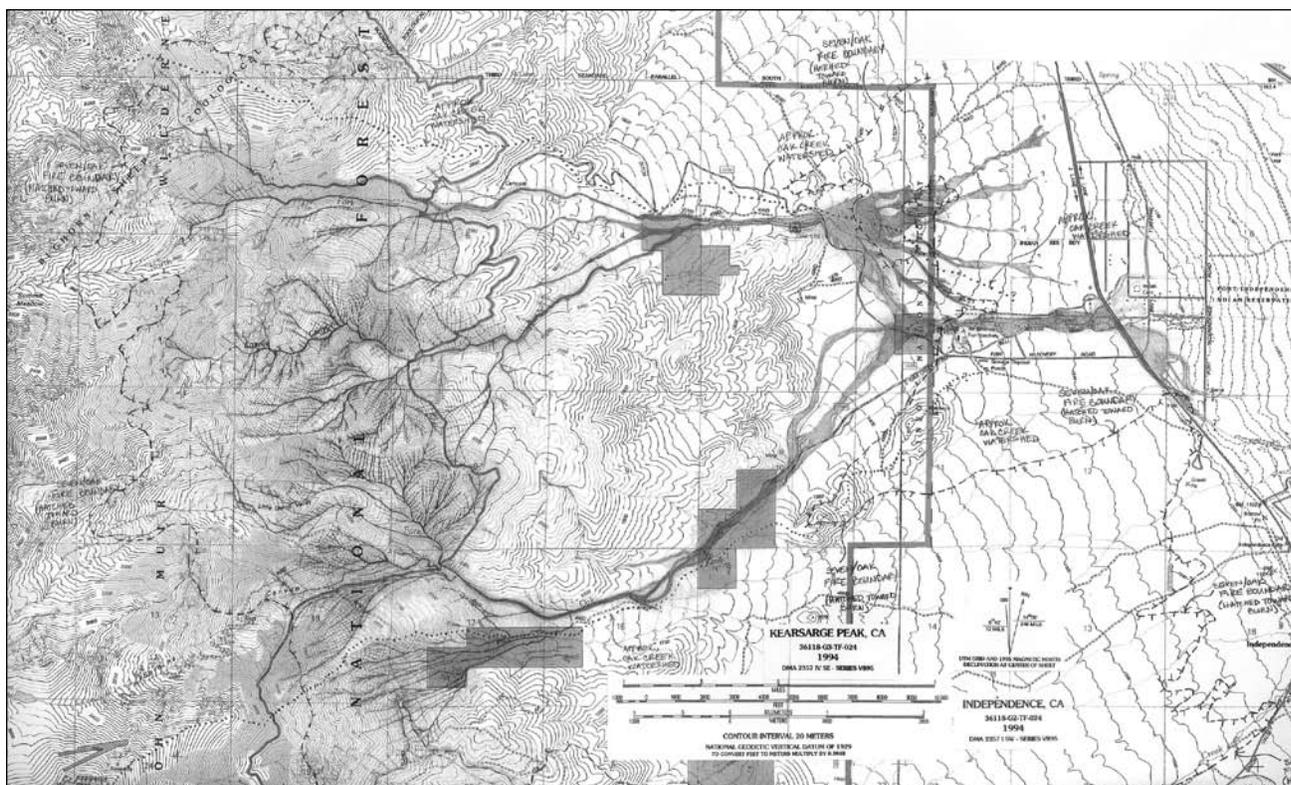


Figure 1. Map of the Oak Creek mudflow of July 12, 2008. Mapping by Margie DeRose, 2008.



Photo 3. Debris flow from north fork of Oak Creek. The Mt. Whitney Fish Hatchery is at the left side of the photo. Mud from the south and north forks of Oak Creek merged just upstream from the hatchery. Photo by Ken Babione.

aeolian sand, decomposed granite, and ash from last year's fire and the channels are filled with sandy, bouldery sediment that is both fluvial and alluvial. Source areas for the south fork of Oak Creek are underlain by glacial till so mud flowing down this part of the drainage contained more and bigger boulders than mud in the north fork.

Mud moved down the narrow canyons for the most part as a hyperconcentrated flow. Mud flowing down the north fork of Oak Creek overflowed below the Oak Creek campground where the channel turns

sharply south and spread across the alluvial fan as a debris flow carrying boulders, logs, and trees. At first, the debris spread across the fan but soon followed existing channels, marked by boulder levees. These channels rapidly filled and the debris flow spread across interflues. The debris flow is nearly a half kilometer across at its widest point and almost 2 km long, nearly reaching Highway 395.

Mud continued to flow in the channel of the north fork of Oak Creek which turns east and heads for the Mt. Whitney Fish Hatchery. Just upstream from the hatchery mud from the south and north forks merged and headed east through the hatchery and destroying homes. Mud then crossed Highway 395, passed through the Ft. Independence Reservation and came to rest in the flat lands west of the Los Angeles Aqueduct.

The man who surfed the mudflow

At about 5:30 on July 12, 2008, Don Rockwood was preparing dinner in his motor home parked at the Oak Creek campground when he noticed the heavy rain. As he told himself, "we need the rain, but this is a little scary," a tree carried by a three-foot wave of mud crashed through the wall of his motor home. Moments later, an eight-foot wave of mud tore the motor home apart and Rockwood was swept away.



Photo 4. Don Rockwood, who surfed over a mile in the mudflow. Photo by Mike Gervais.

An accomplished body surfer, he began ride the mudflow, moving at what he thought was about 30 miles per hour. He surfed for nearly a mile when he saw his pickup truck, also swept away, bearing down on him from behind. He was able to steer sideways, avoiding the truck, and managed to reach shallower and slower mud where he was able to stand. A third wave, about three feet high, swept him away again, submerging him and smashing him against a rock. Now, suffocating, he thought the end was near but the mud slowed and he was able to stand again, making his way to the Fish Hatchery Road. Most of his clothing had been torn away and his shoes were gone. He knew there would be people at the fish hatchery so he began to walk along the road but soon it became impassable due to flowing mud, probably from the south fork of Oak Creek. Lightning began to strike nearby and he was shivering so he looked for shelter but there was none. Rockwood decided he must make it to the hatchery so he picked up a branch to use as a pole and headed through the mudflow toward the hatchery



Photo 5. Remains of Rockwood's motor home in a tree below the Oak Creek campground. Photo by Dave Wagner.



Photo 6. Rockwood's pickup truck mired in the mud after following him down the flow. Photo by Dave Wagner.

and the flashing lights of a fire engine. Mud was up to his chest but he made it through, emerging caked with mud, astonishing a group of fire fighters. He was quickly airlifted to the hospital and treated for major lacerations but otherwise he was ok.



Photo 7. Rilling of slopes in Charlie Canyon. Photo by Jerry DeGraff.



Photo 8. Deeply incised channel of the middle fork of Oak Creek. Note run up of mud above the new channel. Photo by Margie DeRose.

Observations by Don “Rock” Rockwood:

- First “wave” was about 3 feet high; hit his motor home at about 5:30 PM
- Second “wave,” about 8 feet high, hit moments later, destroying the motor home and sweeping him into the mud; he body surfed about a mile until it slowed enough for him to stand
- Third “wave,” about 3 feet high, knocked him off his feet; he surfed about another 0.5 mile before he was able to stand.

After the three “waves” there was a steady flow of mud. Rain had stopped but there were lightning strikes nearby.



Photo 10. Home destroyed along Oak Creek. Photo by Dave Wagner.



Photo 9. Mud from both north and south forks of Oak Creek merge just west of the Mt. Whitney Fish Hatchery. Photo by Ken Babione.

North fork of Oak Creek

The largest single source area for the mudflow is Charlie Canyon, drained by an unnamed tributary to the north fork of Oak Creek, here informally referred to as the middle fork of Oak Creek. Extensive rilling occurred on the slopes in Charlie Canyon, at elevations mostly between 2000 and 2800 m. Debris flow scars were noted in many colluvial swales at the heads of tributaries in the drainage. Deep incision and scouring in the middle fork is apparent above 2200 m elevation and becomes more pronounced downstream. Below 2000 m elevation incision is typically 5 to 15 m; usually bedrock is exposed at the thalweg. Headward and sideward erosion in the form of slumping is ongoing. Deposits exposed in the walls of the new channel are both fluvial boulder gravel and alluvial debris flows attesting to a history of events such as this one. Several other smaller tributaries show similar features but Charlie Canyon is by far the most significant. The north fork of Oak Creek above the confluence with the middle fork appears to be a relatively minor source area.



Photo 11. Mud flowing across Highway 395 requiring CHP escorts to guide traffic. Photo by Ken Babione.



Photo 12. Remnant of the old Parker Ranch buildings along south Oak Creek. Note deep incision of new channel. This partial structure is all that remains of ranch buildings present prior to the mud flow. See also Photo 2. Photo by Jim Stroh.



Photo 13. Double-wide mobile home moved to the right (east) off its foundation and rammed into trees. Trailer was salvaged. Photo by Jim Stroh.



Photo 14. This house took a direct hit by mud and rocks. Note large boulders and tree visible in interior. Flow went in the west side and out the east. View south. Creek on north side of house after partial realignment, but not in original channel. Photo by Jim Stroh.



Photo 15. This cabin on Oak Creek was damaged by mud, but cleaned and made habitable. Photo by Jim Stroh.



Photo 16. This house was damaged by inundation and the foundation undermined after partial realignment of Oak Creek. Photo by Jim Stroh.

The channel of the middle fork of Oak Creek becomes deeply incised after its emergence from Charlie Canyon. In some places the incision is up to 20 m. Mud overflowed the new channel at most of the turns suggesting it was flowing very rapidly.

South fork of Oak Creek

The south fork of Oak Creek has a much different channel morphology than the north fork. It is much wider, as much as 300 to 400 m in places. It had three major sources areas that collectively rival the size of Charlie Canyon. Extensive rilling occurred on the slopes below Onion Valley, Sardine Canyon, and above Tub Springs. As in Charlie Canyon, most of the rilling occurred between 2000 and 2800 m elevation. This area is underlain by boulder-rich glacial till, so the mud that moved down the south fork carried more and bigger boulders than in the north fork. In general, the south fork drainage is far more bouldery

than the north fork because of the abundant large boulders in the glacial till in the source area. Most of the rilled slopes in the south fork drainage are on gruss or aeolian sand, but there is abundant till in the valleys. Mud from the south fork did not spread across a fan like the north fork but stayed in existing channels until it merged with mud from the north fork just upstream from the fish hatchery (Figure 1 and photo to the right).

Damage to homes and Highway 395

Twenty-five homes along Oak Creek just east of the Mt. Whitney Fish Hatchery were destroyed. They were situated very close to the creek and were battered by mud, boulders, and logs.

Mud and debris filled the Oak Creek channel and a new channel was established to the south. The creek was diverted back to its original channel soon after the event by CALTRANS, allowing residents to reach their homes. Mud flowed across Highway 395 and through the Ft. Independence Indian Reservation damaging another 25 homes. Mud and water flowed across the highway for days requiring traffic detours or escorts to pass through the area.

Acknowledgments

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Proceedings of the 2009 Desert Symposium: abstracts and short papers

Evidence that the Big Pine fault is not a segment of the Owens Valley fault and its significance for northern Owens Valley active tectonics

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The Big Pine fault extends 20 kilometers from its southern terminus at the Poverty Hills to its northern terminus a few kilometers north of the town of Big Pine. Although not connected to the 100 km long Owens Valley fault to the south, geologists have long considered the Big Pine fault to be a segment of the Owens Valley fault. New analysis of the Big Pine fault, however, suggests that it might better be considered as a separate fault. First, morphologic and rock varnish evidence argues against the notion that Big Pine fault ruptured during the great 1872 earthquake that was created by certain rupture of the Owens Valley fault to the south. Another reason to consider the Big Pine fault as separate from the Owens Valley fault is evidence that the fault moves as a pure dip-slip structure rather than the right-lateral oblique slip of the Owens Valley fault. Evidence for pure dip-slip displacement of the Big Pine fault is present in the following three places: just north of the Poverty Hills where the fault cuts a cinder cone without right lateral displacement; at the north end of Crater Mountain where an incised gully crosses the fault without lateral offset; and in alluvial fan materials north of Crater Mountain where a line of rocks marking the edge of a debris flow deposit appears to be cut by the fault without horizontal displacement. If the Big Pine fault is a separate fault from the Owens Valley fault that does not carry a right-lateral slip component, then the active tectonics of northern Owens Valley needs new interpretation. It is proposed here that the right-lateral slip of the Owens Valley fault is transferred to the White Mountains fault near the Poverty Hills and that this transfer has created the Tinemaha releasing bend.

Microbial biomass in clays from arid/ hyper-arid regions: testing for habitability in desert Mars analogs

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Next decade planetary/astrobiology missions, including the US 2009 Mars Science Laboratory (MSL) and the ESA 2013

Pasteur ExoMars, will seek key information of the geological and biological history of Mars.

The search for past (organic biomarkers) and present life, as we know it, on Mars has been supported by long-term studies on arid and hyper-arid deserts on Earth. These environments are considered a key analog model for life in dry conditions [1-4, 7]. We evaluated microbial biomass contents in phyllosilicate-rich materials to explore the possible habitability of analog mineral deposits on Mars.

Phyllosilicates, or clay minerals, have been identified on the surface of Mars by the OMEGA-Mars/Express [5], the Mars Reconnaissance Orbiter (MRO) instruments, HiRISE and CRISM, as well as inferred from rover observations in the Gusev Crater [6]. Therefore, these clay minerals will be the primary target for geochemical and microbiological in situ investigations. All of the seven MSL landing site candidates (see <http://marsoweb.nas.nasa.gov/landingsites/index.html>) include clays deposits and have been ranked by relevance with respect to their geological context, diversity, habitability (i.e., the capability to support living microorganisms), and preservation potential of organics. Although habitability has been the most ambiguous criteria to be defined, it will be the most discriminating issue for the final selection.

Methods: Total biomass. We measured ATP-based total biomass in samples with a portable hygiene monitoring system, or luminometer (LIGHTING MVP, BioControl Systems, Inc., WA, USA). As the ATP is the energy system carrier used by all living organisms, this technique is routinely applied to estimate the relative metabolic activity of cells. The ATP assay-based biomass data are expressed as Relative Luminosity Units (RLUs) and calibrated vs. Phospholipid Fatty Acid (PLFA)-based total biomass (cell/g soil) on sub aliquots of primary soil samples tested for ATP [7-8].

Gram-negative bacterial biomass. In order to estimate the endotoxin-producing microbial biomass, we used a portable system (Charles River Laboratories PTS System Package 550), which is based on the limulus amoebocyte (LAL) assay [9].

Basically, the level of endotoxin (EU/mL) is proportional to the amount of lipopolysaccharide (LPS) in the external cellular membrane of Gram-negative-like microorganisms, such as bacteria, cyanobacteria (Raetz and Whithfield, 2002), unicellular algae, and vascular plants in the rock and mineral samples. LPS amount can be then translated into bacterial biomass i.e., cell/g (1EU/mL is equivalent to about 105 cells/mL, *E. coli*-like cells).

Atacama Desert (MAP <0.5-2 mm/y; P/MEP <0.05; [11a-b]). Preliminary results suggest that plant-barren hyper-arid soil (sand to gravel supported) bear higher bio-

mass contents of Gram negatives (samples AT08-25_ANT and AT06-03_ANT = 0.55-1.0 EU/mL equivalent to 3.3-5.0 x10⁵ cells/g) than phyllosilicate-rich deposits i.e., Sample AT08-42C: <0.01 EU/mL (~6.0 x10⁴ cells/g) and hematite-rich materials i.e., Sample AT08-42B: 0.20 EU/mL (~1.2 x10⁵ cells/g) from nearby sites (Yungay Desert Station). These deposits were newly observed and sampled during the April 2008 Atacama Expedition and represent our hyper-arid end-member.

Death Valley Natl. Park (DVNP). Ubehebe crater field (MAP <50-60mm/y; P/MEP 0.05-0.20; [11a-b]). The Death Valley is one of the hottest places on Earth (record high 57 C) (record low -9 to 18 C) with the longest summers and the highest ground temperatures (~99 C). The system comprises over 13 maar volcanoes formed by phreatomagmatic explosions (~2 to 4-7 ka ago). Laminated tuff, deposited by air fall and base-surge processes [12] are plausibly associated with some of the clay mineral deposits present in the main crater area.

The Gram-negatives in surface clays from this area (Sample HUBE08-3C) resulted to be a three-order magnitude higher e.g., 106 EU/mL (~6.4 x10⁷ cells/g) than that we found in the end-member, clay minerals-rich sample i.e., AT08-42C from the extremely arid Atacama region. The clay-rich surface sample (HUBE08-3C), likely of eolian origin, was taken from a plant-colonized area, e.g. desert holly (*Atriplex hymenelytra*) and creosote bush (*L. tridentata*) and yielded higher Gram negatives based biomass e.g., <15 EU/mL (~9.8 x10⁶ cells/g) than a hematite-rich sample (HUBE08-3D). This was taken from the Little Hebe crater's rim, a plant-barren, and possibly arider and exposed wind-blast site affected by dissection. Overall, and last but not least, the Gram-negative biomass contents of Fe-oxide and clay minerals rich materials (i.e., ~10⁷ cells/g) brackets those of massive clays (~1.5 to ~3.0 x10⁷ cells/g) sampled from the semi-arid/Mediterranean coastal California sites (below discussed).

California Coast. Pescadero State Beach. Semi-arid (MAP: 750 mm/y; P/MEP: 0.45; [11]). Green/blue clays i.e., Sample PES08-1B (Qhbm: Holocene bay mud Unit; [10]), contain the highest amount of Gram negative-like cells i.e., 48.8 EU/mL (~6.0 x10⁷ cells/g) as well as total ATP-derived biomass 1475 ± 514 RLUs (N=3), which is about double than average values yielded from the overlying gray clays Unit (Qhb: Holocene basin deposit, [10]). Sample PES08-2D from this Unit yielded 24.1 EU/mL (~1.5 x10⁷ cells/g) and 912 ± 200 RLUs (N=3). The mud also contains a few lenses of well-sorted, fine sand and silt.

When compared with the oxidized, goethite-rich, poorly indurated sandstone Unit i.e., PES08-2C (Qmt: Pleistocene Marine terrace deposits Unit, [10]) the Pescadero clays units contain similar amounts of Gram negative-like cells i.e., 31.4 EU/mL (1.9 x10⁷ cells/g) than the non-clay Unit. However, this coarse-grained oxidized lithotype, appears to contain three-fold higher total ATP assay-based biomass i.e., 2858 ± 1461 RLUs (N=3) than the massive clay units (912 RLUs) below. In addition to horizontal fluxes of

biomass thru coastal fog, and ocean spray, Clay and non-clay units receives gram negative-like organisms from near surface soil horizons. The upper surface of the Qhbm Unit is indeed covered with cordgrass (*Spartina* sp.) and pickleweed (*Salicornia* sp.).

Clays and non-clay units sampled from this coastal foggy site should contain the highest and most diverse biomass, and represent a quite appropriate end-member for comparing habitability in oligotrophic environments such as minerals-rich materials formed in arid and hyperarid deserts.

The present dataset is preliminary in that is unclear whether or not environments rich in clay minerals have a higher habitability potential with respect that of non-clays environments, and a wider number of study sites should be considered. A deeper understanding of habitability of phyllosilicates and hematite-rich materials, achieved by studying new analog sites where these minerals are simultaneously present, will therefore provide critical information to help identify the best locations to search for ancient organic biosignatures on Mars.

Geochemical analyses are underway to complement information on the habitability of these relevant mars analogs in support of next decade missions landing site selection [8].

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Water level monitoring in the southern Amargosa

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Although much research, money, and scientific effort has been poured into the regions in both Nevada and California that border the proposed Yucca Mountain Nuclear Waste Disposal site, little attention has been devoted to areas farther downstream. The Amargosa Conservancy,

a non – profit land trust, is embarking upon a project to install five small monitoring wells and piezometers in the Shoshone and Tecopa area.

The goal is to establish a base line of data about the water levels and any changes in the level that may be taking place over time. The area is rich in springs, seeps, and surface water, yet there is little quantitative information available about any of these water sources. It is also unknown to what extent these water sources may be tied into the regional system. This project, a collaborative effort between the conservancy, private land owners, and the BLM, should provide a good first step toward providing reliable data, and if successful and useful it could be expanded in the future.

This presentation will show views of the five sites and outline the program we are putting in place. Contact information for anyone interested in the project or the data collected will also be provided. The presenter will be Brian Brown from the Amargosa Conservancy.

Gamma-Ray measurements of fissures along the Owens Valley Fault scarp at Lone Pine, Inyo County, California

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Measurements of gamma rays using a portable gamma-spectrometer and scintillometer across the Owens Valley Fault scarp immediately northwest of Lone Pine show a direct correlation to recent fissures and zones of alluvial piping into the fissures. The portable spectrometer and/or scintillometer permits a rapid, nondestructive qualitative identification and, with calibration, quantitative measurement of K, Bi and Tl radionuclides, the major geological sources of gamma rays. Research concluded that the surface spectrometer and scintillometer anomalies are most likely caused by the radioactivity of a uranium daughter, radon gas, that seeps up through fissures or porous alluvium.

Population dynamics of the Joshua tree (*Yucca brevifolia*): twenty-one year analysis, Upper Covington Flat, Joshua Tree National Park

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One of the most recognizable plants of the arid regions of California is the Joshua tree (*Yucca brevifolia*). Its large size, often-contorted shape, and stiff, dagger-shaped leaves give it a distinctive appearance resulting in its status as the symbol of the California deserts (Cornett, 1999). *Yucca brevifolia* is the only native tree found on open desert flatlands, the dominant plant in many areas, and one of the few species with a national park named after it: Joshua

Tree National Park. These were some of the reasons which led to the establishment, in 1988, of ten one-hectare study sites in California, Nevada, Utah and Arizona. This paper describes 21 years of monitoring the population dynamics on the site located at Upper Covington Flat in Joshua Tree National Park.

Upper Covington Flat is an alluvial plain surrounded on all sides by low hills or mountains with drainage occurring through a single, unnamed canyon to the north. Soil is a mix of coarse sand and silt. The study site was situated at an elevation approximately 1,475 meters above sea level. *Y. brevifolia* was the visually dominant plant on site and estimated to account for the greatest biomass of any plant species. Listed in decreasing order of ground cover, the perennial species found on site were *Lycium cooperi*, *Elymus elymoides*, *Yucca brevifolia*, *Atriplex canescens*, *Chrysothamnus nauseosus*, *Gutierrezia microcephala*, *Ephedra viridis*, *Prunus fasciculata*, *Purshia tridentata*, *Tetradymia stenolepis*, *Juniperus californica*, *Sphaeralcea ambigua*, *Eriogonum fasciculatum*, and *Opuntia echinocarpa*.

From 1988 through 2008 Joshua trees on the Upper Covington Flat study site were monitored each spring with regard to their dimensions, vigor, and reproductive status. Thirty-two trees were growing on the site in 1988 representing both mature and immature individuals. Based upon dimensions and leaf cluster status, 22 trees were considered to be enlarging, 7 were stable, and 3 were declining. By 2008, *Y. brevifolia* numbers had dropped to 27, approximately a 16% decrease. Of the remaining trees 15 were enlarging, 7 were stable and 4 were declining.

This data indicates that the Joshua tree was in a period of decline on the study site from 1988 through 2008. Based upon additional data from the immediate area (Cornett, 2006) and twenty years of observations elsewhere on Upper Covington Flat, it appears that this trend is not confined to the study site. Recurring drought exacerbated by a possible increase in average annual temperature seem the most likely explanations for this decline.

The research described in this paper was made possible by financial support from the Garden Club of the Desert and the Joshua Tree National Park Association.

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A perspective on desert vehicle recreation and the proposed expansion of the Marine Corps Air Ground Combat Training Center

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The motivation for this article is not to promote nor denigrate recreational off-highway vehicle (OHV) use, military use, or any other of the many multiple uses in the California Desert (Desert), nor suggest an allocation of lands

to different uses. I do not suggest to what extent—social, environmental, economic, or for national defense—one or another value is important (or more important, including the merit of the proposed expansion of the Marine Corps Air Ground Combat Center into the Bureau of Land Management's (BLM) Johnson Valley Open Area). Resource management allocations result from application of laws, policies, planning, science, public involvement, and other factors. My purpose, rather, is to provide an overview of the very controversial, complicated, and long-evolving loss of vehicle access in the Desert to help explain why OHV enthusiasts are so upset with the proposal. This short history also has application to other current and controversial land use proposals—e.g., energy “farms,” transmission lines, and who knows what other future land uses as our social, economic, and national defense needs change. Politics, conflicting land use agendas and grabs, traditional and new public needs, the Federal Endangered Species Act, etc.—all these swirled around over the years to complicate the resource management picture and invite the writing of books. Given the results today, perhaps they also suggest a subtitle to this article: shoe-horning too many land uses onto too little land.

For 40 years, beginning in the 1920s, vehicle enthusiasts (i.e., people who drive on roads to get somewhere, as well as those who drive cross-country on motorcycles, ATVs, and dune buggies for family fun and competitive racing in places like the Johnson Valley Open Area and the Imperial Sand Dunes) enjoyed an upswing in trails and access. By the 1960s, such vehicle use had created an access v. environment fight, hitting at the heart of public land management difficulties in the Desert, and is one that still rages today. Given the conflict, the many fragile Desert values, new resource protection and management laws, and the realization that land is finite while population grows, it was inevitable and necessary to bring management to the Desert and BLM, largely, was the lightning rod agency that had to do it.

Following is a chronology of major milestone events that came to affect Desert vehicle access and reducing it to the situation of today:

1. Until the 1930s one could drive a vehicle just about anywhere a vehicle could go
2. 1930s: 2 national monuments are designated: Death Valley and Joshua Tree
3. 1940s: several military reservations are created to meet the WWII emergency and conduct training and research-development into the future

By 1945 approximately 25% of the Desert is “off limit” or severely restricted to vehicle use. Environmental effects of vehicle use to this point and into the 1950s are probably light.

In the 1960s public concern began to rise over the growing use and effects of unmanaged vehicle use—particularly new and affordable motorcycles—to the point that in 1976 Congress felt compelled to designate the 25 million acre California Desert Conservation Area (CDCA) and

mandate that, by 1980, BLM address conservation and use needs for the CDCA—with considerable public and stakeholders involvement—through a comprehensive land use plan. The completed CDCA Plan did close a modest amount of areas, and it also defined the end of proliferation of vehicle roads and trails. It also suggested the need to reduce vehicle access, through further planning and management, to bring it in line with other Desert resources/uses sensitivities. An array of conservation and use “zones,” area designations, and other commitments were set in place for a host of resources and uses: e.g., vehicle use on roads/in areas, areas of critical environmental concern, livestock grazing, wild horses and burros, mining, utility corridors, species and habitats, cultural resources, and community expansion. The CDCA Plan also proposed areas for Congress to consider and designate as wilderness.

The fight over land uses did not go away with the completed CDCA Plan, however: native species were increasingly being listed as threatened or endangered under the Federal Endangered Species Act (ESA), due to the effects of disease and human encroachment; and, not satisfied with the outcome nor speed of plan implementation, environmental groups agitated for more wilderness than was recommended, creation and expansion of federal parklands, and more vehicle restrictions. As they watched, vehicle users, mining companies, and ranchers became increasingly concerned as to where all this agitation would lead. At the same time some military installations saw a need for more land to meet new technology and warfare needs. In short, environmentalists did not trust BLM, its multiple use management mandate, the new plan, and budget to get the job done; and all interests were concerned about BLM's ability to “referee” the situation going forward.

4. In 1994 Congress, through the California Desert Protection Act (CDPA), made a declaration on the different wilderness and parkland proposals from BLM and environmental groups. Where the CDCA Plan recommended 2.1 million acres for wilderness designation, Congress designated 3.8 million acres (BLM) and transferred an additional 3.5 million acres of BLM-managed lands to the National Park Service, most of which also became wilderness—in all, 3.75 times what was proposed in 1980.

The CDPA hit vehicle, mining, and other “intensive” uses hard: the amount of Desert now “off limit” or severely restricted to vehicle use jumped from 25% to 50%. Affected were many roads and favorite camping areas, 50% of the favorite rock hounding areas, and the availability for discovery and extraction of 40 out of 49 kinds of minerals (based upon potential). Ominously, CDCA Plan flexibility to provide for future needs (e.g., military expansion and energy needs) was severely reduced. The impact of these numbers is even more severe considering that:

a. half of the remaining 50% of the Desert—another 25% of the Desert “pie”—is private land and unavailable for public uses, and b. most of the designated wilderness is high elevation and remote—out of critical habitat range for most species listed under the ESA—i.e., the CDPA ignored

species and habitats issues. Consequently, the remaining 25% of the Desert (i.e., not military, wilderness, parkland, or private) is still not severely restricted for vehicle use, but half of it does have the dark cloud of species and habitat issues hanging over it and is subject to further controversy and vehicle restrictions as these issues are addressed (below).

5. In 1990 the desert tortoise was listed as a threatened species under the ESA. This listing, geographically the most widespread in the Desert, and other listings (e.g., Pierson's milkvetch in the Imperial Sand Dunes and bighorn sheep in the mountains southwest of Palm Springs–La Quinta) required BLM to consult with the U.S. Fish & Wildlife Service (FWS) to address the adequacy of the CDCA Plan to resolve these species issues (other federal agencies with affected habitat also had to consult on their land use plans). BLM and FWS did agree that the CDCA Plan indeed needed to be amended. For BLM to do this, given its large size, the CDCA was divided into six geographically separate plan amendments in the 1990s. In 2000, the amendments in progress, three environmental groups sued BLM in federal district court in San Francisco to (in my mind) help “guide” amendments to more preservationist conclusions—i.e., to force more vehicle use restrictions than might otherwise have come out of the amendment processes. (Other uses, particularly livestock grazing, were also targeted.) The plan amendments now had to satisfy both ESA administrative requirements and a federal district court judge.

With the completion of the six plan amendments to the satisfaction of the agencies involved, primarily the BLM and FWS, the 2000 lawsuit still rages regarding the desert tortoise. On the one hand are the three environmental groups, while on the other are the combined science, wisdom, and give-and-take of a host of federal, state, and local agencies and a considerable array of stakeholders. The plan amendments combined, all popular vehicle open areas remain open, while 3,671 miles (20%) of 17, 588 miles of inventoried roads in the Desert were closed. This does not include an estimated 4,500 miles of roads closed with the passage of the CDPA in 1994; so together the total closed is closer to 36% of miles of roads on BLM-managed public lands that existed in 1994. Also of significance, huge areas of desert tortoise habitat were dedicated to the recovery of the desert tortoise (i.e., were designated areas of critical environmental concern—ACEC), a large fraction of the remaining 25% of the Desert noted in 4.b., above. The road closures noted above are primarily in these ACEC areas.

The beat goes on. Not one but two military base expansions onto public lands are under consideration today: Ft Irwin, north of Barstow, and the Marine Corps Air Ground Combat Training Center, southeast of Barstow. Additionally, BLM has about 70 solar-wind farm applications on hand amounting to thousands of acres. So as to not affect the desert tortoise and other listed species these and other future intensive land use proposals will be directed as much as possible into non-ACEC areas, further compressing and stressing conflicts and competition involving OHV

recreation. As noted above, there is only a finite amount of land. Vehicle users feel they are painted into a corner and the corner continues to shrink.

Finally, to put the proposed Marine Corps expansion into the Johnson Valley Open Area (JVOA) into perspective: the CDCA Plan placed about 550,000 acres (combined federal and private) into the “Intensive” multiple use management zone. “Intensive” areas are for both large mining areas and OHV recreation. About 135,000 acres are mining related and 415,000 acres are OHV-related. The 415,000 acres amounts to about 1.6% of the 25 million acres CDCA. At about 190,000 acres the JVOA alone comprises about 45% of the total acres of OHV open areas. Of the six Marine Corps alternatives proposed to date three almost entirely drape over the JVOA and would effectively delete it in its entirety as a BLM OHV open area.

The reader should now better understand the issues. This is not the only Desert resource management story, but it is one that hits home for vehicle users and others and should be considered under “cumulative effects” in an environmental impact statement.

Most of the factual information and figures included in this article are taken from BLM documents: 1980 CDCA Plan, 2002 Northern & Eastern Colorado Desert Plan and other amendments to the CDCA Plan completed between 2002 and 2007, and derived from BLM's Geographic Information System. Acreage figures related to Johnson Valley Open Area and other BLM open areas are estimated, however, so BLM should be consulted for more precise figures. The 25 million acres of CDCA does not include the Owens Valley. OHV open area figures do not include the 40,000 acres Ocotillo Wells State Recreation Vehicle Area or other OHV areas located in Southern California but outside the CDCA.

Author

Richard Crowe worked for the BLM for 33 years, 29 of them in the CDCA dealing with a wide range of multiple use management issues. His job assignments included Field Manager for the Needles office, staff manager for Operation for the entire California Desert, and species and habitats planner (was lead for Northern & Eastern Colorado Desert Plan, a species and habitats plan amendment completed in 2002 to the 1980 California Desert Conservation Plan). Richard retired in 2007 and lives in Beaumont, CA.

A photographer's exploration

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A landscape photographer's interpretation under a variety of lighting conditions of a diversity of subjects in the region from the San Gabriel Mountains, north through the West-ern Mojave Desert and Owens Valley to Mono Lake.

Geochronology and paleoenvironment of pluvial Harper Lake, Mojave Desert, California

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The Mojave River is a well-known recorder of Southern California paleoclimate with a complex paleohydrology and past terminations in pluvial (upstream to downstream) Harper (Harper basin), Manix (Afton, Coyote, and Troy basins), and Mojave (Soda and Silver basins) lakes over the last 30,000 years. Previous studies yielded uncalibrated radiocarbon ages ranging from 24 to >30 ka yrs BP for highstand lake deposits near 656 m elevation. Based on several studies, the present hypothesis is that the Mojave River: 1) flowed simultaneously into Harper and Manix lakes ~30 ka; 2) the river then flowed exclusively into Manix Lake 28–25 ka; 3) then, resumed simultaneous flow into Harper and Manix lakes, forming the Harper Lake highstand ~25 ka; 4) the Mojave River ceased flowing into Harper basin and the lake receded. Being upstream and consisting of a single basin without internal sills, pluvial Harper Lake is relatively uncomplicated compared to the other terminal basins. Here we present geologic mapping (1:12,000), a measured stratigraphic section, and radiocarbon ages from the Red Hill area. The 2.1-m-thick continuous stratigraphic section is near the highstand elevation and comprised of interbedded sand, silt, and silty sand capped by a 0.6-m-thick sequence of carbonate mud resting nonconformably on quartz monzonite. Lacustrine sediments contain four shell horizons (*Anodonta californiensis*) and ostracodes (genera *Limnocythere*, *Candona*, and *Heterocypris*). Each shell horizon was sampled yielding seven calibrated radiocarbon ages ranging from $33,645 \pm 343$ to $40,155 \pm 924$ cal yrs B.P. Our radiocarbon ages and the continuity of the section support a single Harper Lake highstand between 40 and 33 ka with no subsequent hiatus or second highstand at 25 ka as previously hypothesized. Preliminary ostracode analysis yields a tentative interpretation that Harper Lake water may have been periodically less saline than other Mojave River-fed pluvial lakes. The radiocarbon data suggests that Harper Lake overlaps phases of Lake Manix.

The Kit Fox Hills, Death Valley, CA: a propagating pressure ridge adjacent to the Northern Death Valley fault zone

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The Kit Fox Hills (KFH) are 17-km-long, 4-km-wide, low hills located in northern Death Valley, California. The KFH are composed of poorly indurated, laminar, thin-bedded fine sand to clay beds deposited in distal alluvial fan to playa lake paleoenvironments. The central KFH attain an elevation of ~250 m amsl and decrease to sea level to the north and south. The base of the sediments exposed at this central highpoint is late Pliocene (~3.5 Ma) playa to perennial lake deposits. The lake sediments change abruptly to coarse conglomerates (alluvial fan environment) upward just above the 3.28 Ma Nomlaki Tuff Member of the Tuscan and Tehama Formations. To the north, the generally conformable section of conglomerates with sparse tongues of playa lake transitions back to playa lake deposits that contain 0.8–1.2 Ma upper Glass Mountain ash beds and the 0.77 Ma Bishop ash bed. To the south, the Pliocene conglomerates are overlain by distal alluvial fan/playa deposits that contain the 0.62 Ma Lava Creek B ash bed. Structurally, the KFH are northeast of the right-lateral Northern Death Valley fault zone (NDVZFZ) with an unnamed fault bounding the east side. In the central KFH, (a) this east-side fault offsets (normal) alluvial fan deposits that cap the hills, (b) relief is greater (182 m), (c) active channel profiles are concave up, and (d) the playa lake sediments are overturned and isoclinal adjacent the NDVZFZ. In contrast, to the north (a) the east-side fault is a series of subparallel, curvilinear, normal faults that offset younger alluvial fans, which are found as terraces within canyons, (b) relief is only 12 m, (c) active channel profiles are convex up, and (d) folds are open. The variation in KFH structure and geomorphology is consistent with a central asperity that has propagated both north and south. Based on proximity to the NDVZFZ, the KFH are interpreted as a flower structure formed by the NDVZFZ. The NDVZFZ is a relatively new element in 8+ Ma years of Death Valley extension and is similar to basin-cutting faults that develop in highly extended modeling studies and observed in other extensional basins.

Natural formations and cultural behavior III

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In the five years that I have been participating in and managing CRM archaeology in southern California, a trilogy of cultural traits that are closely correlated with landscape formation processes have focused some of my attention. They are: (1) black soils purported to be cultural middens; (2) "cupules", or cuplike indentations on granitic boulders that are purported to be of cultural origin (or at the very least to have been culturally enhanced) and to have served a wide variety of ceremonial and practical functions; and (3) yonis, that similar to cupules are thought either to be cultural portrayals of female genitalia, or enhancements of natural occurring rock forms that call such anatomical details to mind. Co-authors and I have dealt with the first two cultural/formation processes in previous presentations at the Desert Studies Symposium.

This presentation is an overview of the cultural and formational aspects of yonis. In particular, I focus on: (1) the anthropologist's/archaeologist's role in influencing the Native American interpretation of natural formational processes; (2) the problems caused when misidentifications by the anthropologist/archaeologist become accepted in cultural interpretations; (3) the difficulties in distinguishing between natural and culturally enhanced rock formations; (4) the difficulties of correlating religious behavior of regional tribes with the natural boundaries of more restricted geological formations that are millions of years old, and (5) the realities of the lack of historical continuity in the identification of yonis in the protohistoric and historic period.

Evaluation of minor and trace elements of the Wilson Creek Formation, Mono County, California

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The Pleistocene Wilson Creek Formation is composed of interbedded mudstones and 19 tephra layers at the type locality along Wilson Creek, northwest of Mono Lake, California. These tephra layers range in age from 32 ka to 13 ka years BP (uncalibrated ^{14}C ages). The distribution of the Wilson Creek ash layers is difficult to assess because the glass shard compositions are indistinguishable from each

other. A possible exception is ash bed #15 that is found as far east as Utah and is also associated with the Mono Lake paleomagnetic excursion (MLE). Single-shard electron probe microanalyses (EPMA) and bulk Instrumental Neutron Activation Analyses (INAA) of the glass shards from many of the tephra layers exist in the U.S. Geological Survey Tephrochronology database. Portions of these data were published in previous studies, however, presentation and examination of the entire data has not been done. In this study, we present similarity coefficient calculations comparing EPMA and INAA data for 18 of these tephra layers. Focusing on ash bed #15, the EPMA data show that #15 is geochemically similar to seven other Wilson Creek ash layers, whereas #15 is distinctive by INAA-measured minor and trace elements. Ash bed #15 also has a unique light rare earth element (La_N/Sm_N) fractionation. We hypothesize that these more accurate TOF-LA-ICP-MS data will allow distinction of individual Wilson Creek Formation ash beds or groups of beds, thus making these ash layers more valuable as time-stratigraphic marker beds for the late Pleistocene in the western United States.

Spatial properties of the San Andreas Fault plate boundary surface trace between Desert Hot Springs and the Bombay Beach based on satellite and B4 imagery.

David K. Lynch and Kenneth W. Hudnut (USGS)

We present preliminary results of a hyper-accurate (\pm few m) inventory of San Andreas Fault (SAF) features between Bombay Beach and Desert Hot Springs based on overhead imagery. The ultimate goal is to identify fault features that can be used to determine offset distance and slip rates. Many fault structures have been previously reported and about two dozen new ones have been found in this study (vegetation lineaments, offset channels, soil color changes, scarps, pressure ridges, etc.). Using approximately one hundred fault components we defined a map view piecewise continuous trace of features (and interpolations when no structure could be discerned) that we are calling the provisional plate boundary (PPB). The resulting PPB trace closely matches the faults reported by Clark (1984) and those in the faults data base. Using a variety of techniques including radius-of-curvature retrieval and wavelet analyses, we analyzed the PPB and present evidence of spatial structure on scales of a few hundred meters.

Reproductive ecology of female Sonora mud turtles (*Kinosternon sonoriense*) in Montezuma Well, Arizona: optimal egg size in a suboptimal environment?

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We studied the reproductive ecology of female Sonora mud turtles (*Kinosternon sonoriense*) at Montezuma Well for two years, 2007 and 2008. Montezuma Well is in central Arizona near the northern limit of distribution for the species; a species believed to be extirpated in California. This unusual natural wetland is characterized by large spring-fed discharges of water at a relatively steady temperature (24°C) and extremely high concentrations of dissolved CO₂ and arsenic, in the form of arsenate. The hydrologic characteristics of Montezuma Well create a chemically challenging environment for aquatic organisms. As a result, there are no fish and the aquatic invertebrate community is depauperate, but includes several endemic species.

Female mud turtles at this site mature between 115.5–125 mm carapace length and the proportion that produced eggs ranged from 23.81–45.65%/yr. Eggs were detected in x-radiographs from 1 May to 28 September 2007 and 23 April to 30 July 2008 and the highest proportion of adult females with eggs occurred in July (both years). Clutch size ranged from 1–8 with a mean of 4.96 and did not differ significantly between years. Clutch size was only weakly correlated with body size and only one female produced more than one clutch per year during the two years of our study. X-ray egg width (XREW) ranged from 17.8–21.7 mm (mean 19.4 mm) and varied more among clutches than within. Mean x-ray egg width (MXREW) of a clutch did not vary with carapace length of females, although the slope of x-ray pelvic aperture width (XRPBW) increased allometrically with the same measure of body size. The heterogeneity of the two slopes demonstrates a lack of morphological constraint by the XRPBW on MXREW in this small species. In addition, greater variation in clutch size, relative to egg width, provides support for the hypothesis that egg size may be optimized in this hydrologically stable

but chemically challenging habitat.

Nesting and hatchling emergence are coincident with the onset of summer rains, the latter facilitated by embryonic diapause and development totaling almost 11 months. The reproductive strategy of this species and its ability to survive elsewhere in ephemeral desert streams and tanks appear to be relatively recent adaptations to the desert climate in the region. The adaptive value of this strategy in the hydrologically stable environment of Montezuma Well is perplexing but may be maintained by genetic exchange with nearby populations in less stable habitats. We observed substantial differences in various traits of reproductive female mud turtles with previous studies including a relatively large size of sexual maturity, small maximum body size and few multiple clutches—a possible result of challenging conditions and associated slow growth rates at Montezuma Well. Our data contradict earlier predictions regarding the existence of morphological constraints on egg size in small turtles. We conclude that the diversity of architectures exhibited by the turtle pelvis, and the associated lack of correspondence to taxonomic or behavioral groupings, explains much of the variation observed in egg size of turtles. Concomitantly the architectural diversity also determines whether egg size is constrained or optimized.

Chronology of pluvial Lake Coyote, California, and implications for Mojave River paleohydrology

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During the late Pleistocene, the Mojave River episodically fed pluvial Lake Coyote, which lay in the central Mojave Desert, California. The Coyote basin was a sub-basin of Lake Manix until about 24.5 cal ka (calendar years before present), when Lake Manix failed and the Mojave River fed either Lake Coyote or Lake Mojave, which was farther downstream (e.g., Meek, 2004). Shoreline geomorphology and dated lacustrine depositional sequences associated with Lake Coyote can be combined to show four pluvial lake episodes: 1) a brief lake rise ~23.6 cal ka; 2) a sustained rise between ~19.7 to ~18.3 cal ka, which included one or more fluctuations in lake level; 3) a sustained rise between ~17.2 and ~16.3 cal ka that included one fluctuation; and 4) a brief ~15.5 to 15.2 cal ka lake that is associated with a shoreline at 540 m. Lake episodes 2 and 3 correspond to shoreline features at 542 and 541 m altitude, respectively, and probably coincide with overflow from the Coyote basin. This Lake Coyote chronology can be combined with Lake Mojave chronology (Wells et al., 2003) to infer times of sustained Mojave River flow, which corresponds to periods of enhanced winter storm precipitation in the Transverse Ranges. The combined chronology suggests sustained river flow and deep lakes between 24.5 and ~13.5 cal ka except for a period of intermittent river flow to these basins

between ~23 and ~22 cal ka. Intermittent flow also characterized the period from ~13.5 to ~9.9 cal ka, but during this last period river flow sustained more and deeper lakes during the Younger Dryas interval than during other times. Our new chronology is similar to the timing of withdrawal of pluvial Lake Bonneville in the northern Great Basin, and suggests that latitudinal shifts in weather patterns at the termination of MIS 2, if they occurred, were rapid.

Assessment of plant height and cover using LIDAR in the Mojave Desert

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The ability to assess desert plant communities at the landscape scale is invaluable for understanding broad changes in community structure and composition, especially where change occurs slowly over time. Increases in atmospheric CO₂, changes in precipitation patterns and increasing temperatures are among some of the factors predicted to affect desert plant communities in the long-term. LIDAR (Light Detection and Ranging) is an emerging technology for the assessment of vegetation structure on a landscape scale. It has been successful in estimating biomass and densities in forest ecosystems but may be much more limited in deserts due to uncertainties caused by the typically diffuse canopies. Airborne LIDAR remote sensing was used to measure vegetation heights across the Hayden piedmont in the Mojave Desert, located just east of Kelso Depot in the Mojave National Preserve, California. We quantified the capability of LIDAR to assess plant heights and canopy areas in a region dominated by creosote bush (*Larrea tridentata*) and white bursage (*Ambrosia dumosa*). Small footprint LIDAR measurements were interpolated into a DEM with 1x1 m grid cells and LIDAR height and cover data were compared with ground-based manual measurements of vegetation. Measurements were also used to assess variation between the vegetation structure above the Cima Road, which bisects the Hayden alluvial-fan complex, and areas below the road into which water from above the road was diverted (water-added) or where natural surface-water flow was cut-off (water-deprived). Vegetative cover based on LIDAR was greatest in the water added areas (4.10 cm² vegetation/m² ground); nearly seven times greater than that of the control areas (0.60) and 100 X greater than that of the water deprived areas (0.04). Measured plant heights were consistently larger in water-added areas below the road (2.46 m ± 0.45, n = 71) compared to the control (1.78 m ± 0.36, n = 27) and water-deprived (1.67 m ± 0.58, n = 24) areas. LIDAR heights were positively correlated with measured heights (r² = 0.44; p < 0.001), however LIDAR consistently underestimated the actual plant heights by about 1 m. The diffuse canopy of the desert shrub *Larrea tridentata* and LIDAR pulses reflecting off the denser inner-plant canopy rather than the maximum height of the plant is believed

to be the reason for the underestimation of plant height. Although LIDAR does not accurately measure plant height, the offset between the actual height and LIDAR height may be consistent enough for this technology to be applied for evaluation of plant biomass and density, and community structure in desert ecosystems.

America's resources and garbage

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Since about 1946, Americans have come to equate the American Dream with accumulating stuff. It is not an exaggeration to say that our economic model boils down to: digging up minerals and cutting down trees, using them to manufacture a range of products, then very quickly transferring the products to dumps, where innocuous substances combine with highly toxic wastes. The speed of this transition is designed to keep consumption high and the economy growing.

The result is that garbage and other wastes—municipal garbage; residential and agricultural wastewaters; mining and industrial solid and liquid wastes; transportation and industrial emissions, and much more—have become our nation's largest-volume product, much of it toxic. Supposedly “clean” information technologies create some of the most poisonous wastes.

The actual grand total of U.S. wastes is 25 billion tons, about a quarter of a ton per person, every day. The nation's annual solid waste production would fill 90 million mine-hauling super dump trucks, the kind that tower three stories tall and have 12-foot-high tires. Placed nose to tail at the equator, they would ring the earth 7 times. Less than 2 percent of all these wastes is now recycled.

We still rely mostly on natural attenuation by soil and dilution by air currents, and soil and water organisms, to reduce even toxic wastes to innocuous substances. But the sheer abundance of hazardous compounds in U.S. landfills and urban and industrial areas overwhelm the processing potential of natural systems. Even worse, natural processes spread unreduced toxins into our soil, our water, and the air, and so into our food and drinking water supplies. No matter where we try to “dispose” of our wastes, natural processes constantly bring them back to us. Global climate warming and water pollution from all sources are today's most prominent examples of the severe threats from our level of waste production.

Toxic wastes degrade the health of the land and water that support us. The cumulative effects of accumulating stuff include depleted soils; depleted industrial minerals; accumulations of toxic materials in and on plants, including food crops; depleted and polluted streams and aquifers; air pollution and de-stabilized climates; species extinctions; and toxic human exposures.

Garbage production is the other side of resource depletion. Most of our natural resources now reside in toxic waste dumps, and energy costs may preclude cleanup. Understanding the source and extent of these issues is the necessary first step toward ameliorating the problems and finding solutions.

Desert writers: the conservationists

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...SO ARE THE GREAT REACHES OF OUR WESTERN DESERTS, SCARRED SOMEWHAT BY PROSPECTORS BUT OTHERWISE OPEN, BEAUTIFUL, WAITING, CLOSE TO WHATEVER GOD YOU WANT TO SEE IN THEM.—Wallace Stegner, from *The Geography of Hope* (1969).

California desert protection and conservation were only vague ideas in the mid 20th century. At that time, the deserts of California—the Mojave and Sonoran, spanning parts of seven counties—were lightly populated, seldom visited, and viewed as mostly impenetrable and vastly indestructible. The relatively small numbers of Native American people who lived in the deserts for centuries prior to that wisely used its precious resources with sustainable practices. Today, however, the California desert competes with other areas in the west for urgent environmental protection, legislation, and practices. In the past fifty years, the desert's human population alone has far surpassed anything previously imaginable; the numbers of visitors, off-road-vehicle recreation-seekers, military bases, urban refuse sites, power-lines, and highways has caused an enormous strain on the desert's resources. Desert conservation literature began to emerge as early as the late 19th century, with the authors Mary Austin and John C. Van Dyke helping shift public awareness of this area from its former depiction as a life-threatening “desert horribilis,” to be avoided at all costs, into a well-rendered portrayal of the desert as a land of beauty and magic, a place indispensable to the renewal of the human spirit and body and indeed, even civilizations themselves.

The power and passion of literature, as it has throughout time and history of the human race, has played a key role in creating and building a body of works that continue to inform, educate, and move citizens and governments at the local, state and federal levels in today's desert conservation and protection efforts. Naturalist and author Edmund Jaeger made the vital understanding and interpretation of the California deserts, often in layman's terms, the focus of his life's work. The often-inflammatory writing of Edward Abbey, starting in the 1960s, helped spark flames of consciousness of the imminent danger to the west's remaining open desert lands in prose that stirs the human heart to action. The writers Mark Reisner and Joseph E. Stephens

have chronicled the miracle and farce of water reclamation of the west and its limited ability to serve human demands on fast-increasing desert populations.

The passion and power in the prose of desert conservation writers such as Wallace Stegner, Aldo Leopold, Ann Zwinger, and Barry Lopez helped shape a public consciousness that led in part to the passage of the California Desert Protection Act in 1994, and more recent works by writers such as Rebecca Solnit, Ruth Nolan, Lawrence Hogue, David Darlington, and the team of Howard Wilshire–Jane Nielsen–Richard W. Hazlett continue their pioneering work: to inspire desert –conservation and protection throughout California's deserts, not only among the scientific and environmental elite, but the public at large. The works of the California desert conservation writers, alongside the efforts of those in the many branches of research and natural sciences, are an indispensable part of this movement, and thanks to their efforts, in so many cases, the pen has truly proven to be... “mightier than the sword.” As the great British poet Percy Bysshe Shelley said: “poets (and writers) are the unacknowledged legislators of the world.”

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The Art of Nature: images from the wildlands of southern Nevada

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This multimedia presentation was produced in collaboration with the Nevada State Museum as a portion of a Mojave Desert natural history exhibit. The solo exhibit involved exploration of nine wildland areas in southern Nevada selected for their unique qualities of landscapes, flora, and fauna. The presentation and resulting museum exhibit include a collection of photos, sketches, and paintings showcasing the unique nature of each area in an attempt foster the appreciation and stewardship of these natural areas through the use of artistic images.

A glimpse of the hidden beauty and grace of nature can change a viewer. By looking at these images the viewer comes away with a different perspective—no longer seeing the public lands of southern Nevada as a desert wasteland, but rather as a place of unparalleled natural beauty and diversity that is deserving of our care and concern.

The museum exhibit was at Nevada State Museum—Las Vegas for 8 months and in September of 2007 traveled to Nevada State Museum—Carson City where it will be on display for a year before traveling on.

The Art of Nature multimedia presentation has been presented at the University of Nevada Las Vegas as part of their Forum Lecture series as well as to over 1400 people in the Las Vegas area at schools, conferences, and meetings. It has proven to be a unique and effective way of promoting knowledge and stewardship of the Mojave Desert wildland areas of southern Nevada.

The Ashford brothers, their lives as miners in Death Valley and as residents of Dublin Gulch

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The landscape around Shoshone is a spectacular jumble of bright colored mountains framed by low-lying white hills. These striking white hills were formed by ancient Lake Tecopa and by volcanic ash deposits resulting from three volcanic eruptions spanning over two million years. The sources of these ash deposits were Huckleberry Ridge in Yellowstone, which took place over two million years ago; the Long Valley caldera near Bishop that took place 760,000 years ago; and Lava Creek in Yellowstone, which erupted 620,000 years ago. The Lavas Creek ash beds are particularly striking since outcrops near the edge of town of Shoshone form stark white cliffs that catch the eyes of visitors as they pass.

In the mid twenties the white ash cliffs, called Dublin Gulch, became the home of several prospectors who carved their dwellings into these cliffs. From the 1920s until the 1950s Dublin Gulch became a well known social center among the many prospectors, miners, and mining promot-

ers who were seeking their fortune in the red hot mining industry that had become the center of the social life of the communities of Shoshone and Tecopa. Death Valley Scotty and Shorty Harris were two of the famous characters in Death Valley history who often could be found at the Gulch.

Three of the residents who made Dublin Gulch their home were the Ashford brothers. These brothers, named Harold, Henry, and Louis Ashford, were Englishmen who came to the United States to seek their fortune. Their first home in the United States was Bakersfield where they bought and operated an alfalfa farm. After hearing many stories from the Rand mining district in California of prospectors with a dream and a grubstake acquiring fabulous wealth, they sold their farm and traveled to the Rand where they became very successful. After several years in Randsburg and Johannesburg they were lured to the rich mining districts of Death Valley where they owned and operated the gold mine they optimistically named the Golden Treasure. They were mining entrepreneurs in Death Valley for over forty years, and their efforts, in many ways, embody the cycle of “boom and bust” inherent in the mining industry.

In this paper I chronicle the lives of the Ashford brothers as individuals, their reasons for coming to Death Valley, their interaction with the community of Shoshone, and their place in the mining milieu of the surrounding communities. Through their lives, and the lives of their neighbors, I explore how the geologic landscape of the Amargosa and Death Valley, as embodied in the mining industry, has shaped the lives of its residents.

Medicinal plants of the Mojave Desert

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Many people look at the desert and see a barren landscape devoid of life. Others see beauty and richness in that very same landscape. Many peoples have lived in the desert, surviving quite well, creatively using the flora and fauna available in the region.

A number of Desert Indian tribes have populated the Mojave including the Cahuilla, the Serrano, the Paiute, the Kawaiisu, the Mohave, and the Chemehuevi. Although there may have been regional differences, these tribes used the local flora in very similar ways, based on the belief that the Earth would provide them what they needed to survive. Prior to the advent of modern medicine, ailments were treated using whatever resources people could find. People experimented, in particular, with plants and often serendipitously discovered some form of medicinal property.

This presentation will discuss a number of extant plant species found in the Mojave Desert that have been used for medicinal purposes. A brief overview of what and how these plants were used by the indigenous people will serve as an introduction to what we see today. Medicinal proper-

ties and preparation for use of some of the extant species will be discussed as well as the efficacy of the “medicine” that is produced.

War in our back yards

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Making war abroad has always involved practicing warfare in America’s back yard—especially the western states. All kinds of military training, including severely contaminating weapons tests, have degraded United States soils, water supplies, and wildlife habitats. The deadly components of some U.S. weaponry have rarely or never been used on foreign soil, but their production and testing have attacked U.S. citizens in their own homes and communities.

During WWII, the U.S. accelerated top secret programs to develop agents and weapons of “unconventional warfare.” Throughout the Cold War, many western locations became sites of weapons building and testing for nuclear, chemical, and biological weapons, and for dumping the wastes. Widespread chemical contamination from weapons research, manufacture, and testing may affect as much as 870,000 acres at 59 western military installations. No one has yet invented a feasible method for cleaning up or disposing of many of those hazardous pollutants.

Some lands will remain unsupportive of wildlife breeding and survival at best, and at worst will harm people, their economic interests and recreational pursuits, far into the future. Turning over training lands laden with unexploded weapons (UXO) for human uses without appropriate mitigations puts members of the public at greater risk of sickness, injury, or even death from the weaponry itself or its constituent contamination. An EPA survey of 61 UXO sites listed five accidental explosions, including three fatalities and two injuries, plus “incidents” at 24 other former DOD facilities.

At many western sites, nuclear bomb development, production, testing, and waste disposal exposed uranium miners; bomb production workers; bomb-test workers; bomb-test observers (some, mostly U.S. soldiers, were forced participants); civilians living downwind; and the atomic scientists themselves to radiation. Many suffered severe health effects. During WWII, the United States also accelerated top secret programs to develop “unconventional warfare” agents and weapons, detonating more than 55,000 chemical rockets, artillery shells, bombs, and land mines at the former Dugway Proving Ground west of Salt Lake City (now Utah Test and Training Center), releasing deadly nerve agents into the air, as well as biological agents that cause fatal diseases. The agents sometimes drifted over nearby ranches and reservations—in 1969 a VX nerve agent release killed sheep on a ranch halfway to densely populated Utah cities.

Since 2001, the U.S. government has proliferated new biochemical research and test programs, and proposed to revive making nuclear weaponry that supposedly needs no testing. Since 9/11 and the 2001 anthrax exposures, funding has poured into bioweapons research. The U.S. now has 12 biosafety level-4 labs (handling the worst pathogens, lacking antidotes). Six universities are among 8 applicants for new level-4 labs. Eighty-four biosafety level-3 labs are operating or under construction. Audits reveal many security violations in these labs and 13 recent serious problems, ranging from accidental to careless exposure. As a nation, we have to face the fact that we are armed and dangerous to ourselves, and decide where our real security resides.

—Notes—