

OVERBOARD IN THE MOJAVE

20 MILLION YEARS OF LAKES AND WETLANDS

edited and compiled by

Robert E. Reynolds and David M. Miller



and

**ABSTRACTS OF PROCEEDINGS:
THE 2010 DESERT SYMPOSIUM**

compiled by

Robert E. Reynolds

Desert Studies Consortium
California State University, Fullerton
April, 2010

FRONT COVER:

The Troy Lake arm of Manix Lake laps against the Newberry Mountains, January 24, 2010.

BACK COVER:

Above: The Barstow Formation in the Mud Hills is the type locality for the Barstovian land mammal age of the North American continent. Biostratigraphic ranges of taxa are constrained by dated volcanic ashes and magnetostratigraphic polarity.

Below: Strata of the Barstow Formation have been folded, faulted, and planed by erosion in the last 10 million years. The colorful Miocene sediments are unconformably overlain by flat-lying tan arkosic sandstone.

TITLE PAGE:

Rob Fulton kayaked on Silver Lake in 2005.

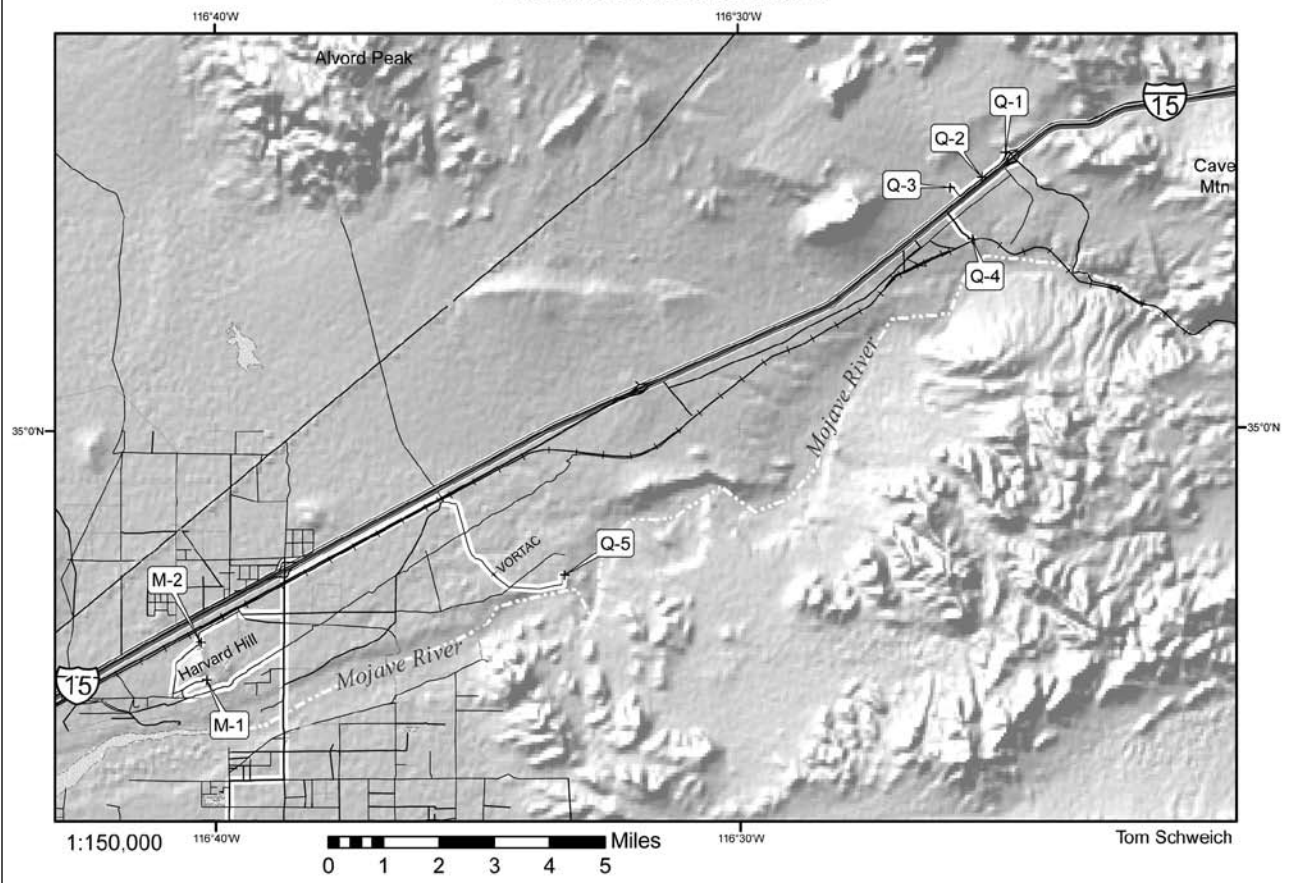
Table of contents

Overboard in the Mojave: the field guide	7
Robert E. Reynolds and David M. Miller	
Environments of nearshore lacustrine deposition in the Pleistocene Lake Manix basin, south-central California	24
Marith C. Reheis and David M. Miller	
Stratigraphy and paleontology of the middle to late Pleistocene Manix Formation, and paleoenvironments of the central Mojave River, southern California	38
George T. Jefferson	
An interview with Elmer Ellsworth, the first researcher to study the Pleistocene deposits in Afton basin	57
Norman Meek	
A brief comparison of central Mojave Desert lakes: Miocene Barstow Formation and Pleistocene Lake Manix	66
Robert E. Reynolds and George T. Jefferson	
Reconnaissance geochronology of tuffs in the Miocene Barstow Formation: implications for basin evolution and tectonics in the central Mojave Desert	70
D. M. Miller, S. R. Leslie, J. W. Hillhouse, J. L. Wooden, J. A. Vazquez, and R. E. Reynolds	
Stratigraphy, age, and depositional setting of the Miocene Barstow Formation at Harvard Hill, central Mojave Desert, California	85
Shannon R. Leslie, David M. Miller, Joseph L Wooden, and Jorge A.Vazquez	
Correlation of the Miocene Peach Spring Tuff with the geomagnetic polarity time scale and new constraints on tectonic rotations in the Mojave Desert, California	105
John W. Hillhouse, David M. Miller, and Brent D. Turrin	
John C. Merriam and the Barstow Syncline	122
Thomas Schweich	
The mammalian litho- and biochronology of the Mojave Desert Province	124
Michael O. Woodburne and Robert E. Reynolds	
Extending the boundaries of the Barstow Formation in the central Mojave Desert	148
R. E. Reynolds, D. M. Miller, M. O. Woodburne, and L. B. Albright	
Miocene and Late Pleistocene stickleback spines from the Mojave Desert, California	162
Michael A. Bell and Robert E. Reynolds	
75 years of fieldwork in the Barstow Formation by the Raymond Alf Museum of Paleontology	169
Don Lofgren and Rajsavi Anand	
Paleomagnetism of Miocene volcanic rocks in the Newberry Mountains, California: vertical-axis rotation and a polarity transition	177
John W. Hillhouse, Ray E. Wells, and Brett F. Cox	
Geology and tectonic development of Alvord Mountain, Mojave Desert, CA	196
John H. Foster, Ph.D.	
Camel tracks from the Early Miocene Hector Formation, Cady Mountains, California	206
Robert E. Reynolds	
The Palm Borate Company operations, Calico	209
Larry M. Vredenburg	

Geology and ore genesis of epithermal silver–barite mineralization in the central Mojave Desert, California	213
David R. Jessey	
Biota of the Willow Tank Formation, late Early Cretaceous (Albian) from southern Nevada	224
Joshua W. Bonde, David J. Varricchio, and Frankie D. Jackson	
New marine sites from Member 4D of the Goler Formation of California	232
Donald L. Lofgren, Brent A. Silver, Theodora A. Hinkle, and Faizan Ali-Khan	
A juvenile specimen of <i>Archaeohippus mourningi</i> (Perissodactyla: Equidae) from the Cajon Valley Formation (Middle Miocene) of California	241
Darrin C. Pagnac and Robert E. Reynolds	
Early Pleistocene range extension for <i>Microtus mexicanus</i> in the southwestern Mojave Desert	248
Robert E. Reynolds	
The first tungsten-mining boom at Atolia, as recorded in the Randsburg Miner, 1904–1907	253
Alan Hensher, compiler	
Thermal infrared airborne hyperspectral detection of ammonia venting on the Calipatria fault in the Salton Sea Geothermal Field, Imperial County	273
David K. Lynch, David M. Tratt, Kerry N. Buckland, Jeffrey L. Hall, Brian P. Kasper, Michael G. Martino, Luis J. Ortega, Karl Westberg, Stephen J. Young, and Patrick D. Johnson	
Abstracts of proceedings: the 2010 Desert Symposium	280
Robert E. Reynolds, compiler	
Desert preservation: when is enough enough?	280
Richard Crowe	
Geochronology and paleoenvironment of pluvial Harper Lake, Mojave Desert, California	281
Anna L. Garcia, Jeffrey R. Knott, Jordon Bright, Shannon Mahan	
Coalescent analysis of fifteen nuclear loci reveals low genetic diversity and Pleistocene speciation in the Mojave Fringe-toed Lizard, <i>Uma scoparia</i>	282
Andrew Gottscho	
Water on the Moon: how we know and what it means	282
David E. Harker	
Expanding the late Oligocene/early Miocene tectonic, magmatic and sedimentary history in the South Bristol Mountains	283
Janet Harvey, Joann Stock, and David Miller	
The effects of drought on host plant canopy condition and survival of the endangered <i>Astragalus jaegerianus</i> (Fabaceae)	284
T. R. Huggins, B. A. Prigge, M. R. Sharifi, and P. W. Rundel	
Pliocene and Pleistocene lakes of Death Valley, California	284
Jeffrey R. Knott, Joseph C. Liddicoat, Elmira Wan, David Wahl, and Andrei M. Sarna-Wojcicki	
Paleomagnetic and radiocarbon record of the Searles Lake Formation (subunits A and ab) at Poison Canyon, San Bernardino County, California	285
Jeffrey R. Knott, Joseph C. Liddicoat, and Robert S. Coe	
Stratigraphy, age, and depositional setting of the Miocene Barstow Formation at Harvard Hill, central Mojave Desert, California	285
Shannon R. Leslie, David M. Miller, Joseph L. Wooden, and Jorge A Vazquez	
A preliminary review of the effects of utility-scale renewable energy development on terrestrial wildlife in the desert southwest with emphasis on the desert tortoise	286
Jeff Lovich	

Aerial surveys using consumer electronics: fast, cheap and best of all: useful!	288
David K. Lynch, Kenneth W. Hudnut, and David S. P. Dearborn	
Comparison of two Pliocene (Blancan) vertebrate fossil assemblages: Panaca Local Fauna (Lincoln County, Nevada) and Hagerman Local Fauna (Twin Falls County, Idaho)	288
Vicki Meyers and Stephen Rowland	
Reconnaissance geochronology of Miocene basins in the central Mojave Desert; implications for basin evolution and tectonics	289
D. M. Miller, S. R. Leslie, J. W. Hillhouse, J. L. Wooden, J. A. Vazquez, and R.E. Reynolds	
Physiological responses of Mojave Desert shrubs to simulated summer stream channel flow	289
April R. Newlander, David R. Bedford, David M. Miller, and Darren R. Sandquist	
Factors affecting recruitment of desert holly (<i>Atriplex hymenelytra</i>) in the Mojave National Preserve	290
Odara E. Pineda	
A half-million years of paleoclimate and effects on landscape, Mojave Desert, California	290
Marith Reheis and David Miller	
Trackways of a gregarious, Early Jurassic therapsid, Aztec Sandstone, Valley of Fire State Park, southern Nevada	291
Stephen M. Rowland	
Late Holocene Lakes and Horticulture in the Mojave Sink	291
Claude N. Warren	
Rock hounds and mining claims	291
Greg Wilkerson	
Fast-tracking solar development in the desert	292
Howard Wilshire	

2010 Desert Symposium Field Trip Eastern Barstow Basin



Overboard in the Mojave: the field guide

Robert E. Reynolds¹ and David M. Miller²
¹Redlands CA 92373, rreynolds220@verizon.net
²U.S. Geological Survey, Menlo Park CA

Before we start

Each of the three days on this trip starts at the Desert Studies Center at Zzyzx, California. Be certain you have a full tank of gas at the start of each day—services are available in Baker.

Discussion stops on this trip are numbered with respect to theme: Pleistocene (**Q**), Pliocene (**P**), and Miocene (**M**).

Stops labeled **Q** are in Lake Manix sediments and discuss its Pleistocene history, depositional record and tectonic deformation.

Stops labeled **M** are in Miocene sediments presumed to represent the Barstow Formation, which has an extended depositional record during the early Miocene.

Stops labeled **P** exhibit white carbonate deposits that represent ground water discharge (GWD) or marsh land along the Mojave River Valley.

Day 1

What we will see: early on Day 1, we will visit outcrops of late Pleistocene (~500 to 24 ka) Lake Manix sediments. Our goal is to examine the lake sediments in their geomorphic context, in order to describe depositional environments of several distinctive lacustrine facies. Some of these facies are somewhat unusual in that they are not recognized in other paleolakes. In particular, interaction between alluvial fan sediments and the lake produce cryptic lacustrine sediments that look very similar to alluvial fan sediments. Combining depositional facies and dated packages of sediment help describe the events during the evolution of Lake Manix. The lacustrine facies and transgressive/regressive sequences can be compared to facies that we will examine in the Miocene Barstow Formation.

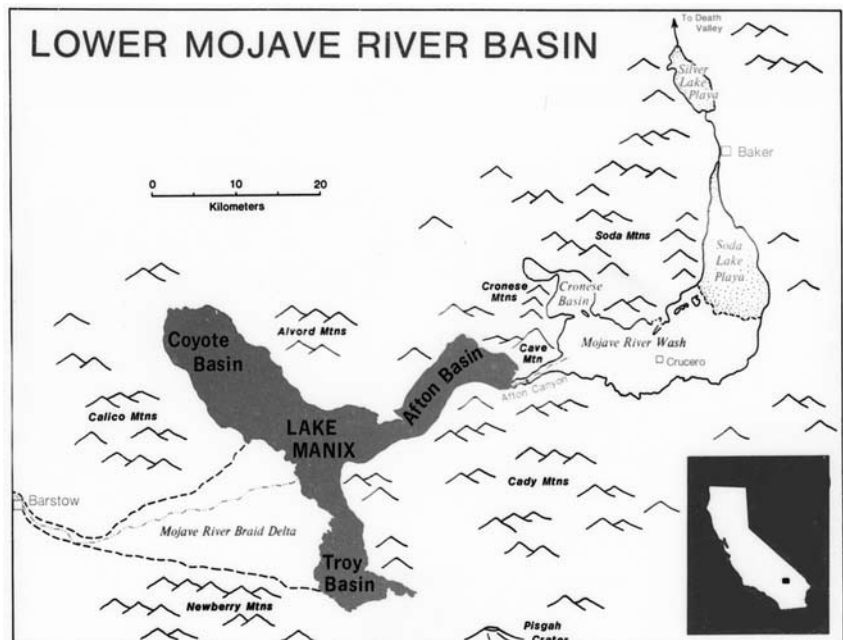
Late on Day 1 we will visit newly dated lacustrine deposits of the Miocene Barstow Formation (~20 to 13 Ma), using zircons from tuffs (dated by U-Pb) and a widespread marker unit, the Peach Spring Tuff, to establish age controls on the sediment. The deposits illustrate depositional environments at

different times, and mostly near the margin of the evolving Barstow Basin.

Convene at Zzyzx Desert Studies Center; proceed north to I-15.

0.0 (0.0) Enter I-15 westbound. Re-set odometer.

5.9 (5.9) Pass Razor Road interchange.



Lake Manix filled the lower Mojave River basin during the Late Pleistocene. (adapted from a graphic courtesy Norman Meek).

9.7 (3.8) Continue past the Basin Road interchange. East Cronese playa, visible to the north, is fed by the Mojave River during floods. It has an extensive record of past lakes and human occupations (Schneider, 1989; Warren and Schneider, 2000). A recent paper (Miller and others, this volume) demonstrates late Holocene lakes present during the Little Ice Age (~AD 1650) and Medieval Warm Period (~AD 1290).

17.3 (7.6) Pass through faulted Miocene sediments in road cut. View west into Afton Basin which was filled by Lake Manix in late Pleistocene time.

18.3 (1.0) Exit I-15 at Afton Road.

18.6 (0.3) Stop at Afton Road, TURN RIGHT (north).

18.8 (0.5) PARK north of frontage road, on north side of off ramp.

STOP Q-1—Afton Barrier Beach. Cuts along the freeway and in drainage channels show that lagoonal deposits capped by weak soils are overlain by beach gravels, illustrating multiple lake occupations at this elevation. Lake Manix reached highstands at or just below 543 masl (1781 ft) at least three times during the late Pleistocene, as suggested by Meek (1990, 2000). Highstands at this elevation are marked by well-preserved constructional beach barriers. Most such barriers are flat-topped with sloping, gently rilled flanks. Retrace and PROCEED WEST along freeway frontage road.

19.3 (0.5) **STOP Q-2—Stromatolites.** PARK. Cuts along the road show breaks in lake deposits, one prominently decorated by a row of stromatolite-capped pebbles, indicating precipitation of calcium carbonate in clear, shallow, sun-lit lake water. CONTINUE SOUTHWEST on paved highway frontage road.

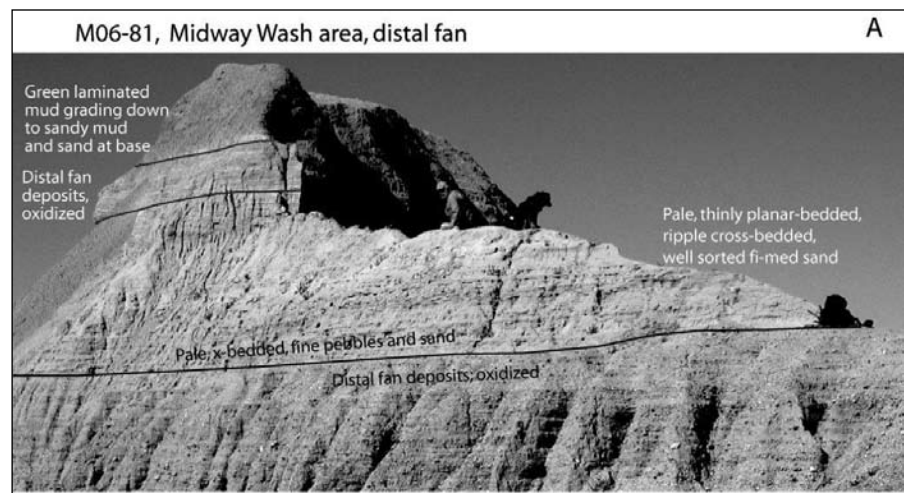
20.1 (0.8) Pavement turns sharply right.

20.2 (0.1) End of pavement.

20.3 (0.1) TURN LEFT on pipeline road and BEAR RIGHT (north).

20.5 (0.2) PARK in quarry. Caution! Do not enter the deep gully at the wash.

STOP Q-3—Dunn Wash Gravel Quarry. “Dunn” wash exhibits lake expression in alluvial fan sediments (GPS coordinates are 551615 E, 3880169 N, Reheis and others,



Stop Q-3. Medial distal fan deposits complexly interfinger with lacustrine sands and muds. M. Reheis photo.

2007). The following is summarized from the report in Reheis and Redwine (2008; Figs. 2 and 4), which contains detailed description and measured sections.

At the Dunn Wash sites, at least four lake fluctuations are recorded. Dunn Wash was an active drainage during these fluctuations, and therefore lacustrine and alluvial sediments are interbedded. These two depositional environments are very difficult to distinguish in places where alluvial gravel was only slightly reworked during a subsequent lake-level rise.

The oldest lake unit (unit 1 in section M05-19, fig. 4 of Reheis and Redwine, 2008) has basal lacustrine gravel with thick tufa coats on clasts, overlain by green mud, silt, and sand that coarsens and thins shoreward. The tufa-coated gravel persists to an altitude of at least 539 masl. This unit is overlain by alluvial gravel of unit 2 and a buried soil with a Btk horizon. The buried soil is overlain by three packages of beach gravel and sand (units 3, 4, and 5) that rise and thin shoreward, typically with tufa-coated clasts at the base of each package and separated by weak soils or alluvial units. Lake units 4 and 5 can be traced to an altitude of about 543 masl. Anodonta shells near the base of unit 1 yielded a finite but minimum-limiting age of $49,800 \pm 2,000$ 14C yr B.P. (table 1 of Reheis and Redwine, 2008). A single preliminary U-series age on the basal tufa is 71 ± 25 ka (J. Paces, U.S. Geological Survey, written commun. to Miller, 2007). Lake unit 3 is only locally preserved and may represent a minor fluctuation. Anodonta shells within unit 4 below a weak buried soil, yielded ages of $34,680 \pm 260$ and $31,900 \pm 200$ 14C yr B.P. At two other sites the uppermost unit 5 contained Anodonta shells that yielded ages of $25,420 \pm 120$ and $26,030 \pm 100$ 14C yr B.P.

The gravels here are generally similar to alluvial fan gravels, but the presence of Anodonta and lateral tracing

to distinctive lake beds indicates that part of the section represents gravels reworked at the lake margin. This facies is common at several places in Lake Manix (Reheis and Miller, this volume). RETRACE toward Afton Road.

20.8 (0.3) TURN RIGHT (south) on graded road.

23.1 (2.3) Stop at Afton Road. TURN RIGHT (south) over I-15.

23.3 (0.2) TURN RIGHT (west) on Dunn Road.

24.5 (1.2) Dip. Prepare to turn left (south) at a white talc pile on the left.

24.9 (0.4) TURN LEFT (south) on wide dirt road.

25.3 (0.4) Cross under powerline.

25.5 (0.2) Continue past a cabin.

25.6 (0.1) PARK in broad cleared area of former railroad station (Reheis and others, 2007)..

STOP Q-4—New Dunn. Lake Manix originally was one cohesive basin that lay west of Buwalda Ridge. The lake threshold near Buwalda Ridge failed midway through the lake history, ~185 ka, flooding the Afton arm. This railroad cut is in the oldest lake deposits of the Afton arm of Lake Manix. The Manix ash helps determine when the Afton Arm integrated with the rest of Lake Manix.

The railroad cut (552284 E, 3878805 N, 478 masl; 1568 ft) exposes, from top to bottom, a thin fan gravel overlying ~4.5 m of well-sorted sands exhibiting long sweeping crossbeds indicating east-directed flow. The quartz- and feldspar-rich sands are similar to modern Mojave River sand. Green muddy sand forms a few thin (centimeters) layers in the upper part of the exposure on the south side of the cut; although these muds contained no ostracodes, deposits in similar stratigraphic positions contain *Limnocythere ceriotuberosa* (J. Bright, Northern Arizona University, written commun. to Reheis, 2006) typical of Lake Manix deposits. The basal sand contains carbonate-cemented layers and nodules that resemble root casts suggesting soil profiles. The sand overlies poorly sorted, indurated, distal alluvial-fan deposits, and represents a valley-axis facies, the first lake to occupy the Afton sub-basin. In other locations to the south and west, this sand is thicker and contains thin interbeds of distal-fan deposits, lacking soils, suggesting brief fluctuations in lake level. To the southwest, along the Mojave River, this same sand directly and conformably overlies fluvial gravel and sand, with clasts of dominantly mafic volcanic rocks probably derived from older fan deposits upstream near Buwalda Ridge. In turn, these fluvial deposits overlie the “Mayhem terrane,” a zone of chaotically mixed and plastically de-

formed blocks of Lake Manix muds and sands mixed with locally derived playa muds formed by flooding when the waters from the lake basin cut through the threshold at Buwalda Ridge. Strong currents evidently carried gravels of the threshold as well as former bottom sediment of Lake Manix down a channel and dumped them near the depocenter of the Afton arm. The Manix ash, ~185 ka, has been found in one place in the Afton sub-basin within these well-sorted sands; it provides a temporal tie to deposits of the main lake, showing that the integration of the Afton arm took place shortly before 185 ka. The ash is well exposed at stop Q-5 and will be examined later today. Retrace to paved Dunn Road and Afton Canyon onramp.

26.3 (0.7) Stop at Dunn Road. TURN RIGHT (east).

27.9 (1.6) Stop at Afton Canyon Road. TURN LEFT (north) across I-15.

28.1 (0.2) TURN LEFT (west) and enter westbound I-15.

31.9 (3.8) Continue past the rest stop.

35.9 (4.0) Continue past Field Road and east-dipping Pliocene gravels.

40.0 (4.1) Pass under the Alvord Mountain overpass.

41.6 (1.6) Continue past the southeast arm of Lime Hill, with east-dipping Pliocene gravels.

42.6 (1.0) EXIT at Harvard Road.

42.9 (0.3) Stop at Harvard Road. TURN LEFT (south) onto Harvard Road.

43.2 (0.3) Stop at Yermo Road. TURN LEFT (east) on Yermo Road.

46.4 (3.2) Manix. TURN RIGHT (south) across tracks. The central road winds southeast to Vortac. SLOW through curves.

46.7 (0.3) BEAR RIGHT (south) at road junction.

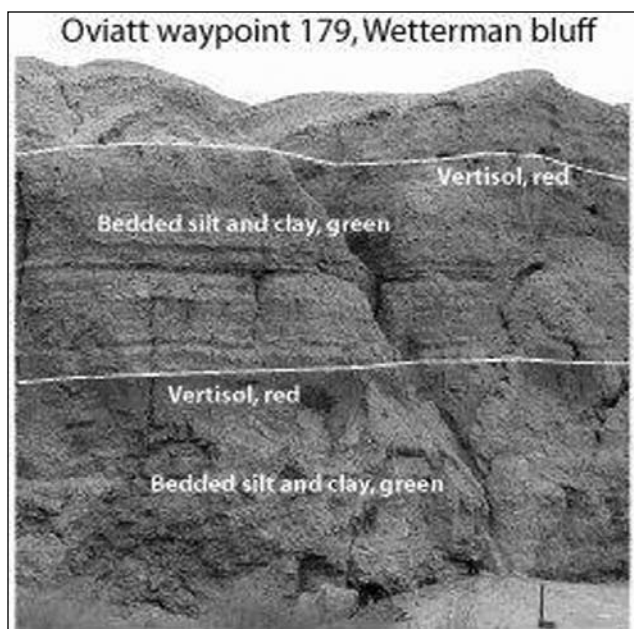
47.6 (0.9) Slow through curves. The road bears southeast toward Vortac.

48.3 (0.7) The road forks at Vortac; BEAR LEFT.

48.5 (0.2) TURN LEFT (east) on graded road.

49.6 (1.1) PARK near view of Lake Manix badlands.

STOP Q-5—Wetterman (Blaire) Ranch. Discuss depositional environments at this location (Figure 5 in Reheis and Miller, this volume). Walk down through the badlands to examine sediments deposited in deep lake, alluvial tongues into lake, mudflat, and fluviodeltaic environments. Paleosols provide records of local lake drying.



Stop Q-5. Wetterman Ranch Bluffs. Two red soil profiles developed on two sequences of lacustrine silts and clays between lake filling events. M. Reheis photo.

The white, fine-grained, water-laid Manix ash lies about mid-section here. These exposures will be compared to sediment recovered in a long core taken a few hundred meters to the north. RETRACE to Vortac and Manix.

50.7 (1.1) BEAR RIGHT (northwest) toward Vortac.

51.0 (0.3) Continue past Vortac.

51.3 (0.3) Slow through curves.

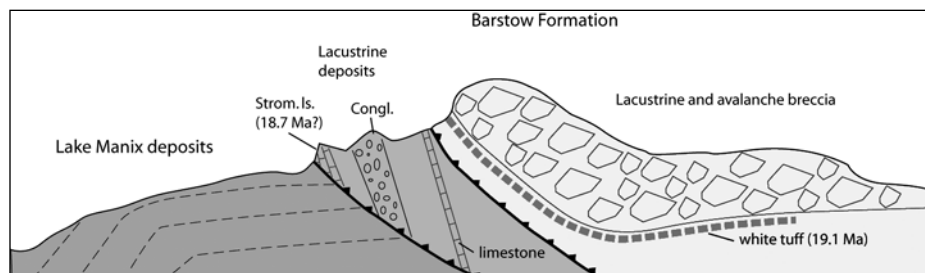
52.4 (1.1) Slow through curves.

52.8 (0.4) Cross railroad tracks and stop at Yermo Road. TURN LEFT (west).

54.6 (1.8) Continue past the southern ridge of Lime Hill. Note the southeast-dipping Pliocene gravels (Byers, 1960).

55.0 (0.4) Continue past a ridge supported by indurated round-cobble conglomerate (CRC, Reynolds and others, this volume).

56.0 (1.0) TURN LEFT (south) at Harvard Road.



Stop Q-6. South Harvard Hill. Branches of the Manix Fault thrust Miocene sediments of the Barstow Formation over late Pleistocene Manix Lake sediments.

56.5 (0.6) Continue past Cherokee Road.

60.0 (0.5) TURN RIGHT (west) on Golden Desert (power line) road.

60.8 (0.8) The road bends.

61.1 (0.3) The road bends; prepare for dips.

61.7 (0.6) Continue past deformed Lake Manix sediments.

62.0 (0.3) BEAR RIGHT (north) between legs of double metal tower.

62.1 (0.1) BEAR RIGHT on rutted dirt road heading east-northeast toward mining cuts on Harvard Hill.

62.2 (0.1) Cross north over ruts in road and PARK near mining cuts for stops Q-6 and M-1.

62.7 (0.5) **STOP Q-6—Harvard Hill Mining Cuts.** A Pleistocene thrust fault places Miocene limestone of the Barstow Formation over Pleistocene Lake Manix sediments. Lake Manix sand, beach gravel, and minor muds are exposed in the pale badlands to the west. They represent folded sets of beach gravels interbedded with sands. These beds at their highest point lie at least 50 m above their original level, representing significant uplift in the late Pleistocene and Holocene. Strands of the Manix fault lie to the north and south of the outcrops, and probably intersect with the Dolores Lake fault on the west to cause local compression. The thrust fault over the lake beds probably represents a more impressive manifestation of this same compression.

STOP M-1—South Harvard Hill stratigraphy and structure. WALK SOUTH to the contact between a high ridge of silicified limestone and the sands that lie to the southwest of that ridge, then proceed upsection across the ridge, across another ridge, and to the dark hill on the northeast.

Key observations here are the facing directions (northeast) for Miocene sediments, the ~19.1 Ma age of the white tuff underlying avalanche breccia, and correlation of the limestone. Leslie and others (this volume) argue

that a thrust fault must lie in this section because of older Miocene sediment lying on younger massive Miocene limestone that elsewhere lies on a 18.7 Ma tuff (Stop M-2). The coarse conglomerate and avalanche breccia in the Barstow Formation here indicate proximity to a steep basin margin to



Stop M-2. The Shamrock Tuff at north Harvard Hill lies below thin-bedded to massive lacustrine limestone. The Shamrock Tuff is considered to be a lacustrine facies of the 18.7 Ma Peach Spring Tuff.

the south during the early Miocene. Walk back to vehicles and drive west along access road.

63.2 (0.5) TURN RIGHT (northwest) at the intersection.

63.4 (0.2) TURN RIGHT (north) on the road leading toward north Harvard Hill.

63.9 (0.5) BEAR RIGHT (east).

64.5 (0.6) At the complex intersection, PROCEED EAST 0.1 miles.

64.6 (0.1) PARK and walk southeast.

STOP M-2—North Harvard Hill. Walk southeast to the pale green Shamrock tuff under silicified limestone beds. The Shamrock tuff may correlate with the ~18.7 Ma Peach Spring Tuff (Leslie and others, this volume). The Shamrock tuff shows that the Barstow basin was receiving water and lacustrine silts prior to 18.7 Ma, and that the MSL at Harvard Hill may have been deposited earlier than the 16.7–16.9 MSL in the Mud Hills. North Harvard Hill stratigraphy differs from that at Stop M-1. The northern section contains thin limestone beds interbedded with sands representing reworked Shamrock tuff, overlain by more massive limestone, partly silicified. The south section is virtually entirely massive limestone, strongly silicified. Correlating limestone units such as these is difficult because they change facies rapidly along strike (Leslie and others, this volume). Walk back and RETRACE to complex intersection.

64.7 (0.1) TURN RIGHT (north).

64.9 (0.2) TURN RIGHT (east) on the south side of the railroad tracks.

65.6 (0.7) Stop at Harvard Road, watch for traffic, and TURN RIGHT (south).

66.3 (0.7) Stop at Harvard Road. Check for traffic, then TURN RIGHT (south) on Harvard Road.

67.4 (1.1) Continue past Golden Desert (power line) road.

68.2 (0.8) Cross the Mojave River.

69.7 (1.5) The road turns right (west).

70.7 (1.0) Stop, TURN LEFT (south) on Newberry Road.

73.6 (2.9) Continue past Newberry School. The Newberry Mountains tower ahead to the south. Magnetostratigraphic studies in the Newberry Mountains are discussed herein (Hillhouse and others, this volume).

76.1 (2.5) Cross the railroad tracks and TURN RIGHT (west) on Pioneer Road.

78.1 (2.0) Stop at National Trails Highway. TURN LEFT and then RIGHT onto westbound I-40.

83.9 (5.8) Continue past the ramp to Barstow–Daggett Airport.

88.9 (5.0) EXIT at Daggett Road.

89.2 (0.3) Stop at Daggett Road (A Street) and TURN LEFT (south).

89.6 (0.4) Stop at Pendleton Road on the south side of I-40 and TURN LEFT (east).

90.6 (1.0) Pavement ends. BEAR RIGHT on Camp Rock Road.

91.5 (0.9) Continue past the first powerline road; wooden posts are on the right. Proceed 0.2 miles and TURN RIGHT at yellow metal posts on the south side of third powerline.

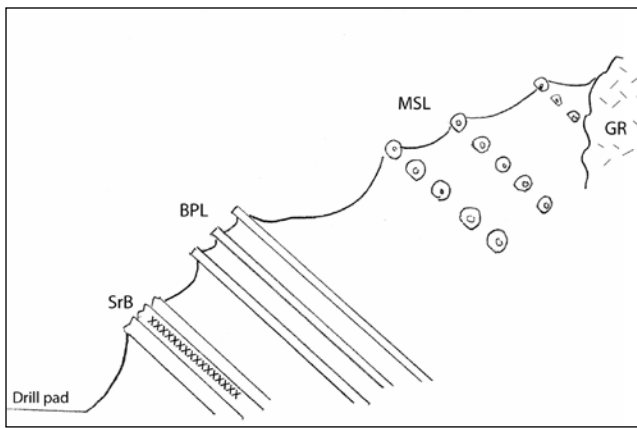
91.7 (0.2) TURN RIGHT at yellow metal posts. Proceed 0.5 miles west-southwest along the powerline road.

92.2 (0.5) TURN LEFT (south) at Tower #449.2 (marked on its southeast leg).

93.8 (1.6) Continue past a right turn to the west.

94.2 (0.4) Continue past the right fork and proceed south to the Columbus/Gem Mine.

94.4 (0.2) BEAR LEFT (east) and PARK on drill pad.



Stop M-3. Marker beds at the Columbus/Gem mine, overturned stratigraphy, view east. SrB: strontium and borate beds; BPL: brown platy limestone; MSL: massive stromatolitic limestone.

STOP M-3—Columbus/Gem Mine. This overturned section of Barstow Formation shows, in reverse stratigraphic order, pink 18.7 Ma tuff (PST), strontium–borate layer (SRB), three brown platy limestones (BPL), thick granitic fanglomerate, and three strata of massive cylindrical stromatolites (MSL), the stratigraphic lowest (now the uppermost) of which is silicified. Farther south, across a branch of the Camp Rock Fault, are aplitic granite outcrops that, when uplifted in the Daggett Ridge event, contributed to the granitic debris now found elsewhere in the lower part of the stratigraphic section. Zircons from the tuff at this location are ~18.7 Ma, indicating correlation with the Peach Spring Tuff (Miller and others, this volume). This suggests that the entire ~0.8 Ma duration marker sequence at this locality is older than 18.7 Ma and started being deposited around 19.5 Ma. Volcaniclastic sediments between the brown platy limestones look like their source was the Pickhandle Formation.

Stoddard Valley to the south has a similar, but abbreviated, section of marker beds, with large tubular stromatolites representing the MSL, a single BPL layer and the PST (Wells and Hillhouse, 1989; Miller and others, this volume). The SrB horizon is apparently not present in Stoddard Valley. RETRACE toward powerline road.

94.8 (0.5) TURN LEFT (west) on dirt track.

95.3 (0.5) TURN LEFT (south) to avoid gully.

96.2 (0.9) BEAR RIGHT (west) at Y turn.

97.7 (1.5) PARK.

STOP M-4—West Gem Exposures. Look east at colorful badlands exposures which in-

clude a pink tuff similar to the tuff we saw at the last stop. This pink marker tuff is useful in tracing this portion of the stratigraphic section westward. The pink tuff sampled at this locality has the same remnant magnetism as the PST, suggesting a correlation (Hillhouse and others, this volume). Here, unlike the section at Columbus/Gem, a thin section of lacustrine beds lie above the pink tuff. The marker sequence is not present here, apparently due to the influx of granitic-sourced gravels that have thin lacustrine beds within them. A Pleistocene analog can be seen in Lake Manix alluvial gravel interbedded with lacustrine silts (Miller and Reheis, this volume).

Central Daggett Ridge is visible to the west, past Ord Mountain Road. That portion of the Miocene sedimentary section is apparently separated from this portion by a north–south fault, as the pink tuff is not present. The central Daggett Ridge section contains the marker sequence (MSL, BPL and SrB in drill records of US Borax). The Daggett Ridge P/L locality (SBCM 1-109-2; Reynolds, 1991) has a fauna that contains small and large horses suggesting the Hemingfordian/Barstovian transition at 16 Ma, similar to the faunal transition in the Toomey Hills (next stop). Proceed northwest to power line road.

98.0 (0.3) TURN RIGHT (north) on the section line road toward the power line road.

99.0 (1.0) Stop at the power line road. TURN RIGHT (northeast) on power line road and retrace to Camp Rock Road.

101.5 (2.5) Stop at Camp Rock Road. TURN LEFT (north) and proceed to I-40.

103.5 (2.0) TURN RIGHT (north) on “A” Street and cross I-40, entering the town of Daggett. The arrival of the



Stop M-4. West Gem locality. The pink tuff exposed here appears similar to the friable, pink Peach Spring Tuff at the Columbus/Gem Mine (Stop M-3). The Tuff is overlain by a thin sequence of lacustrine sediments.

Atlantic and Pacific Railroad at Waterman's (northwest Barstow) in 1882 improved shipping of colemanite from Borate in the Calico Mountains and from Death Valley (Myrick, 1992). Colemanite was offloaded at Daggett, then shipped to Alameda, California. Borax came to this railhead from Death Valley via the 1883 Saratoga Springs–Cave Springs–Garlic Spring route and via the Wingate Wash–Pilot Knob route about the same time. Borax freighting from Death Valley moved to a westerly route via Pilot Knob/Blackwater Well to Mojave between 1884 and 1888. Borate minerals from the Calico Mountains were freighted to Daggett until 1900 (Myrick, 1992).

104.2 (0.7) Cross National Trails Highway; cross railroad tracks.

104.7 (0.5) Cross the Mojave River.

105.1 (0.4) Continue past the Waterloo Mill site the to left (west) at the base of Elephant Mountain. The original 1884 mill was owned by the Oro Grande Mining Company, and boasted 75 stamps when it was purchased by the Waterloo Mining Co. in 1889. A narrow gauge railroad connected the Atlantic and Pacific Railroad (later Santa Fe) at Daggett to the mill on the north side of the river,

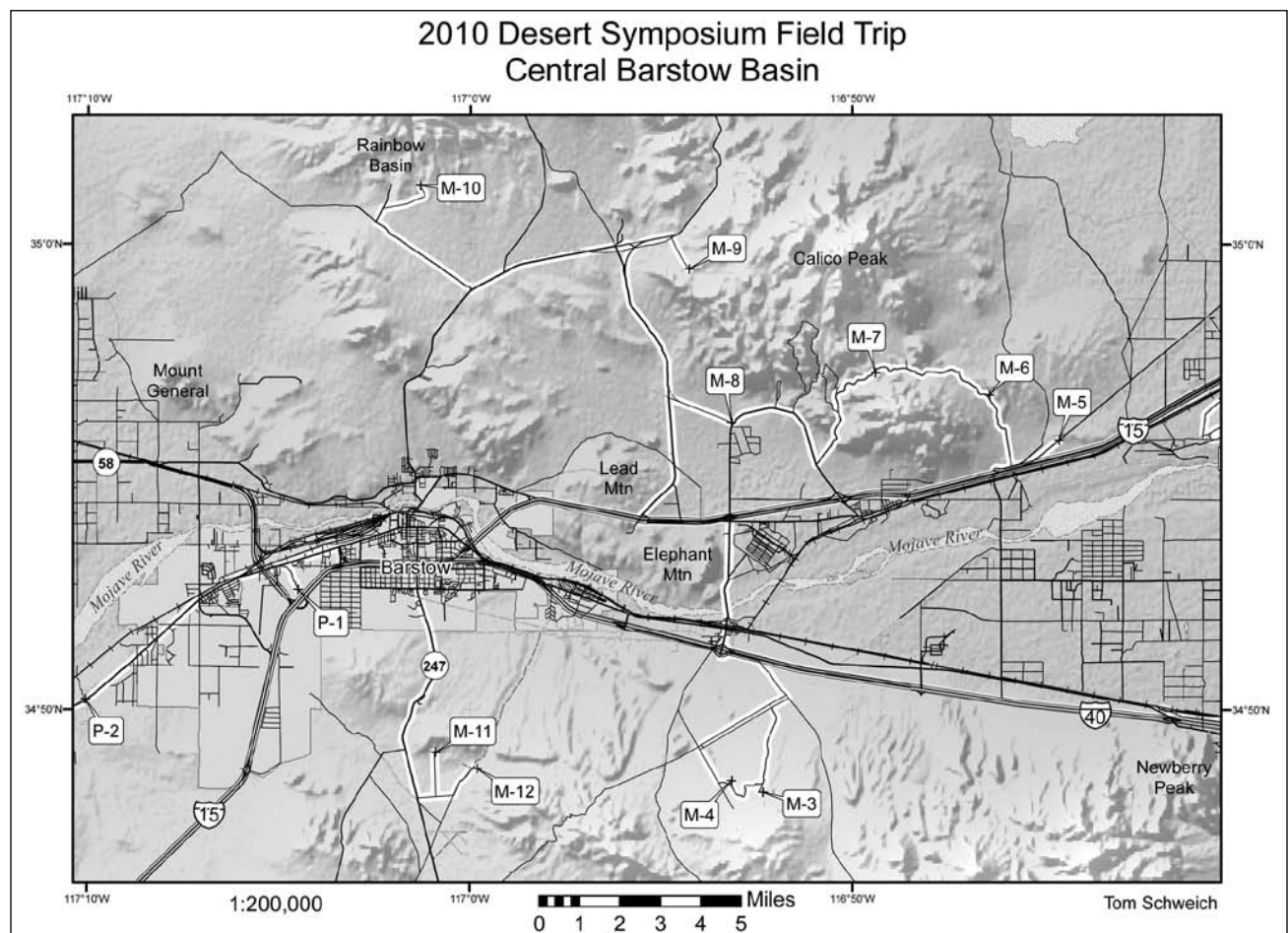
to the Waterloo and Silver King Mines at Calico, and to the narrow gauge railroad that led to the American Borax Company's Columbia Mine (not Columbus) on the north side of Lead Mountain (Myrick, 1992) .

106.8 (1.7) Stop at Yermo Road. Gas and snacks available. Proceed a short distance north on Ghost Town Road and TURN RIGHT (east) onto I-15 eastbound. Return to the Desert Studies Center at Zzyzx.

End of Day 1

Day 2

What we will see: On Day 2 we will visit outcrops of lake deposits in the Barstow Formation at the Toomey and Yermo hills, at the Calico Mountains, and near its type section in the Mud Hills. New dates support previously determined ages based on fossil mammals, and clarify stratigraphic and depositional relationships. The ages and sedimentary facies in the Yermo Hills–Calico Mountains area will be compared with deposits of the Barstow Formation that are of similar age, but somewhat different facies, at Owl Canyon in the Mud Hills. The Barstow Formation at the Calico Mountains was an important source



of borate minerals during the very early 20th century and ruins of mines, railroads, and dwellings testify to the importance of the mining industry at this location.

CONVENE at the Desert Studies Center with a full tank of gas. Enter westbound I-15 toward Minneola Road.

0.0 (0.0) EXIT at Minneola Road. Re-set odometer.

0.3 (0.3) Stop at Minneola Road. TURN RIGHT (north). The road will bear east.

0.9 (0.6) Continue past a left turn to the landfill and the Calico Mountains Archaeological Site .

1.1 (0.2) BEAR RIGHT. When pavement ends, take the middle power line access road and bear northeast.

1.3 (0.2) Avoid the left turn.

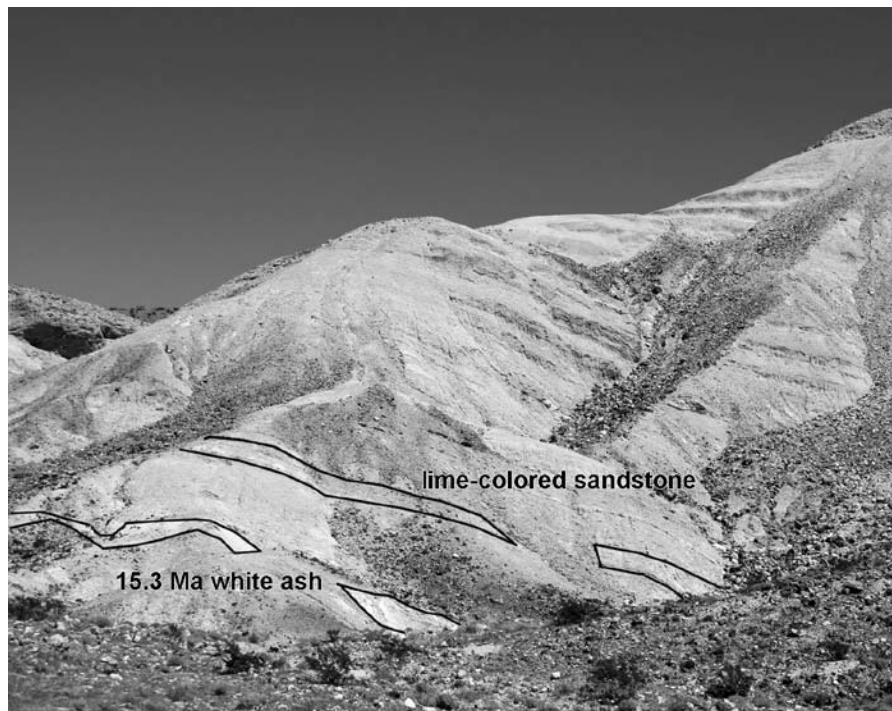
1.4 (0.1) BEAR LEFT on power line road.

2.3 (0.9) **STOP M-5—Toomey Hills.** PARK in open area north of road. The Barstow Formation is exposed on the north side of the Manix fault, as it is at Harvard Hill. Near the base of the section are fluvial gravels with well-rounded clasts and tubular stromatolites (CRC, Reynolds and others, this volume) in a section of sandstone cemented by calcite. Above are fine-grained lacustrine sandy silt beds interlaced with tuffs. This outcrop produced fossil mammals which document the 16 Ma transition from the Hemingfordian to Barstovian NALMA (Woodburne, 1991; Woodburne and Reynolds, this volume; Tedford and others, 2004). A local sequence of beds (CRC, Emerald Tuff, Lime Tuff) allow the Toomey sequence to be correlated with a datable section in the Yermo Hills to the northwest (Miller and others this volume). The crest of the ridge here is a thick, strongly cemented soil, above which is the Pliocene–Pleistocene Yermo Gravel. RE-TRACE toward Minneola Road.

3.6 (1.3) TURN RIGHT (north) at road to the county landfill and Calico Mountains Archaeological Site.

4.8 (1.2) TURN LEFT (west) away from entrance to the Calico site and landfill.

5.6 (0.8) TURN RIGHT (north) into Emerald Basin.



Stop M-6. Yermo Hills, showing lime-green sandstone and 15.3 Ma ash layer.

6.0 (0.3) PARK in open area, avoiding nails from burnt wood.

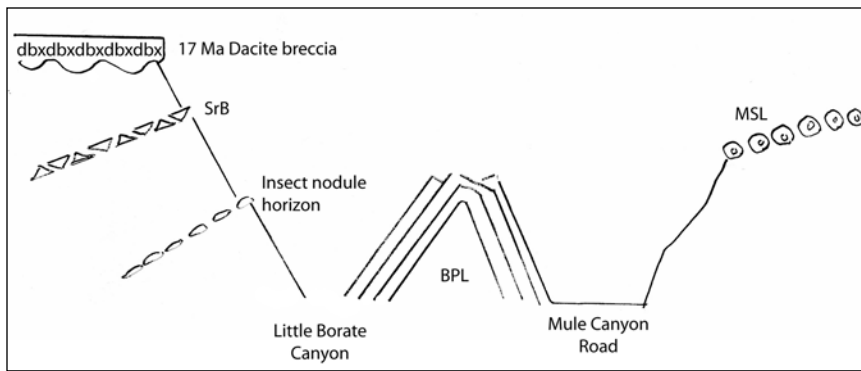
STOP M-6—Overview of Emerald Basin in the Yermo Hills. This section contains a sequence of beds (CRC, Emerald Tuff, Lime Tuff) similar to the Toomey Hills. A dated tuff near the top of the section is about 15.3 Ma (Miller and others, this volume). If the CRC here is equivalent to the CRC at the 16+ Ma Toomey Hills stop, then the gray siltstone section represents at least 0.7 Ma. A thick, silicified limestone on the ridge line can be traced westward into the eastern Calico Mountains where it overlies the 16.2+ Ma BPL. Retrace to main dirt road and proceed northwest.

6.1 (0.2) TURN RIGHT (northwest) onto main dirt road and proceed northwest.

7.0 (0.9) Join the pole line road and proceed north. This road coincides with the approximate position of the north-trending Tin Can Alley fault, described and named by Dudash (2006). It is a right-lateral fault that in places exhibits scarps cutting Holocene materials.

7.2 (0.2) TURN LEFT (west) onto Mule Canyon Road.

8.3 (1.1) Continue past the Sulfur Hole quarry in the BPL which produced an unusual assemblage of hydrous iron sulfate minerals oxide products from pyrite deposited by hot springs (Dunning and Copper, 2002; Schuil-ling, 1999).



Stop M-7. View southwest at Little Borate Canyon. Massive stromatolitic limestone (MSL) to the north. Look southwest at the chevron folds of brown platy limestone (BPL). The historic mines containing borate and strontium minerals (SrB) are upsection on the south side of Little Borate Canyon. The nodular stromatolites containing silicified insect fossils are in a horizon below the SRM (Reynolds and others, this volume). To the south, the 17 Ma dacite breccias (Singleton and Gahns, 2008) unconformably overlie the compressionally folded 0.8 Ma-duration marker bed sequence (Reynolds and others, this volume).



M-7. Little Borate nodule horizon contains silicified insects, including midges showing detailed preservation of compound eyes. SEM image courtesy of John Jenkins (1986).

- 8.5 (0.2) Continue past Big Borate Canyon (Reynolds, 1999b) on the south.
- 8.6 (0.1) Continue past Tin Can Alley on the north.
- 9.1 (0.5) Continue past the entrance to Little Borate Canyon (Trestle Canyon) and the site of Happy Hollow (Reynolds, 1999b). The Barstow Formation section here consists of MSL on the north side of road and BPL and SrB on south side of road. The BPL nearby contains the oldest proboscidean (gomphothere) tracks known from the North American continent (Reynolds 1999a; Reynolds and Woodburne, 2001, 2002). Proceed uphill.
- 9.7 (0.6) PULL LEFT (south) into the turnout at the top of the grade and PARK.

STOP M-7—Overview of Little Borate. We are at the point where the narrow gauge Borate and Daggett Railroad (Myrick, 1991; Hildebrand, 1982) separated from Mule Canyon Road. The marker section here contains the MSL, BPL, and SrB. This section of the Barstow Formation lies on upper Pickhandle Formation beds dated at 19.0 Ma (Singleton and Gans, 2005, 2008). They considered andesite breccia (dated at 16.9 Ma) to cap the Barstow section; however, mapping by Sadler (p. c. to RER, 2010) indicates that these volcanic breccias interfinger with a section of Barstow Formation that ranges in age from 16 to 14.8 Ma (Reynolds and others, this volume). Proceed

westerly downhill on Mule Canyon Road.

- 10.9 (1.2) Continue past the Phillips Drive exit.
- 11.4 (0.5) Camp Rock and the entrance to Phillips Drive. Camp Rock was the halfway point on the two-day mule team run from Borate to Daggett (Reynolds, 1999b; Hildebrand, 1982). To the north, in the beds of the Barstow Formation, fragments of bird eggs have been recovered from the MSL, which sits on a flat pebble conglomerate suggestive of a shale beach (Leggitt, 2002).
- 12.9 (1.5) Stop at paved Calico Road and TURN RIGHT (north).
- 13.9 (1.0) Continue past Barber's Mill (1880; Weber, 1966, 1967) on the southwest side of the road.
- 14.3 (0.4) Continue past right turn to Doran Drive.
- 14.6 (0.3) Continue past the Garfield millsite for the 1880 Silver King mine (Weber, 1966, 1967) on the right (north).
- 14.8 (0.2) Continue past the entrance to Calico Ghost Town.
- 15.0 (0.2) Continue past Cemetery Ridge on the right.
- 15.8 (0.8) TURN RIGHT on Yermo Cutoff. PULL RIGHT and stop.

STOP M-8— Cemetery Ridge: view northeast. The section west of the Calico cemetery is east of the Burcham Gold Mine (Weber, 1966, 1967). The upper of two white tuffaceous sandstone beds contain zircons that give an interesting age of 19.0 Ma (Miller and others, this volume). The section contains the BPL, and north, across a strand



Stop M-8. Cemetery Ridge. A white tuff (above backpack) provided zircon dates of 19.1 Ma. Resistant beds to the right are brown limey sandstone.

covered by rounded and sub-rounded pebbles suggesting a long period of weathering, therefore, the age of the BPL is later than the weathered dacite. The BPL contains approximately 50% pebbles at its base, but becomes finer upward and has vertically oriented pores that resemble the texture of the tufa at Travertine Point on the west side of the Salton Sea. This five-m-thick section of BPL may represent a lacustrine onlap sequence. RETRACE to Ft. Irwin Road.

24.3 (1.0) Stop at Ft. Irwin Road, TURN LEFT (southwest).

25.1 (0.8) TURN RIGHT (northwest) on Old Ft Irwin Road; proceed west toward Fossil Bed Road.

29.4 (4.3) Continue past Copper City Road.



Stop M-8. Replica of fossil flamingo egg (center) found below Burcham mine, flanked by eggs of lesser flamingo (left) and greater flamingo (right).

30.3 (0.9) TURN RIGHT (west) on Fossil Bed Road.

33.2 (2.9) TURN RIGHT (northwest) on Rainbow Basin Loop Road.

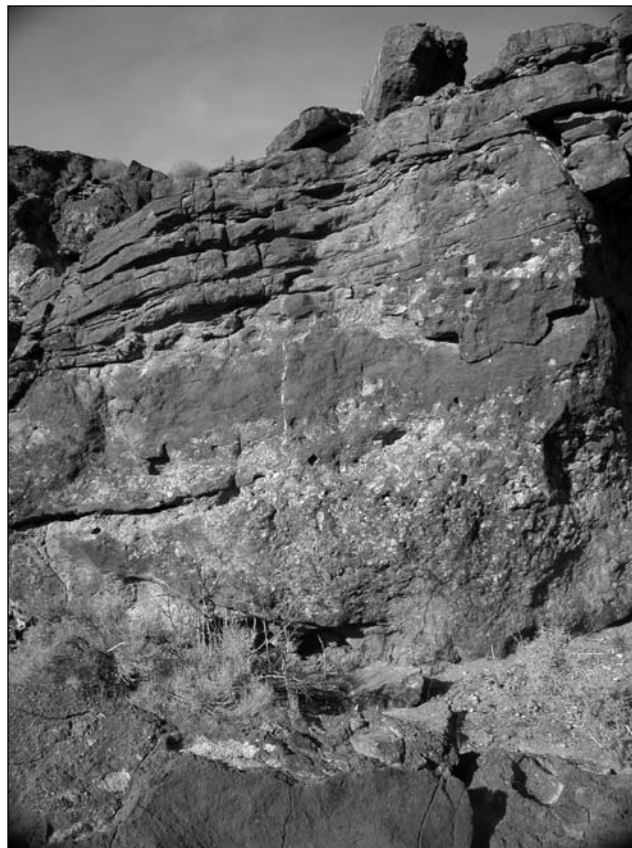
of the Calico Fault, tubular stromatolites of the MSL. The internal mold of a flamingo egg was found below the Burcham Mine, reinforcing the environment of a shallow lake margin (Reynolds, 1998). PROCEED WEST to Ft. Irwin Road.

17.5 (234.0) (1.7) Stop at Ft. Irwin Road; TURN RIGHT (north).

21.5 (4.0) Continue past Old Ft Irwin Road and proceed northeast toward Pickhandle Pass.

22.4 (0.9) Slow, TURN RIGHT (southeast) at the BLM kiosk onto a road to the northwest Calico Mountains. Proceed approximately one mile to the southeast until the road bends and enters narrows.

23.3 (0.9) **STOP M-9—Russ' Mailbox Locality** in northwest Calico Mountains. The thick section of BPL overlies 19Ma Pickhandle dacite. The dacite outcrop is



Stop M-9. Northwestern Calico Mountains. A single brown platy limestone (BPL) at this locality is 5+ meters thick, and is interpreted as a near-shore, on-lapping sequence. The BPL contains sub-rounded dacite pebbles at its base, and becomes platy and spongy textured toward the top.



Stop M-10. Owl Canyon Campground-Space 28. Concentric tufa deposits around plant. View west shows stromatolites (MSL) and two brown platy limestone (BPL) beds. U-Pb dates on stromatolite calcite at the campground are 15.3 Ma. Stromatolites of the MSL were deposited over a 1 my period at different portions of the basin. One mile west, the marker sequence (MSL, BPL, SrB) spans 0.8 Ma, with the MSL dating to approximately 16.8 Ma.

33.5 (0.3) Turn Right (north) on Owl Canyon Road, enter Campground and PARK at space #28.

35.0 (1.5) **STOP M-10—Owl Canyon Campground.** View west across wash of stromatolitic limestone phytoherms (tufa around branching bushes) and two brown platy limestone beds. Cole et al (2005) used U-Pb methods to date branching tufa mounds. Results from three dated localities are 15.39 ± 0.15 Ma and 15.30 ± 0.25 Ma at Owl Canyon Campground, and 16.25 ± 0.25 Ma on the north limb of the Barstow Syncline. Apparently, the MSL was actively depositing over a period of nearly 1 my along the south and north portion of the basin. This would be expected during lacustrine deposition of an expanding basin event in tandem with alluvial fans developing from rising highlands to the north and south. The U-Pb dates may be useful in providing relative ages of lacustrine car-

bonates deposited across different portions of the basin. In summary, dates on tubular stromatolites and phytoherms of the MSL from the Mud Hills span a period from 15.3–16.2 Ma. One mile to the west, the entire marker sequence (MSL, BPL, SrB) spans 0.8 Ma, with the MSL dating to approximately 16.7–16.9 Ma (Reynolds and others, this volume). MSL tubes) RETRACE.

36.8 (1.8) Stop at Fossil Bed Road, TURN LEFT (east). Retrace to Old Ft Irwin Rd.

39.7 (2.9) STOP at Old Ft. Irwin Road. TURN LEFT (northeast) to Ft. Irwin Road. Proceed south to I-15, then east on I-15 to Zzyzx Road and the Desert Studies Center.

End Day 2

Day 3

What we will see: early on Day 3, we will continue to explore outcrops of the Peach Spring Tuff, an 18.7 Ma time marker that has been identified in several sub-basins of the Barstow Formation. We will examine the overlying lacustrine deposits of the Barstow Formation near the west end of Daggett Ridge and contrast depositional facies and ages in various sub-basins in the Daggett Ridge area, using the Peach Spring Tuff as the tie point for all stratigraphic sections.

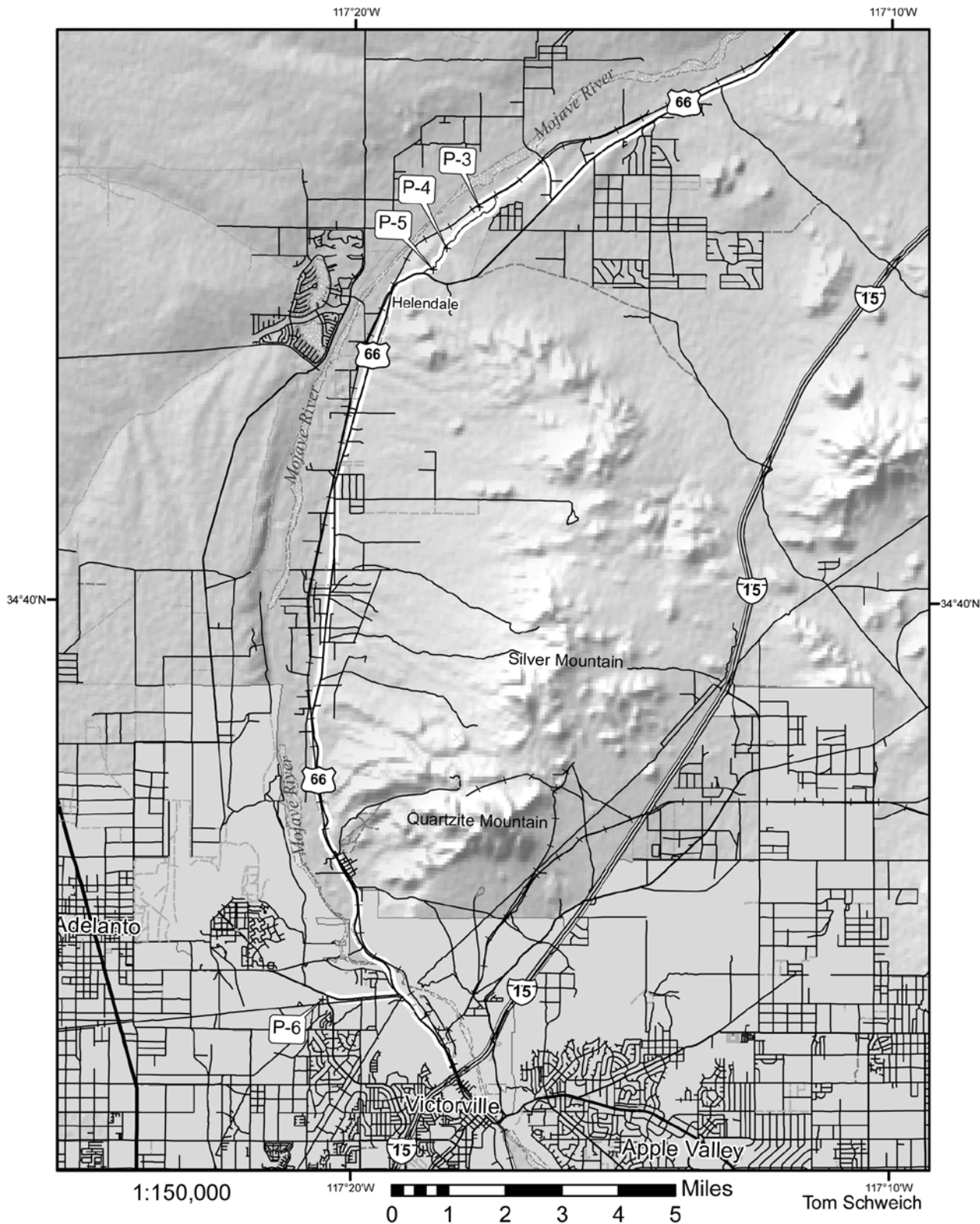
Later on Day 3, we will turn to Pliocene (~5 to 2 Ma) deposits, when instead of lakes there were marshes, small shallow ponds, and wetlands (collectively grouped as groundwater discharge deposits). These deposits are typified by white carbonate-rich sand and silt. We will see that several Pliocene deposits of this type were deposited in an east-trending low area from near Helendale to near Barstow, and an even broader area farther south near Victorville. The deposits at Victorville represent the damming of old drainage systems to form a long-lived marshy basin that eventually was filled with sediment, from which the ancestral Mojave River Valley flowed to Barstow and later to Lake Manix.

CONVENE at the Desert Studies Center with a full tank of gas. Enter westbound I-15 toward Barstow Road. CAUTION: when reaching Barstow, I-40 merges with I-15 near Barstow Road. You must be in the far right lane to exit on Barstow Road.

0.0 (0.0) STOP at Barstow Road. Reset odometer. Services are available to the right.

TURN LEFT (south) on Highway 247 (Barstow Road) toward Lucerne Valley and proceed through several traffic light intersections.

2010 Desert Symposium Field Trip Western Barstow Basin



0.0 (0.3) Continue past Armory Road.

0.6 (0.6) Continue past signal at Rimrock Road.

1.0 (0.4) Continue past Barstow College, BLM headquarters, and National Park Service offices.

2.0 (1.0) Barstow Road (Highway 247) ascends hills composed of Pliocene and Pleistocene gravels, sands, and groundwater discharge (GWD) deposits.

3.4 (1.4) Continue past the summit of Daggett Ridge and the entrance to a landfill. Proceed downhill with caution.

4.8 (1.4) Continue past a right turn (west) to Stoddard Road.

5.9 (1.1) TURN LEFT on unnamed Stoddard Road (graded dirt) section line road.

6.6 (0.6) Continue past a road on the left to Pink Tuff Quarry.

Optional STOP M-11—Pink Tuff Quarry. Drive north 1.4 miles and PARK on low hill south of quarry. The Peach Spring Tuff crops out discontinuously from Outlet Center Drive on I-15 eastward along the north flank of Daggett Ridge to where the Stoddard Cutoff Road runs through Opal Canyon. Where crossed by branches of the Lenwood Fault, the sections of the Peach Spring Tuff (Dibblee, 1968, 1970) appear to be right-laterally offset a total of 3.2 km (2 miles), slightly more than the 2.5 km (1.6 miles) of other estimates (Jachens and others, 2002). Retrace south to Stoddard Road, and proceed east on Stoddard Road.

7.2 (0.6) TURN LEFT (northeast) onto Oxford Road.

8.0 (0.8) PULL RIGHT (east) and PARK in loop.

STOP M-12—Peach Spring Tuff. We are parked just east of a portion of the northwest-trending Lenwood fault. The PST to the east consists of pink non-welded and purple welded facies. Hike east to a vantage point and look north-northeast toward outcrops of the MSL, approximately one-quarter mile northeast. The PST outcrop pattern has an attitude (N 32°W) that suggest concordance with the MSL (N 30°W, 20°E). Hillhouse and others (this volume) describe paleomagnetic results for this PST site; different orientations on either side of a gully indicate a buried fault in that gully. In contrast to the M-3 stop at Columbus/Gem Mine on eastern Daggett Ridge, the marker sequence in Barstow lake beds apparently lies above the 18.7 Ma PST. The Barstow Fm in this area is capped by a calcareous silcrete—a paleosol composed of pedogenic carbonate cemented by silica, a siliceous duricrust). RETRACE route to Highway 247. 8.8 (0.8) TURN RIGHT (west) onto “Gloucester” (Stoddard) Road.

10.1 (1.3) Stop at Highway 247. Watch for traffic. TURN RIGHT (north) and proceed toward Barstow on Main Street.

15.3 (5.2) Continue past signal at Rimrock Road.

15.8 (0.5) Continue past signal at Armory Road on Main Street.

16.0 (0.2) TURN LEFT (west) and ENTER westbound I-15. Stay in the right lane and prepare to exit at “L” Street.

17.5 (1.5) EXIT at “L” Street.

17.9 (0.4) Stop at “L” Street. TURN RIGHT (north) and proceed to Main Street.

18.3 (0.4) Stop at Main Street, TURN LEFT (west) onto Main Street and proceed west.

19.6 (1.3) Continue past Diamond Pacific

19.8 (0.2) Slow, watch for oncoming traffic. TURN LEFT (south) on dirt track (Osborne Road) toward the interchange of I-15 and Highway 58.

20.0 (0.2) TURN LEFT (southeast) into the wash.

20.3 (0.3) **STOP P-1—West Barstow Ground Water Discharge (GWD).** We have stopped near prominent white outcrops of gently north-dipping calcareous silt and sand. The vague bedding, relatively coarse grain size and poor sorting, cementation, and lack of beach facies suggest that the beds primarily represent GWD. To the west, in the freeway road cut and at an outcrop even farther west, Brett Cox recovered tephra from two beds of this unit (Cox and others, 2003). The tephra are identical, and have a poorly established correlation to tephra recovered at Honey Lake, CA, where beds are thought to be latest Pliocene or earliest Pleistocene between 2.0 and 2.5 Ma. A weaker correlation is possible, with a tephra recovered by Jeff Knott at Zabriskie Point; that tephra is bracketed between 3.35 and 3.58 Ma (Elmira Wan, USGS, written commun. to Cox and Miller, 2010). RETRACE to Main Street.

20.8 (0.5) Stop at Main Street, watch for traffic. TURN LEFT (west) toward the Highway 58 underpass.

21.2 (0.4) Pass under Highway 58.

23.0 (1.8) Stop at Lenwood Boulevard. Proceed west.

24.4 (1.4) Continue past Delaney Road.

25.9 (1.5) Continue past “The Dunes” Motel on the left (south). Prepare to turn right (north) in 2.7 miles.

26.6 (0.7) Ahead on the right at the top of the rise is a pole with reflectors. Prepare to turn right (north).

27.1 (0.5) TURN RIGHT (north) at the pole with reflectors (Post # 2 2404—Oak Grove Road).

27.2 (0.1) PARK at utility road.

STOP P-2—Johnston Corners

Pliocene GWD. The sedimentary section at this locality starts at the railroad tracks to the north. It begins with coarse gravels that are angularly overlain by a section of fine-grained sand, very fine gravelly sand, and calcareous sand and silt. Some calcareous beds are strongly cemented, similar to the upper carbonate cemented caps of late Pleistocene GWD (e.g., Forester et al, 2003). The section continues south of the road and is capped on the skyline by a pediment marked by a veneer of boulders. A carbonate-cemented bed nearby contains silica, which yielded an approximate U-Pb age of 2.3 Ma (K. Maher, written commun. to Miller, 2005). This date suggests a correlation with the beds we saw at Stop P-1 in Barstow. Return to vehicles, Retrace south to National Trails Highway.

27.7 (0.4) Stop at National Trails Highway. TURN RIGHT and proceed west.

28.7 (1.0) Continue past Hinkley Road and Hodge.

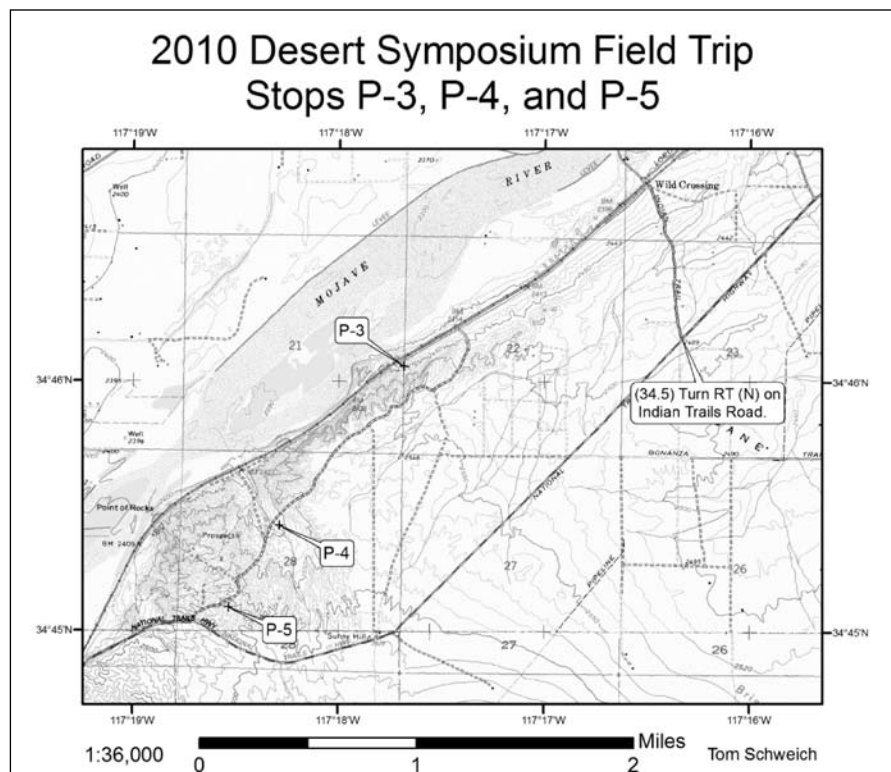
33.6 (5.6) Continue past Holcomb Ranch Road.

34.5 (0.9) TURN RIGHT (north) on Indian Trails Road. Proceed toward railroad.

35.2 (0.7) TURN LEFT (west) on the south railroad frontage road.

36.3 (1.1) Pass through a dip and proceed 0.2 miles to an outcrop on the left (south).

36.5 (0.2) **STOP P-3—Helendale Ash** (Reynolds and Cox, 1999). Helendale Ash sampled here is a Mono “W” type tephra, indicating an age range from 5.0 to 0.7 Ma. It best correlates with a sample from the Ventura section that is between 0.9 and 1.18 Ma. The Helendale Ash is located at the base of the Mojave River gravels (2500 ft elev.), and overlies the fine sediments of the Victorville Basin. This sequence suggests that the river had finished filling the Victorville Basin by ~1 Ma, and then began carrying coarse sands farther east. The elevation of this ~1 Ma ash helps constrain the section that we will see at



Stops P-4 and P-5. RETRACE east along the south side of the railroad for 0.2 miles.

36.7 (0.2) TURN RIGHT (south) through a break in the fence and follow OHV tracks in wash to reach terrace top.

37.2 (0.5) Complex intersection. TURN NORTH, then bear west along bluff road. Do not turn south toward houses or pole line road.

37.5 (0.3) Proceed west, passing a left (south) turn. Note lag gravels on the terrace top.

37.9 (0.4) Bluff Road drops off terrace top (elevation 2600 ft) into broad canyon.

38.1 (0.2) **Stop P-4—Helendale Section** (2480 ft elevation) in wash bottom. Note red paleosols, and green pond sediments that probably represent fine sediments of the early Pleistocene Victorville basin. We will see Mojave River sands at higher elevations as we follow the road uphill to the southwest. The GWD capping the bluffs may correlate with those at Johnston Corners and farther east at Lenwood, suggesting a broad, marshy, east–west valley bottom during the late Pliocene and earliest Pleistocene.

38.3 (0.2) Join a road that proceeds south to the pole line road. PARK on hill crest where space is available.

38.6 (0.3) **Stop P-5—Helendale GWD** at 2600 ft on ridge top supported by GWD carbonates. This unit is similar to those above Stop P-4. About 10 km to the west-northwest, Dibblee mapped similar white beds as lake

deposits (Vitrinite deposits); if these GWD beds correlate with the Helendale, Johnston Corners, and west Barstow beds, a length of nearly 40 km of possibly contemporaneous GWD conditions are indicated. This white GWD horizon is arched along the northwest-trending Helendale fault, suggesting activity along that fault within the last few million years. Proceed southwest to National Trails Highway.

38.0 (0.4) STOP at National Trails Highway. WATCH FOR cross traffic. Enter National Trails Highway west-bound toward Helendale. Each car must watch for on-coming down hill traffic. ENTER Highway ONE CAR AT A TIME!

38.2 (0.2) Slow for curves.

39.5 (1.3) Continue past Vista Road to Silver Lakes.

43.5 (4.0) Continue past North Bryman Road.

45.0 (1.5) Continue past South Bryman Road.

47.9 (2.9) Slow; pass under the railroad bridge at the Riverside Cement Company Plant. Proceed slowly through Oro Grande.

50.8 (2.8) Cross the bridge over the Mojave River and continue past Rockview Nature Park.

51.4 (0.6) TURN RIGHT (west) at the signal for Air Expressway. Move to left lane.

52.0 (0.6) Continue past a left turn to Gas Line Road.

52.7 (0.7) TURN LEFT (south) on Village Drive and immediately pull right into turnout and PARK.

STOP P-6—Victorville

Stratigraphy. View northwest toward Southern California International Airport (formerly George Air Force Base). Stratigraphy described here consists of three units: lower, middle and upper (Reynolds and Cox, 1999; Cox and Tinsley, 1999; Sibbett, 1999; Cox and others, 1998, 2003).

The lower unit is a pebble-cobble arkosic sand deposited by a southward flowing axial stream system. Clasts of Jurassic and early Miocene volcanic rocks are derived from the Mojave Desert, not from the Transverse Ranges,

and include lithologies from the Kramer Hills and outcrops near Barstow. Borehole magnetostratigraphy suggests that the top of the south-flowing lower unit is about 2.5 Ma (Cox and others, 1999). This would be within the 4.1 to 1.5 Ma range of the Phelan Peak Formation on the north slope of the San Gabriel Mountains west of Cajon Pass. The Phelan Peak Formation records a reversal in its drainage direction from southward to northward in its depositional history (Meisling and Weldon, 1989), corresponding approximately to the age of the upper limits of the south-flowing lower unit in Victorville.

The middle unit of lacustrine sandy silt is seen in outcrop and on bore holes over a wide area. To the south, magnetostratigraphic data from bore holes indicate that the lacustrine middle unit is late Pliocene to early Pleistocene, ranging from 2.5 to ~2 Ma (Cox and others, 1999; 2003). At this stop, ostracode data suggests a Pleistocene age for the lacustrine middle unit.

The upper unit overlies the lacustrine middle unit, and is an upward coarsening fluvial sequence of silt, sand, and gravel. Clast lithologies indicate a source for these ancestral Mojave River gravels from the northwestern San Bernardino Mountains. A suggested depositional history is that stream flow from the southeast deposited the lacustrine middle unit in the Victorville basin north of the San Bernardino Mountains. As the ancestral Mojave became a cohesive drainage, and as the mountains continued to rise, sediments overwhelmed the basin, and the river then flowed northward and then eastward to terminate in Lake Manix (Reynolds and Cox, 1999).



Stop P-6. Mammoth tusk and rib, Lexington development near Village Drive and Amargosa Road. Dire wolf tooth associated with the mammoth tusk indicates a Rancholabrean (late Pleistocene) age for the greenish-gray clay overlain by coarse, orange arkosic sands (above white bucket) of the ancestral Mojave River. R. Hilburn photo.

Fossil dire wolf, mammoth, and horse were recovered in coarse Mojave River sands during monitoring of development near Village Drive and Amargosa Road (Reynolds, 2006). The presence of the dire wolf (*Canis dirus*) suggests a 0.2 Ma Rancholabrean North American Land Mammal Age (Bell and others, 2004) for the uppermost ancestral Mojave River sediments immediately below the Victorville Fan sequence. Fossil mammals from north of Air Expressway (Air Base Road) include giant ground sloth (*Paramylodon* sp.), short-faced bear (*Arctodus* sp.), mammoth, horse, camel, llama, giant camel (*Titanotylopus* sp.), and a meadow vole (*Microtus* sp.) (Scott and others, 1997). These mammals, particularly *Titanotylopus* and *Microtus*, suggest an early Pleistocene age, less than 1.3 Ma, for ancestral Mojave River sediments. The occurrence of the fossil cotton rat (*Signodon minor*) on Hesperia Road (Reynolds and Reynolds, 1994) suggests a Pliocene age for sediments along the Mojave River south of I-15.

52.8 (0.1) TURN RIGHT (east) on to Air Expressway and proceed to National Trails Highway.

54.0 (1.2) TURN RIGHT (south) on to Air Expressway and proceed toward Victorville and I-15.

55.1 (1.1) Continue past a thick sequence of sediments containing thin sections of the lower fluvial unit, the thin, lacustrine middle unit, and a thick section of the ancestral Mojave River gravels as the upper unit (Reynolds and Cox, 1999).

55.7 (0.6) Junction with I-15 in Victorville (services available) and the end of the trip. Options for routes home include following I-15 east toward Barstow to reach Arizona and Nevada. I-15 west (southbound) reaches San Bernardino and the Los Angeles basin. Bakersfield can be reached by backtracking to Air Expressway, proceeding west to Highway 395, then north to Highway 58.

End Day 3

References cited

- Bell, C. J., and E. J. Lundelius Jr., and seven others. 2004. The Blancan, Irvingtonian and Rancholabrean mammal ages. Mammalian biochronology of the Arikareean through Hemphillian interval, in Late Cretaceous and Cenozoic Mammals of North America: Biostratigraphy and Geochronology, M.O. Woodburne, ed. New York: Columbia University Press: 232-314.
- Byers, F. M., Jr. 1960. Geology of the Alvord Mountain quadrangle, San Bernardino County, California. U. S. Geological Survey Bulletin, 1089:-71P.
- Cole, J. M., E. T. Rasbury, G. N. Hanson, I. P. Montañez and V. A. Pedone. 2005. Using U-Pb ages of Miocene tufa for correlation in a terrestrial succession, Barstow Formation, California. Geological Society of America Bulletin 2005;117;276-287.
- Cox, B. F., J. W. Hillhouse, A. M. Sarna-Wojcicki, and J. C. Tinsley. 1998. Pliocene-Pleistocene depositional history along the Mojave River north of Cajon Pass, California—regional tilting and drainage reversal during uplift of the central Transverse Ranges: Geol. Soc. Amer. Abstracts with Programs, v. 30, no. 5, p.11.
- Cox, B. F., and J. C. Tinsley, III. 1999. Origin of the late Pliocene and Pleistocene Mojave River between Cajon Pass and Barstow, California. San Bernardino County Museum Association Quarterly 46(3):49-54.
- Cox, B.F., Hillhouse, J.W., and Owen, L.A. 2003. Pliocene and Pleistocene evolution of the Mojave River, and associated tectonic development of the Transverse Ranges and Mojave Desert, based on borehole stratigraphy studies and mapping of landforms and sediments near Victorville, California, in Paleoenvironments and paleohydrology of the Mojave and southern Great Basin Deserts, Y. Enzel, S.G. Wells, and N. Lancaster, eds. Geological Society of America Special Paper 368, p. 1-42.
- Cooper, J. F., Jr., G. E. Dunning, T. A. Hadley, W. P. Moller, and R. E. Reynolds. 2002. The Sulfur Hole, Calico District, San Bernardino County, California, California. Calif. State Univ. Fullerton, Desert Symposium 2002 Proceedings Volume. p. 29-36.
- Dibblee, T. W., Jr. 1968. Geology of the Opal Mountain and Fremont Peak quadrangles, California. California Division of Mines and Geology, Bulletin 188:64p.
- Dibblee, T. W., Jr. 1970. Geologic map of the Daggett quadrangle, California. San Bernardino County, California. U. S. Geological Survey Miscellaneous Geologic Investigations Map I-592, 1:62,500.
- Dudash, Stephanie L. & Geological Survey (U.S.) 2006 Preliminary surficial geologic map of a Calico Mountains piedmont and part of Coyote Lake, Mojave Desert, San Bernardino County, California [electronic resource] / by Stephanie L. Dudash ; U.S. Department of the Interior, U.S. Geological Survey.
- Forester, R.M., D.M. Miller, and V.A. Pedone. 2003. Ground water and ground-water discharge carbonate deposits in warm deserts, in Land of Lost Lakes, R.E. Reynolds, ed. Desert Studies Consortium, California State University, p. 27-36.
- Hildebrand, G. H. 1982. Borax Pioneer: Francis Marion Smith: San Diego, Howell-North Books, 318p.
- Hillhouse, J.W., R.E. Wells and B.F. Cox. 2010. Paleomagnetism of Miocene volcanic rocks in the Newberry Mountains, California: vertical-axis rotation and a polarity transition, in Overboard in the Mojave: 20 million years of lakes and wetlands, R.E. Reynolds and D.M. Miller, eds. California State University Fullerton, Desert Studies Consortium, this volume.
- Jachens, R. C., V. E. Langenheim, and J. C. Matti. 2002. Relationship of the 1999 Hector Mine and 1992 Landers fault ruptures to offsets on Neogene faults and distribution of Late Cenozoic basins in the Eastern California Shear Zone. Bull. Seismological Soc. Amer., 92(4):1592-1605.
- Leggitt, V. L. 2002. Preliminary report on the stratigraphic setting and microstructure of avian dggshell fragments from the Calico Mountains: Barstow Formation, Mojave Desert, California. in Between the basins: exploring the western Mojave and southern Basin and Range Province, R.E. Reynolds, ed. California State University, Fullerton, Desert Studies Symposium: 51-57.
- Leslie, S.R., D.M. Miller, J.L. Wooden, and J.A. Vazquez. 2010. Stratigraphy, age, and depositional setting of the Miocene Barstow Formation at Harvard Hill, central Mojave Desert, California, in Overboard in the Mojave: 20 million years of lakes and wetlands, R.E. Reynolds and D.M. Miller, eds. California State University Fullerton, Desert Studies Consortium, this volume.
- Meek, Norman. 1990. Late Quaternary geochronology and geomorphology of the Manix Basin, San Bernardino County, California. Ph. D. dissertation: Los Angeles, CA. Univ. of California, 212 p.

- Meek, Norman. 2000. The late Wisconsinan history of the Afton Canyon area. San Bernardino County Museum Association Quarterly 47(2):32-34.
- Meisling, K.E., and R. J. Weldon. 1989. Late Cenozoic tectonics of the northwestern San Bernardino Mountains, southern California. Geological Society of America Bulletin 101:106-128.
- Miller, D.M., K.M. Schmidt, S.A. Mahan, J.P. McGeehin, L.A. Owen, J.A. Barron, F. Lehmkuhl, and R. Löhner. 2010. Holocene landscape response to seasonality of storms in the Mojave Desert: Quaternary International, v. 215, p. 45-61.
- Miller, D.M., S.R. Leslie, J.W. Hillhouse, J.L. Wooden, J.A. Vazquez, and R.E. Reynolds. 2010. Reconnaissance geochronology of tuffs in the Miocene Barstow Formation: implications for basin evolution and tectonics in the central Mojave Desert, *in* Overboard in the Mojave: 20 million years of lakes and wetlands, R.E. Reynolds and D.M. Miller, eds. California State University Fullerton, Desert Studies Consortium, this volume.
- Myrick, D. F. 1992. Railroads of Nevada and Eastern California, Vol. II, the southern roads. University of Nevada Press, Reno, 933p.
- Reheis, M.C., and D.M. Miller. 2010. Environments of near shore lacustrine deposition in the Pleistocene Lake Manix basin, south-central California: in Overboard in the Mojave: 20 million years of lakes and wetlands, R.E. Reynolds and D.M. Miller, eds. California State University Fullerton, Desert Studies Consortium, this volume.
- Reheis, M.C., D.M. Miller, and J.L. Redwine. 2007. Quaternary stratigraphy, drainage-basin development, and geomorphology of the Lake Manix basin, Mojave Desert: Guidebook for Fall Field Trip, Friends of the Pleistocene, Pacific Cell, October 4-7, 2007: U.S. Geological Survey Open-File Report v. 2007-1281, 31 p.
- Reheis, M.C., and J. L. Redwine. 2008. Lake Manix shorelines and Afton Canyon terraces: Implications for incision of Afton Canyon, *in* Late Cenozoic drainage history of the southwestern Great Basin and lower Colorado River region, M.C. Reheis, R. Hershler, and D.M. Miller, eds. Geological and biotic perspectives: Geological Society of America Special Paper 439, p. 227-260.
- Reynolds, R. E. 1991. Hemingfordian/Barstovian Land Mammal Age faunas in the central Mojave Desert, exclusive of the Barstow Fossil Beds, *in* Inland Southern California: The last 70 million years, M.O. Woodburne, R.E. Reynolds, and D.P. Whistler, eds. Redlands: San Bernardino County Museum Association Quarterly 38(3,4):88-90.
- Reynolds, R. E. 1998. Flamingo egg from the Miocene sediments of the Calico Mountains, San Bernardino County, California. San Bernardino County Museum Association Quarterly, 45(1, 2):106.
- Reynolds, R.E. 1999a. Gomphothere tracks in southern California, *in* Fossil footprints, R.E. Reynolds, ed. San Bernardino County Museum Association Quarterly, 46(2): 31-32.
- Reynolds, R.E. 1999b. A walk through Borate, rediscovering a borax mining town in the Calico Mountains. San Bernardino County Museum Association Quarterly, 46(1): 31.
- Reynolds, R. E., and B.F. Cox. 1999. Tracks along the Mojave: a field guide from Cajon Pass to the Manix Basin and Coyote Lake. San Bernardino County Museum Association Quarterly 46(3):1-26.
- Reynolds, R.E., and M.O. Woodburne. 2001. Review of the Proboscidean datum within the Barstow Formation, Mojave Desert, California. Journal of Vertebrate Paleontology, Abstracts of Papers, 21(3): 93A.
- Reynolds, R.E., and M.O. Woodburne. 2002. Review of the Proboscidean datum within the Barstow Formation, Mojave Desert, California [abs], *in* Between the basins: exploring the western Mojave and southern Basin and Range Province, R.E. Reynolds, ed. California State University, Desert Studies Consortium: 82-83.
- Reynolds, R.E. 2006. Paleontological resources monitoring program, KB Homes Saratoga Project (Tract 13908), Victorville, California. LSA Associates, Inc. for KB Homes, 16 p.
- Reynolds, R.E. and R.L. Reynolds, 1994. The Victorville Fan and the Occurrence of Sigmodon. San Bernardino County Museum Association Special Publication, pp. 31–33.
- Sibbett, B. S. 1999. Pleistocene channels of the Mojave River near Victorville, California. San Bernardino County Museum Association Quarterly 46(3):p. 65-68.
- Scott, E., K. Springer, and L. K. Murray. 1997. New records of Early Pleistocene vertebrates from the west-central Mojave Desert, San Bernardino County, CA. Jour. Vert. Paleo. Vol. 17(3), p. 75A.
- Schneider, J. S. 1989. Fresh Water Bivalves as Paleoenvironmental Indicators. Abstracts of Proceedings, 1989 Mojave Desert Quaternary Research Symposium, San Bernardino County Museum Association Quarterly, V. XXXVI, No. 2, p. 65
- Schuiling, W. T., 1999. A Miocene hot spring exhalite in the southern Calico Mountains. San Bernardino County Museum Association Quarterly 46(3):89-94.
- Singleton, J. S., and P. B. Gans. 2005. Geologic evolution of the southeastern Calico Mountains, central Mojave Desert, California: Geologic Society of America, Cordilleran Section, v. 37, p. 73.
- Singleton, J.S., and P.B. Gans. 2008. Structural and stratigraphic evolution of the Calico Mountains: Implications for early Miocene extensions and Neogene transpression in the central Mojave Desert, California: Geosphere, v. 4, p. 459-479.
- Tedford, R. H., L. B. Albright III, A. D. Barnosky, I. Ferrusquia-Villafranca, R. M. Hunt Jr., J. E. Storer, C. C. Swisher III, M. R. Voorhies, S. D. Webb and D. P. Whistler. 2004. Mammalian Biochronology of the Arikarean through Hemphillian Interval, *in* Late Cretaceous and Cenozoic Mammals of North America: Biostratigraphy and Geochronology, M.O. Woodburne, ed. New York: Columbia University Press: 169-231.
- Warren, C. N. and J. S. Schneider. 2000. Archaeology in Cronese Basin: a history. San Bernardino County Museum Association Quarterly 47(2):40-41.
- Weber, H. F., Jr. 1966. Silver Mining in Old Calico. Mineral Information Service, Calif. Div. Mines & Geol., Vol. 19 (5), p. 71-80.
- Weber, H. F., Jr. 1967. Silver Deposits of the Calico District. Mineral Information Service, Calif. Div. Mines & Geol., Vol. 20 (1), p. 3-8.
- Wells, R. E., and J. W. Hillhouse. 1989. Paleomagnetism and tectonic rotation of the lower Miocene Peach Springs Tuff: Colorado Plateau, Arizona, to Barstow, California, Geol. Soc. Am. Bull., 101, 846-863.
- Woodburne, M.O. 1991. The Mojave Desert Province, *in* Inland Southern California: the last 70 million years, M.O. Woodburne, R.E. Reynolds, and D.P. Whistler, eds. Redlands: San Bernardino County Museum Association Quarterly 38(3,4):60-77.
- Woodburne, M.O. and R. E. Reynolds. 2010. The mammalian litho- and biochronology of the Mojave Desert Province, *in* Overboard in the Mojave: 20 million years of lakes and wetlands, R.E. Reynolds and D.M. Miller, eds. California State University Fullerton, Desert Studies Consortium, this volume.

Environments of nearshore lacustrine deposition in the Pleistocene Lake Manix basin, south-central California

Marith C. Reheis and David M. Miller

USGS, Menlo Park, CA mreheis@usgs.gov, dmiller@usgs.gov

Abstract

Lake Manix, in the central Mojave Desert of southern California, was the terminus of the Mojave River from about 500 to 25 ka. Deposits of this lake provide excellent examples of depositional environments of a low-desert, clastic-dominated lake sustained by a mostly perennial river, a setting atypical for most Great Basin pluvial lakes. We characterize and provide examples of the nearshore environments of deposition of Lake Manix, which include extensive low-gradient fluvial-deltaic and mudflat deposits and ephemeral alluvial-fan-lake-marginal deposits, to aid in their recognition in other settings and in the geologic record. The sedimentary character and stratigraphic architecture of these nearshore deposits depend on proximity to the mouth of the Mojave River, to position on fringing alluvial fans, to steepness of fan slope, and to presence of active channels. Recognition of these nearshore deposits, some of which are enigmatic or difficult to discern in the geologic record, may provide valuable information on response of desert lakes to past climate change.

Introduction

Deposits of paleolakes have been studied for well over a century (e.g., Russell, 1885; Gilbert, 1890; Morrison, 1991; Adams, 2007). Characterization of lake sediments and stratigraphy provides information on age, hydrologic environments (water temperature and chemistry), water depth, relations to feeder streams, and basin evolution. Sedimentary and chemical facies of lakes also provide analogues for interpretation of ancient sediments in the rock record (e.g., Smith et al., 1983; Lowenstein et al., 1999; Adams, 2007; Waldmann et al., 2009). The large pluvial lakes in the Great Basin, such as Lakes Bonneville, Lahontan, Russell, and Owens, have been magnets for geological investigations of both outcrops and cores. These lakes lie in the higher, less arid part of the Great Basin, received direct runoff from extensive mountain glaciers, and were fed by sizable rivers. Nearshore deposits of these lakes, such as beach barrier and deltaic complexes along steep range fronts, have been thoroughly studied (Russell, 1885; Gilbert, 1890; Komar, 1998; Adams and Wesnousky, 1998).

In contrast, only a few sedimentologic studies have targeted lake basins in the southern low deserts of the Great Basin, and none have focused on nearshore deposits other than beach complexes. Southern lake records include Lake Estancia, in central New Mexico (outcrop studies

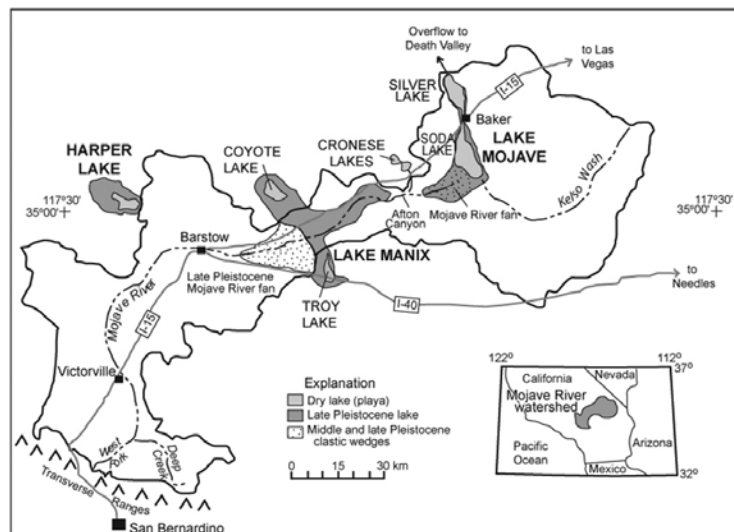


Figure 1. Location map of the Mojave River watershed and paleolakes sustained by the river. Modified from Wells et al. (2003).

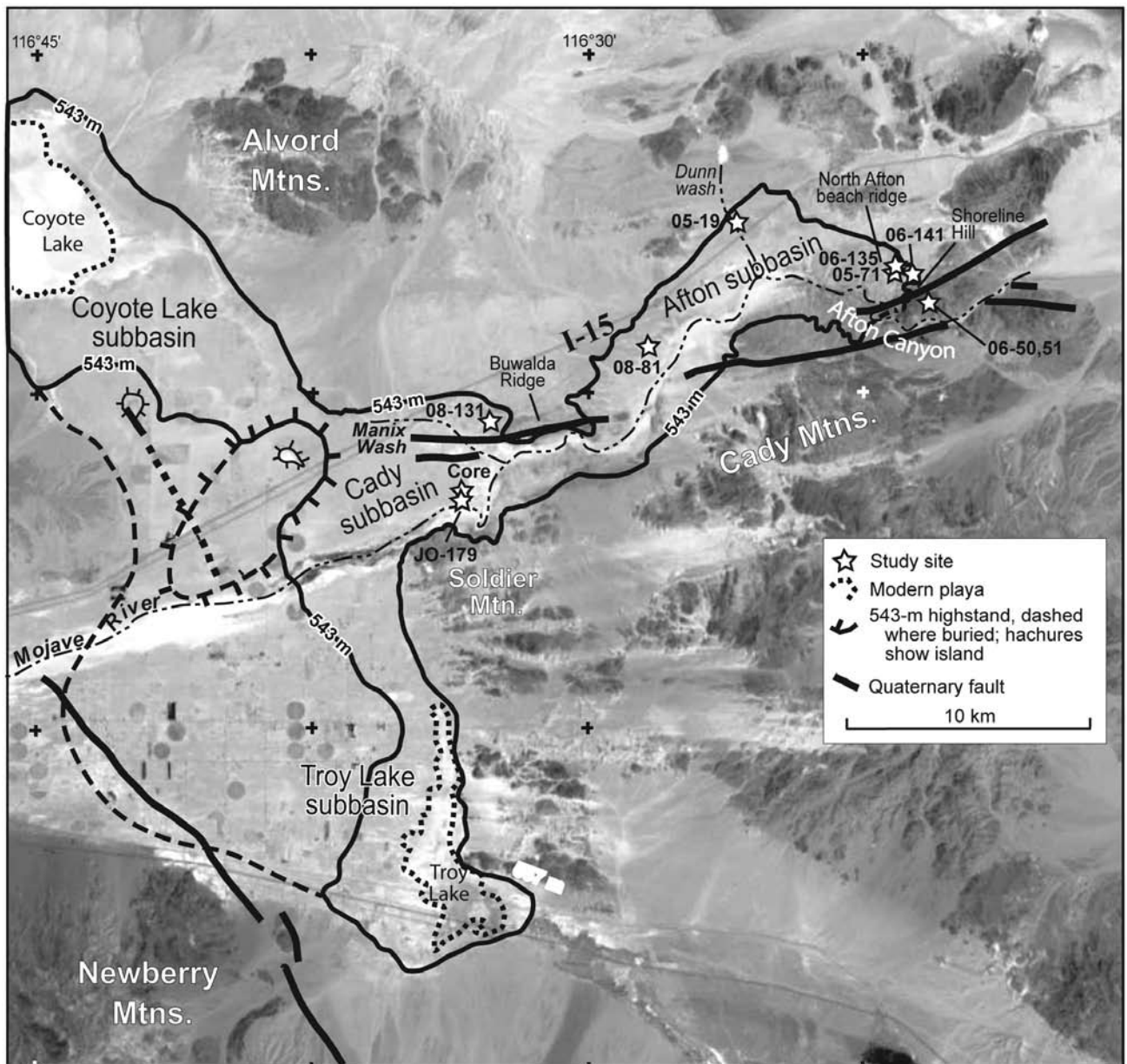


Figure 2. Geographic features of the Manix basin, showing subbasins, playas (dotted), the Mojave River fluvial fan, and major faults (modified from Reheis and Redwine, 2008). Medium-weight line is the 543-m highstand level that depicts the minimum extent of Lake Manix during the late Pleistocene; western margin is not known due to progradation during highstands and later burial by Mojave River alluvium. Stars and numbers (e.g., 06-135; prefix "M" omitted for convenience) show locations of sites described in text and figures 3-9.

summarized in Allen and Anderson, 2000) and Lake Babicora of Chihuahua state in Mexico (coring studies by Metcalfe et al., 2002). Neither lake was sustained by substantial rivers, and Babicora lay at relatively high altitude (~2140 m). In the low Mojave Desert, well-characterized pluvial lakes include Lake Mojave (Enzel et al., 1992; Wells et al., 2003), the present terminus of the Mojave River (Fig. 1), and Lakes Searles, Panamint, and Manly (Smith, 1976; Smith et al., 1983; Forester et al., 2005; Jayko et al., 2008; Knott et al., 2008; Phillips, 2008). These lakes lay along the episodic overflow route of the Owens River and, for Lake Manly, also the Amargosa

River (Menges, 2008). Most sedimentologic interpretations of the latter three lakes focus on the features in their deep-water settings and especially chemical precipitates as seen in cores.

Lake Manix, the terminal basin of the Mojave River from about 500,000 to 25,000 years ago (Jefferson, 2003) (Figs. 1 and 2), was sustained by the Mojave River, which has its source in the Transverse Ranges. During historic time, and likely during much of the Holocene, the Mojave River was ephemeral in its lower reaches below Victorville (Enzel and Wells, 1997). Short-duration Holocene lakes that formed in the Silver Lake basin are interpreted

to have been sustained by extreme winter storms in the headwaters that persisted over years to decades (Enzel et al., 1992; Enzel and Wells, 1997). Enzel et al. (2003) inferred that sustaining a lake during the late Pleistocene, when evaporation rates were reduced and transmission losses to groundwater may have been lower, would have required an average annual discharge at least an order of magnitude larger than today. Although Lake Manix was farther upstream in the fluvial system than the younger Lake Mojave studied by Enzel et al., and likely had lower transmission losses as a result, river inputs that sustained Lake Manix were almost certainly highly variable. Preliminary results from outcrop study and from analysis of sedimentology and stable isotopes in the U.S. Geological Survey (USGS) Manix core (Reheis et al., 2009a, b) suggest large variability in lake level on timescales ranging from possibly seasonal to millennial. Variable river input, combined with relatively high evaporation rates in this low-desert setting (Meyers, 1962), would have caused lake level to fluctuate continually and would have created dynamic and generally unstable nearshore environments.

The general history of Lake Manix was established by Jefferson and Meek (Ellsworth, 1932; Meek, 1990, 2000; Jefferson, 2003; Meek, 2004), and our recent mapping, stratigraphic studies, and dating (Miller and McGeehin, 2007; Reheis et al., 2007b; Reheis and Redwine, 2008; Reheis et al., 2009a, 2009b) have added much new information. The Manix basin contains a spectacularly well-exposed record of lake fluctuations in settings that are somewhat atypical for Great Basin pluvial lakes, in that the sediments include extensive low-gradient fluvial-deltaic and mudflat deposits, as well as ephemeral alluvial-fan-lake-marginal deposits. The exposures are unusual because the lake drained by threshold failure, with stream flow eventually cutting into the floor of the lake, incising lake and alluvial deposits in a wide variety of geomorphic settings. The purpose of this short paper is to characterize and provide examples of the nearshore environments of deposition of Lake Manix, to aid in their recognition in other settings and in the geologic record. We make no attempt to provide detailed sedimentologic or stratigraphic characteristics or to adopt formal basin-analysis procedures, as these are abundantly well documented in textbooks (e.g., Reading and Collinson, 1996; Talbot and Allen, 1996; Komar, 1998) and research papers. Rather, our intent is to point out some unique depositional features of this low-desert, fluvially controlled paleolake.

Nearshore environments of Lake Manix

Throughout the geologic history of Lake Manix, the lake interacted with a variety of depositional settings influ-

enced by proximity to the Mojave River, the amount of alluvial activity, surface slope, and basin configuration. There was an abrupt contrast in sedimentary environments near the migrating mouth of the Mojave River and the rest of the basin's margins, which are alluvial fans, in places mantled by or interbedded with eolian deposits. The Manix basin consists of four interlinked subbasins (Fig. 2). Although all of these subbasins are traversed by faults, some of them active as recently as the Holocene, their margins are not fault-bounded; hence, most of the nearshore environments are characterized by fluvial, alluvial-fan, and locally, eolian sediments. Three exceptions of steep topography adjacent to the lake are (1) the south face of Buwalda Ridge, which is underlain by a Pliocene fanglomerate that is bounded by the Manix fault, (2) the north face of Soldier Mountain, and (3) the eastern end of the Afton subbasin, where the lake abutted the bedrock of Shoreline Hill and slopes to the south. The Cady and Troy Lake subbasins lay nearest the entry point of the Mojave River, and thus their sediments record interaction of the lake with the encroaching fluvial fan of the Mojave River, as well as with alluvial fans of local drainages. The Coyote Lake subbasin, surrounded by alluvial fans, is separated by a shallow bedrock sill from the Cady subbasin (Meek, 1990, 1994), and hence served as a release valve for high lake levels in the rest of the Manix basin. As such, Coyote Lake was shallow or even dry during part of the history of Lake Manix. The Afton subbasin is most distant from the Mojave River inlet, and is bounded on the north and south by alluvial fans. On the east end, this subbasin is characterized by steep margins with relatively high-gradient fans and bedrock. The Afton subbasin was abruptly integrated into Lake Manix shortly before the deposition of the Manix tephra (Reheis et al., 2007b; Reheis et al., 2009a) at about 185 ka (Jefferson, 2003).

The nearshore depositional environments of Lake Manix are subdivided on the basis of adjacent deposits and their gradients as follows: (1) deltaic, interacting with fluvial inputs of the Mojave River, either proximal or distal to the river mouth; (2) interacting with active alluvial fan channels of either gentle or moderate to steep gradients; (3) interacting with alluvial-fan deposits with no active channels, with gentle or moderate gradients; and (4) interacting with colluvium and very steep fans along mountain fronts. These interactions resulted in characteristic suites of sediments and stratigraphic architectures, many of which of course overlap among these groups.

One reliable indicator of nearshore environments in Lake Manix is the presence of stromatolitic carbonate-coated clasts, termed "oncoids" by Awramik et al. (2000). These coats consist of relatively dense, commonly laminar, single to multiple, wavy layers ranging from <1 mm

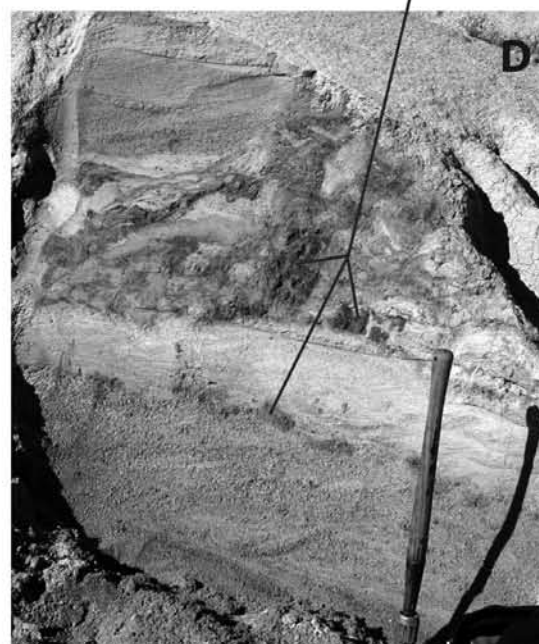
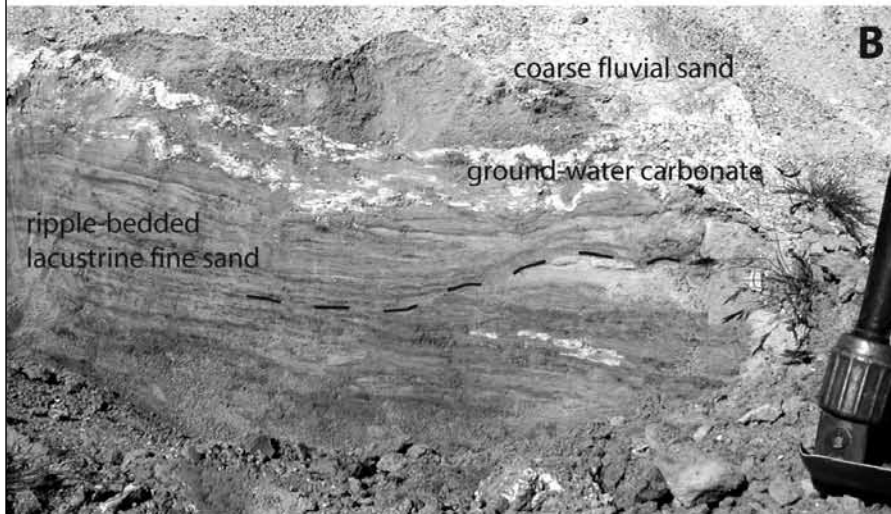
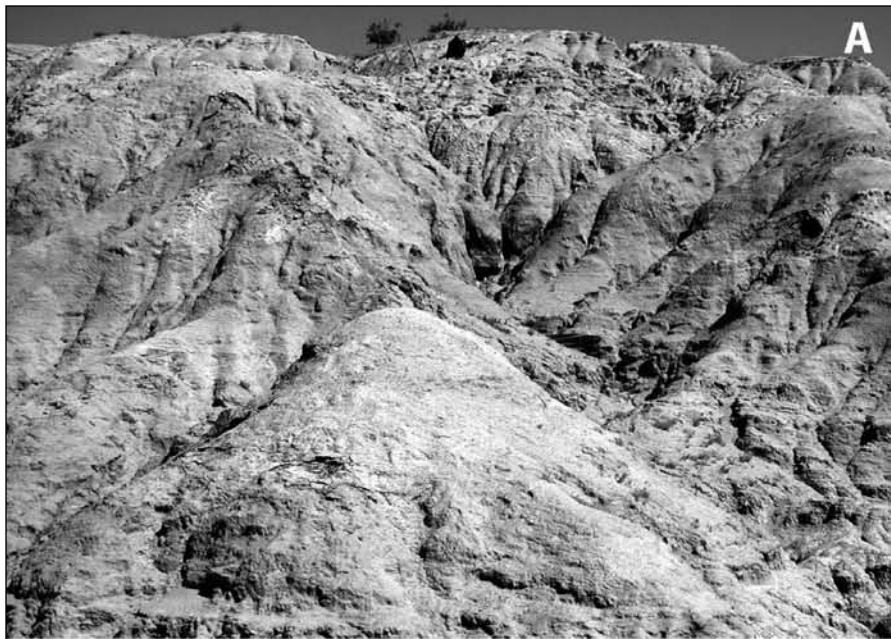


Figure 3. Upper ~8 m of section ("member D" of Jefferson, 2003) exposed along bluffs north of river at intersection of Mojave River and Manix Wash. A, weathered outcrop appearance of parallel-bedded sand and intercalated finer-grained sediment consisting of stacked fluvial and minor lacustrine deposits. Most beds of fluvial sand are oxidized (darker bands). Pack for scale at top. B, oscillatory ripple-bedded laminated fine-grained sand (lacustrine) overlain by wavy white bands of ground-water carbonate, in turn overlain by poorly sorted coarse-grained sand (fluvial). Dashed line indicates possible erosional disconformity truncating white sand in center right. C, cross-bedded sands with thin mud drapes. In lower part, beds fine upward; in upper part, beds coarsen upward from a basal mud. D, basal coarse-grained, planar-bedded sand is overlain by mud rip-ups, then by pale cross-bedded well-sorted sand, in turn capped by more mud rip-ups. Photographs by D. Miller.

to several cm in thickness. They are nearly always found at or just above the contact of alluvial fan gravels with overlying lacustrine deposits and are associated with lake transgression (Awramik et al., 2000). Our observations indicate that oncoid coats tend to be thin (or absent) in distal fan positions, where single layers of coated clasts may be found, and on active fan-lake margins, where multiple layers of clasts with thin coats may be deposited. Distal fans in the Manix basin are typified by oxidized muddy sand and fine gravel with little soil development. Fine gravel is easily wave-transported and this, combined with rapid lake transgression over low-gradient distal fans, may inhibit oncoid formation. Thick oncoid coats are most common along stable medial to proximal fan-lake margins, where they accreted on pebbles to boulders composing, or reworked from, a desert pavement overlying moderately to well-developed soils formed on fan

gravel. These relations, together with the absence of tufa towers or mounds constructed by groundwater—lake interaction as is common today at Mono Lake and Pyramid Lake, suggest that the calcium required to construct the oncoids is derived from near-surface pedogenic carbonate of the subjacent alluvial fan during lake transgression.

1. Fluvial-deltaic deposits

Fluvial-deltaic deposits at the junction of the perennial Mojave River and Lake Manix are widely exposed as the uppermost 8-10 m of outcrop along the northern bluffs of the modern Mojave River near and west of the USGS Manix core site (Fig. 2). They were described as “Member D” of the Manix Formation by Jefferson (2003) and accumulated between about 50 and 25 ka (Miller and McGeehin, 2007; Reheis et al., 2007b). These deposits form nearly parallel-bedded exposures on faces parallel to the

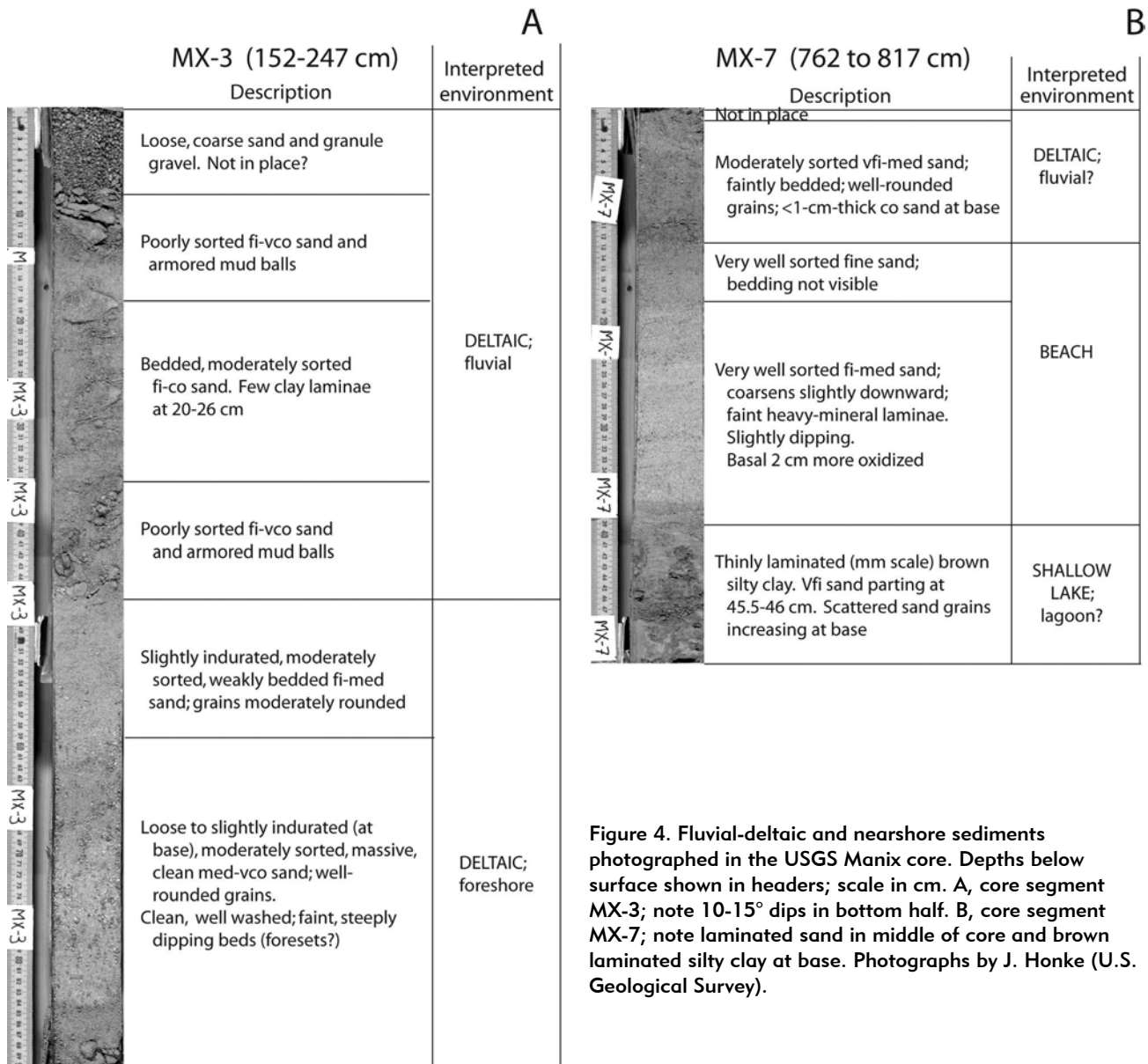


Figure 4. Fluvial-deltaic and nearshore sediments photographed in the USGS Manix core. Depths below surface shown in headers; scale in cm. A, core segment MX-3; note 10-15° dips in bottom half. B, core segment MX-7; note laminated sand in middle of core and brown laminated silty clay at base. Photographs by J. Honke (U.S. Geological Survey).

modern Mojave River (Fig. 3A), that is, oriented parallel to past streamflow, and form notably lenticular-bedded deposits in faces perpendicular to stream flow. Lenses fill channels cut into adjacent deposits and into the top of older Lake Manix deposits, and generally channels are draped by mud and sand interpreted as overbank deposits. The overbank deposits fine upward from pebble gravel to mud beds with rip-up mud balls common near the base and mudcracks common at the top. The sand beds in places contain calcium-carbonate nodules and plates (Fig. 3B) that we interpret as forming from groundwater discharge; locally, shell beds were dissolved and re-precipitated as soft nodules (G. Jefferson, written commun., 2010). Most beds are composed of grussy sand and fine-grained gravel, with less common mud beds and laminae (Figs. 3C and D); the clast composition indicates deposition by the Mojave River. Sedimentary facies and structures are largely fluvial and include “cross-bedded channel-fill deposits, crevasse-splay deposits, marsh, and floodplain deposits” (D. Miller, p. 23-24 in Reheis et al., 2007b). The rare beds that least ambiguously indicate lacustrine deposition are planar-bedded, very thin, and composed of well-sorted fine-grained sand (Fig. 3B), some of which are diatomaceous. Intervals of poorly sorted sand and silt, including armored mud balls, are locally observed in the USGS Manix core (Fig. 4A). Poorly sorted sands and those with curving crossbeds are interpreted as fluvial, whereas well-sorted, oscillatory ripple-bedded sands are interpreted as lacustrine (Fig. 3B). Thin intervals (10-20 cm) of very well sorted, sometimes gently dipping, sand and silt are also interpreted as lacustrine (beach) deposits (Fig. 4B), and thin intervals of steeply dipping deltaic foreset beds are present (Reheis et al., 2007). Moderately sorted, planar-bedded sands (basal sand, Fig. 3D) may represent either lacustrine or fluvial deposition.

The fluvial-deltaic sediments contain fossils of freshwater invertebrates and some vertebrates (Jefferson, 1987, 2003). Beds with abundant *Anodonta* shells found in these deposits have provided radiocarbon ages. Some beds also contain freshwater snails and turtle remains (*Clemmys marmorata*; Jefferson, 2003). However, modern representatives of these fossils live in a variety of freshwater habitats including streams, rivers, ponds, lakes, and bogs. Jefferson (1987) interpreted the mixture of fossils from different habitats in two beds near the bottom and top of the fluvial-deltaic section to represent deposition by floods. *Anodonta* occupy both fluvial and lacustrine habitats, and were historically present in the Mojave River (reported in Meek, 1990); thus, their presence is not diagnostic of depositional setting. The *Anodonta* beds we sampled for dating were composed of clusters of whole and half shells with no other species present, and likely grew in-situ or

were little reworked. Fish fossils (*Gila bicolor mojavensis*; Jefferson, 2003) and lacustrine ostracodes (Steinmetz, 1987; Bright et al., 2006), have been identified in older, finer grained sediments of Lake Manix and also in younger lacustrine sands in the Afton subbasin (K. Gobalet, CSU Bakersfield, written commun., 2005), but have not been reported in the fluvial-deltaic deposits. Thin white beds of diatomaceous sand occur in the central part of “Member D” in some sections north of the core site. Diatoms identified by Scott Starratt (USGS, written commun., 2008), are mostly benthic forms that populate shallow (<10-20 m depth), fresh to slightly brackish waters that were mesotrophic to eutrophic.

The limited presence of clearly lacustrine sediment and classic Gilbert-type delta foreset and bottomset beds and the dominance of fluvial sedimentary features suggest that these deposits represent mainly fluvial aggradation, with temporary transitions to a fluvial-dominated delta on a gently sloping lake margin with shallow lake depths. In addition, the frequent 5-15 m fluctuations in lake level during the period of deposition of these deltaic sediments (Reheis et al., 2009b) would have resulted in significant fluvial reworking, during lower lake levels, of any lacustrine sediment deposited at higher lake levels. The dominant westerly wind regime in this region also would have limited reworking by wave action at this western edge of the lake with minimal fetch. Similar deposits in the rock record, with limited occurrence of clearly nearshore or deltaic features, would likely not be recognized as being proximal to a lake.

Distal fluvial-deltaic and mudflat deposits are well exposed in the upper part of Jefferson’s (2003) “Member C” along the bluffs of the Mojave River and Manix Wash (Fig. 2). Although Jefferson interpreted most of these sediments to represent relatively high lake levels, detailed study of equivalent intervals in the USGS Manix core and in outcrop suggests instead that they represent shallow lacustrine to mudflat environments, with short intervals of deeper water, and reflect proximity to sediment inputs from the Mojave River (Oviatt et al., 2007; Reheis et al., 2007a). The primary reason for the change from relatively deep-water sedimentation recorded in older deposits to shallow water and mudflats in “upper C” is the shift of the lake depocenter to the newly incorporated Afton subbasin just prior to 185 ka (Reheis et al., 2007a; Reheis et al., 2009a). Because much of the accommodation space in the older part of the lake basin had been filled with Mojave river sediment, the area of the lake bottom near the present Mojave River-Manix Wash confluence was frequently exposed as a mudflat or playa after the depocenter shift, and was only submerged to moderate depths when Lake Manix approached highstand levels. In addi-

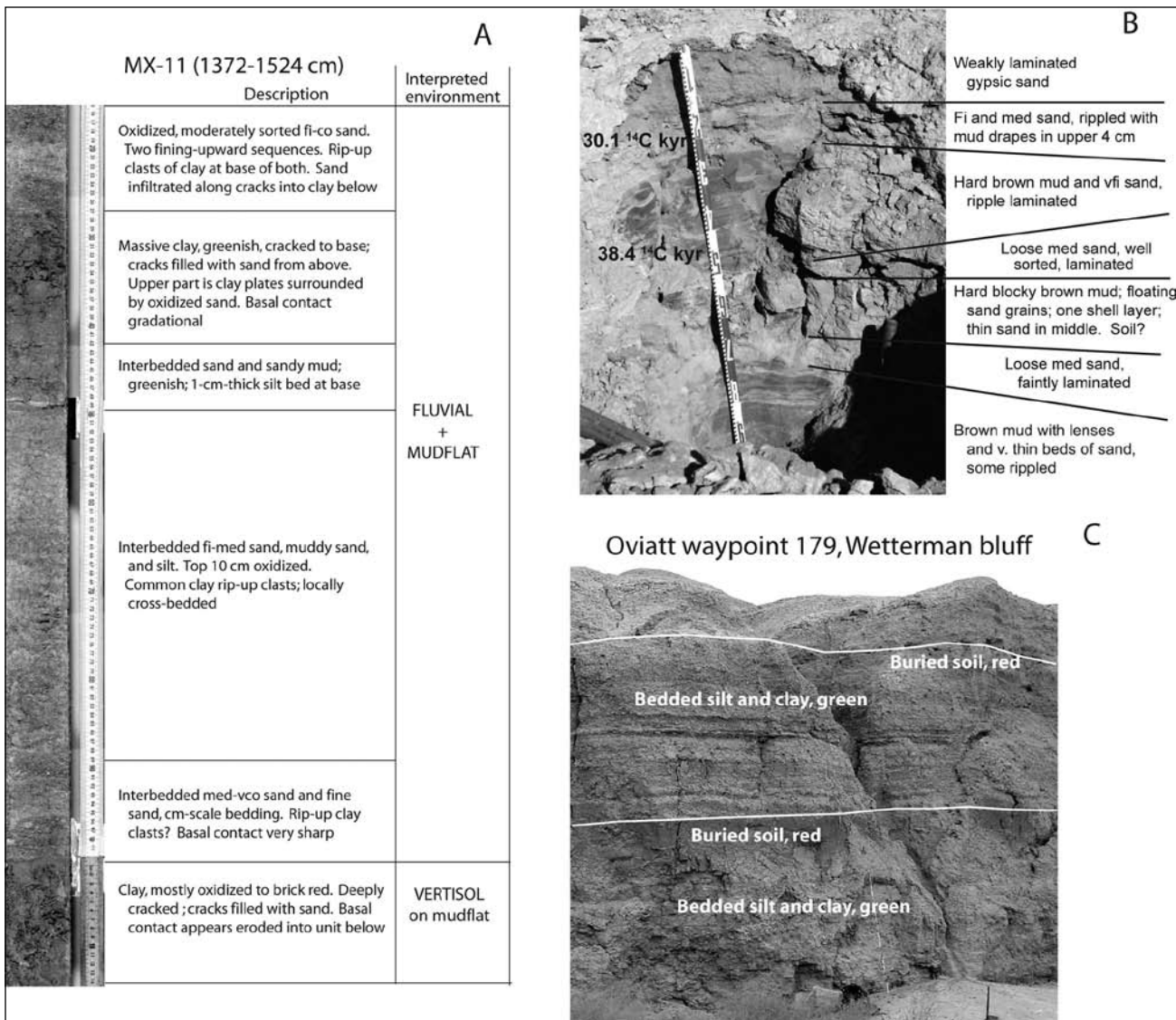


Figure 5. Distal fluvial-deltaic and mudflat deposits. A, segment MX-11 of the USGS Manix core shows fluvial fining-upward sequences with clay rip-ups and mudcracks. Photograph by J. Honke, U.S. Geological Survey. B, oscillating shallow-water and mudflat deposits exposed on the southeastern margin of the Coyote Lake subbasin. Numbers mark 10-cm increments on scale. Photograph by D. Miller. C, vertisols formed on desiccated lake sediments

tion, proximity to the encroaching front of the Mojave fluvial fan increased the frequency of incursion of fluvial sediment during low to moderate lake levels.

The distal deltaic and mudflat deposits typically consist of normally graded sequences of sand, silt, and clay (Figure 5A). The basal arkosic, coarse- to medium-grained sands are commonly oxidized and locally cross-bedded. In some intervals they grade up through thinly bedded and sorted, pale gray, very fine-grained sand and silt to laminated clay and silt, suggesting delta-front sedimentation. In other intervals, interpreted as fluvial sedimentation on a mudflat surface possibly from a single flooding event, the sands grade up through brown muddy sand and sandy mud to blocky clay with burrows and sand-filled cracks. Some fine-grained beds appear con-

torted and brecciated, and some sand beds are mixed with rip-up mud clasts. Muds that are cracked and oxidized (7.5 to 5 YR colors) are interpreted to represent episodes of soil formation on an exposed mudflat. These sequences differ from sandflats and mudflats previously described for the margins of arid closed basins (e.g., Talbot and Allen, 1996) because of the influence of the Mojave River.

Other mudflat deposits in the Manix basin, such as in the Coyote subbasin (Fig. 5C), share some characteristics with the distal deltaic deposits in that they are characterized by rapidly changing, interbedded, fining-upward sand, silt, and clay beds, with the clays cracked and oxidized. However, they lack evidence for fluvial sediment transport, and instead are similar to the sandflat and mudflat environments of other closed-basin lakes (Allen

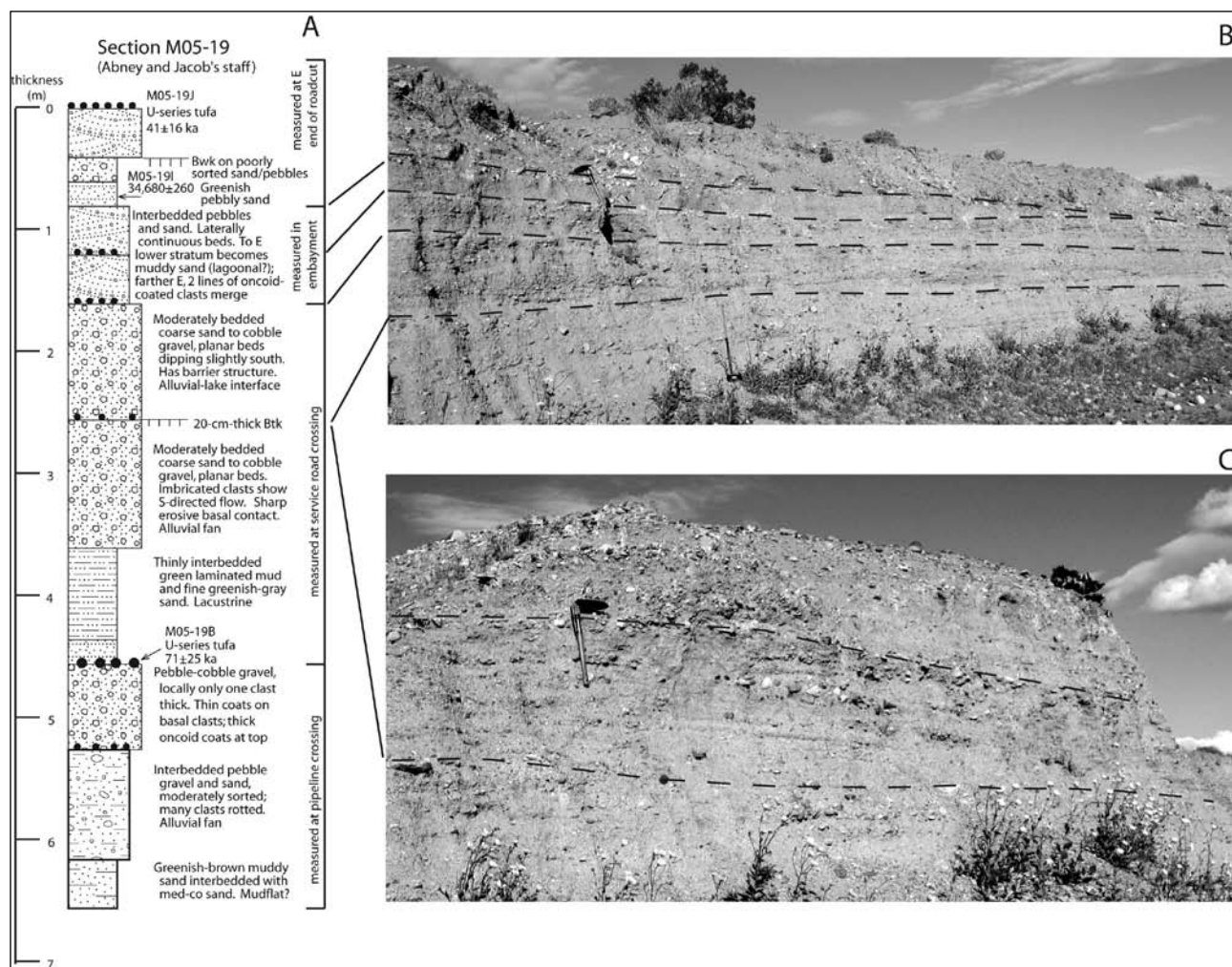


Figure 6. Deposits of gently sloping active fan-lake margins. A, measured section M05-19 in upper Dunn wash at pipeline crossing (modified from Reheis and Redwine, 2008). Filled black circles are lines of oncoid carbonate-coated clasts; size of circle proportional to oncoid thickness. Ticks on horizontal lines represent buried soils. M05-191 is a radiocarbon age obtained from *Anodonta* shell. B and C, photographs keyed to measured section; note barrier-bar structure with backset and frontset beds in second unit from base. Dashed lines are unit boundaries commonly marked by oncoid-coated clasts; only the lowest unit is alluvial. 50-cm trenching shovel for scale in both photographs (by M. Reheis).

and Anderson, 2000; Wells et al., 2003). In addition, intervals of bedded to massive silt and clay in the older, deeper-water sediments of Lake Manix (unit “lower C” of Jefferson, 2003) are commonly capped by prominent red soils (SYR colors; Fig. 5B), similar to those in the mudflat deposits. These soils resemble vertisols and represent subaerial weathering on desiccated lake sediments.

2. Active fan-lake interface deposits

Active alluvial fans are characterized by ephemeral drainages that episodically transport water and sediment. Due to the dynamic nature of the fan channels, which may shift laterally across the fan, the episodic nature of runoff, and rapid lake-level fluctuations, the fan-lake margin is a very dynamic environment. The sedimentary structures and stratigraphic architecture of these environments around the Manix basin depend primarily on slope. Previ-

ous studies have not well characterized such deposits formed in low-gradient settings.

Gently sloping active fan-lake margins are characterized by sediments that strongly resemble clast-supported alluvial-fan sediments. However, close examination reveals that in some beds, the sediments have been modified by lacustrine processes and (or) contain lacustrine fossils such as *Anodonta* shells and lacustrine ostracodes. Such sequences are well exposed in the upper reaches of Dunn wash, in northeastern Afton subbasin (Figs. 2 and 6; Reheis and Redwine, 2008). Here, the upper several meters consist of gravel and sand atop an erosional unconformity channeled into older, fine-grained lacustrine sediment (Fig. 6A). The gravelly deposits exhibit rapid lateral and vertical changes in bedding and sorting characteristics, but individual beds are laterally continuous and can be traced 25 m or more along outcrop unlike typical

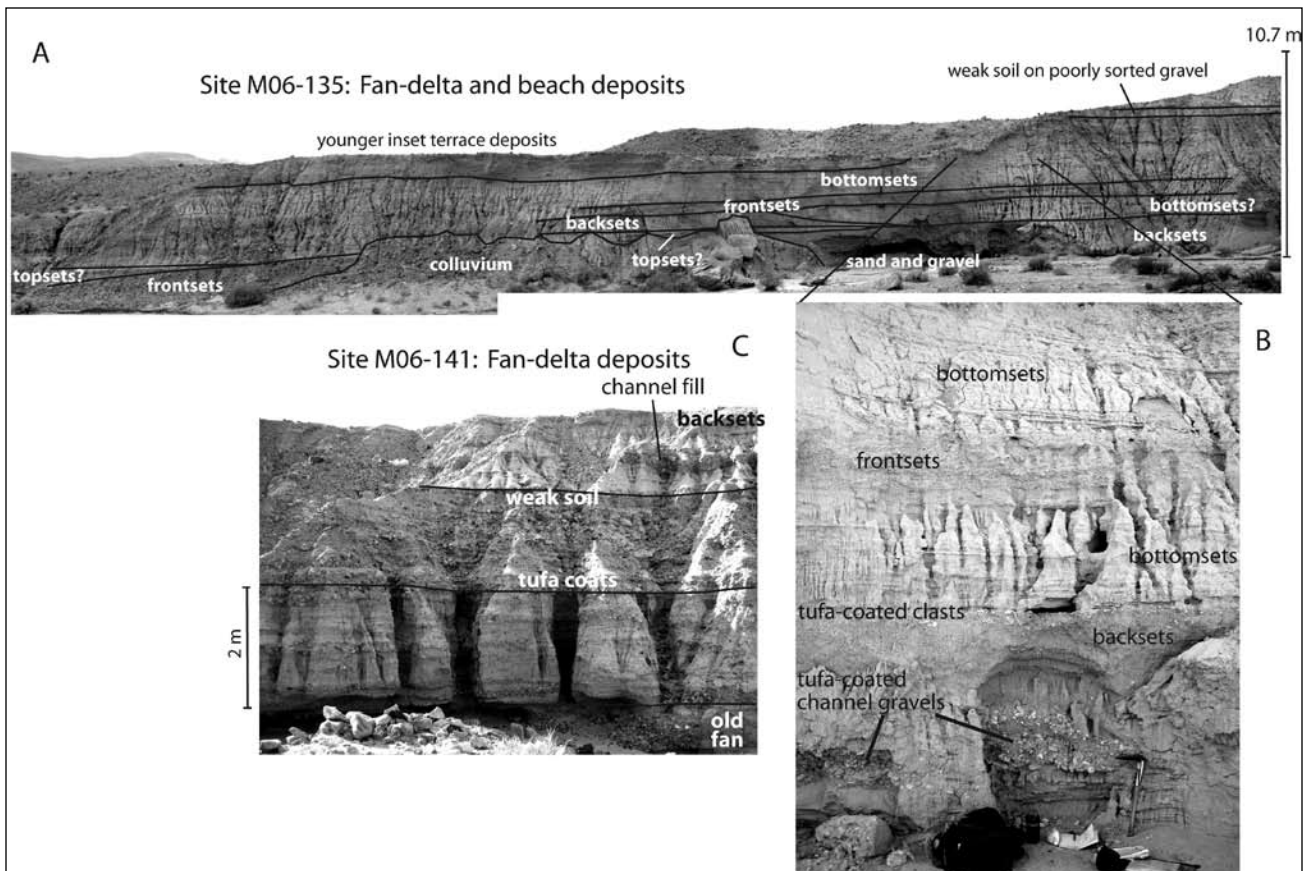


Figure 7. Deposits of steeply sloping active fan-lake margins. A, fan-delta and offshore-bar features exposed at site M06-135, east of the North Afton beach ridge. This section, oriented nearly N-S, lies at an oblique angle to SW-trending depositional slope, thus apparent bedding dip is low; where parallel to depositional slope, basal fan deposits slope $\sim 7^\circ$ basinward. Section represents alternating sequences of barrier structures with frontset and backset beds and deltaic structures with foreset and bottomset beds. Modified from Reheis et al. (2007b). B, closeup of bedding in part of upper photograph. 50-cm trenching shovel for scale. C, fan-delta deposits exposed at site M06-141 just north of Shoreline Hill; note rightward dip of gravel beds in middle unit. Photographs by M. Reheis.

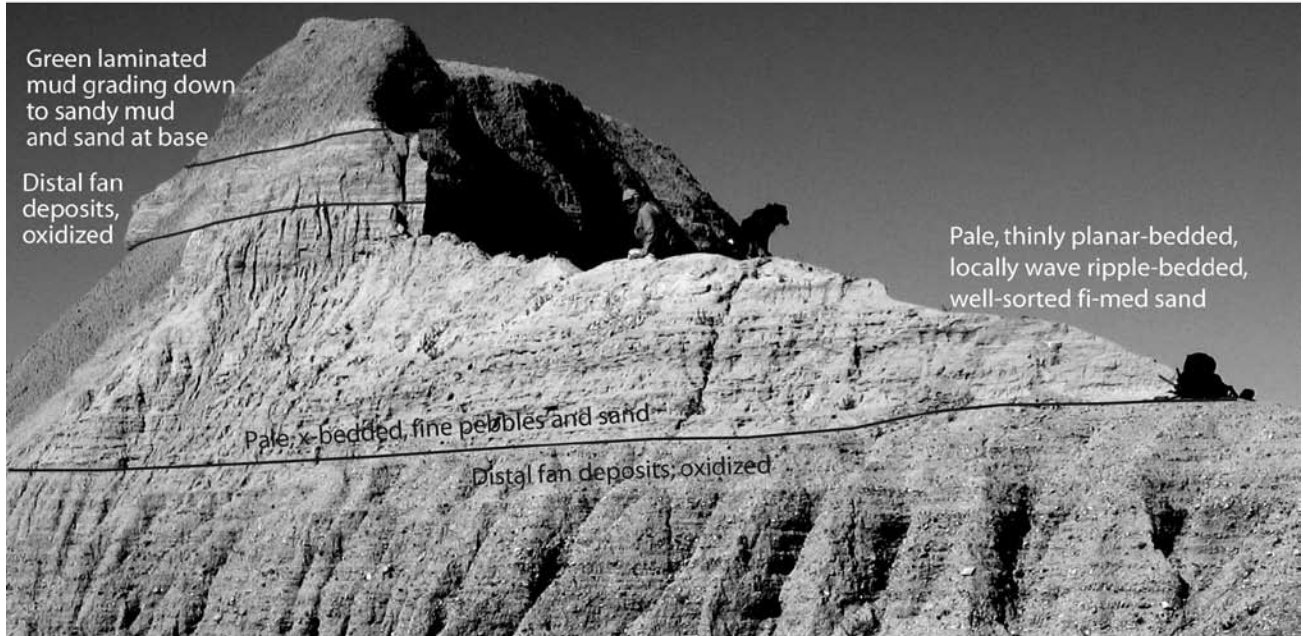
lenticular-bedded fan deposits. The dominantly fluvial units consist of planar-bedded coarse-grained sand to cobble gravel and in places, exhibit alluvial clast imbrication indicating downslope flow (Fig. 6C). Due to shifting fan channels and lake levels, fluvial channel fills may be inset within dominantly lacustrine sequences. Fluvial units that have been modified by nearshore processes have similar grain-size distribution, but are better sorted and somewhat better bedded (Fig. 6B). As a result, they are commonly looser and less indurated than the purely alluvial units because they have fewer fines mixed with the coarser sediment. Lake-reworked fluvial deposits are typified by thin lacustrine oncoid carbonate coatings on clasts (Awramik et al., 2000) especially at the base of each unit, locally contain *Anodonta* shells, and may fine upward into muddy sand. In addition, they may exhibit gently dipping frontset (toward the lake) and backset beds of a barrier beach (Fig. 6B).

Steeply sloping active fan-lake margins are similar to the gently sloping margins in that they are characterized

by lake-modified alluvial deposits. Although such deposits locally exhibit oncoid coatings on clasts, they rarely contain fossils due to the higher-energy environment, and must be recognized on the basis of bedding, sorting, and stratigraphy. Such deposits are abundantly exposed in the deeply dissected drainages around and north of Shoreline Hill, in the eastern end of the Afton subbasin (Fig. 2). Here, an alluvial-fan complex was built on the steep eastern flank of Cave Mountain. After the integration of the Afton subbasin into Lake Manix, the style of deposition shifted abruptly from that of coarse-grained, angular clast-supported and debris-flow beds with depositional slopes of $5\text{--}7^\circ$ and common buried soils formed during periods of stability, to shoreline-modified fluvial beds (Fig. 7; Reheis et al., 2007b). Site M06-135 is unique in this area in exposing the architecture of a Gilbert-type fan-delta and lacustrine bar. The basal unit consists of interbedded moderately sorted, well-bedded medium-grained to coarse-grained sand and pebble to cobble gravel; the clasts have thin scattered oncoid coats, but the gravelly

M06-81, Midway Wash area, distal fan

A



M08-131, E side Manix Wash

B

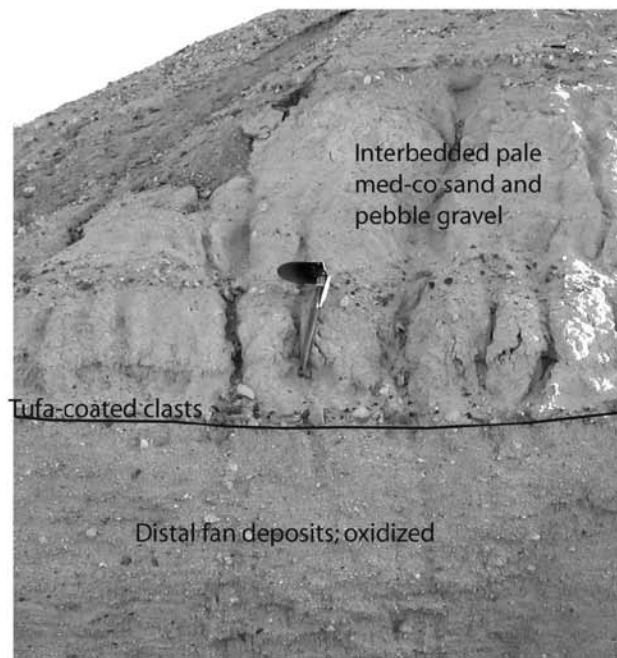
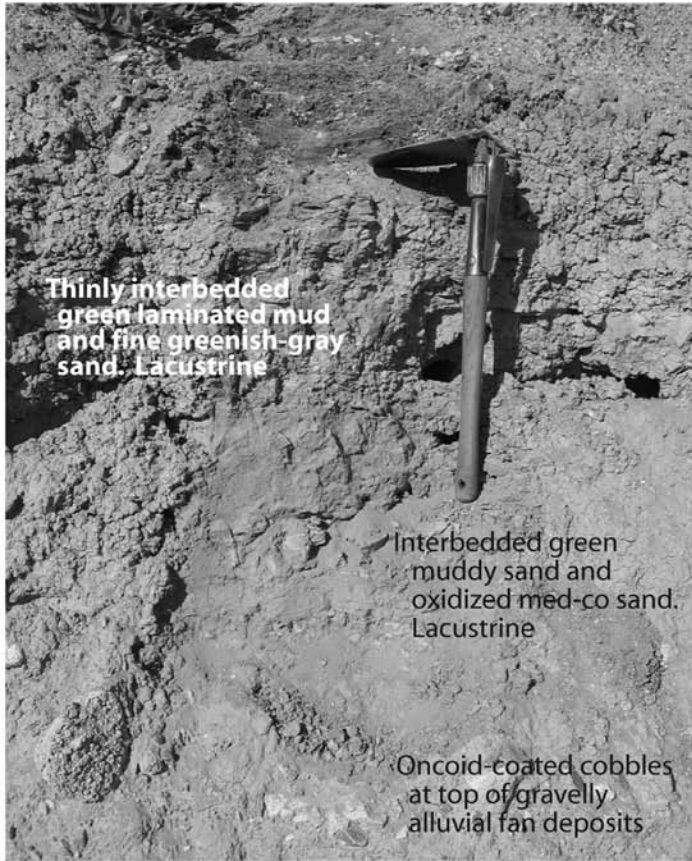


Figure 8. Deposits of distal fan-lake margins. A, distal-fan and overlying nearshore sediments at site M06-81 near Midway wash. Nearshore sediments are capped by a thin distal-fan unit and another transgressive sequence with laminated, deeper-water muds at top. Geologist, dog, and daypack for scale. B, distal-fan and overlying nearshore sediments at site M08-131 on the east side of Manix Wash. Note transgressive lag of oncoid carbonate-coated pebbles and cobbles at base. 50-cm trenching shovel for scale. Photographs by M. Reheis.

beds form channel fills (Fig. 7B). To the left and right, however, the channel gravel appears to grade into beach frontset and backset beds, respectively. Stratigraphically above this unit are sets of fining-upward, well-bedded sand and locally, pebble gravel arranged in alternating sequences of bar structures with frontset and backset beds and deltaic structures with foreset and bottomset beds (Fig. 7A).

Other exposures of the steeply sloping fan-lake environment are more like that at site M06-141, where deposits of two lake cycles overlie dipping, indurated older fan deposits capped by a strong calcic buried soil ("old fan," Fig. 7C). The basal unit consists of two parts: the lower part is 2 m of stratified, interbedded moderately to poorly sorted sand and lesser angular pebble-cobble gravel, with local thin lenses of sandy mud, interpreted as alluvial-fan deposits at the photograph site. However,

M05-19, basal part; upper Dunn Wash **A**



M05-6, S flank Buwalda Ridge **B**

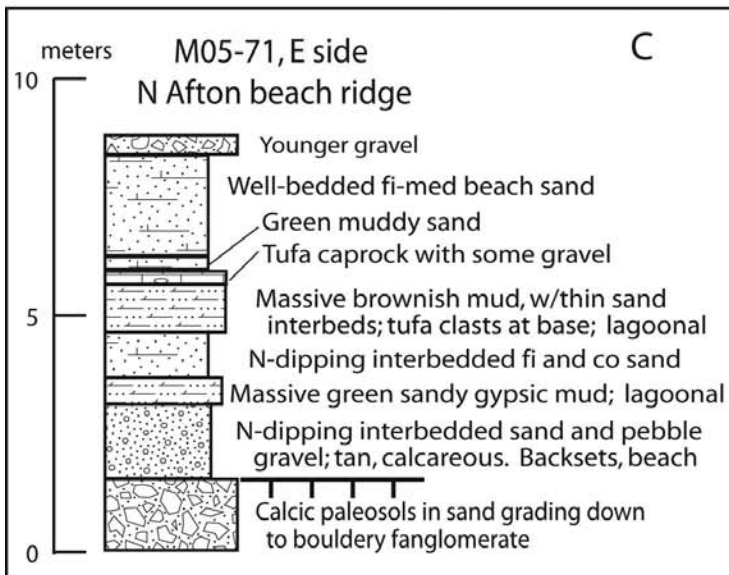


Figure 9. Deposits of medial to proximal fan-lake margins. A, photograph of basal part of section M05-19 (see measured section in Fig. 5A). 50-cm trenching shovel for scale. B, beach sand and gravel overlie oncooid-coated clasts at top of a prominent, red calcic soil formed on alluvial-fan gravel; many clasts are rotted. 20-cm trowel for scale. C, measured section M05-71 on east side of North Afton beach ridge. Beach gravel was deposited during initial transgression over calcic soils formed in massive eolian (?) sand. Photographs by M. Reheis.

downstream (to the right) these beds become sandier and better sorted, suggesting this lower unit was deposited in a lake-marginal setting. This lower part is capped by about 30 cm of coarse-grained gravel with abundant thin oncooid carbonate coats, and this in turn is overlain by the upper part, about 2 m of interbedded, fining-upward packets of angular to subangular cobble gravel and coarse-grained

to medium-grained sand capped by a buried soil. Beds in this upper part dip 10° and fine in a downslope direction, and are interpreted as delta foresets. The upper unit above the buried soil is overall sandier and better sorted and bedded than the lower unit, but still contains channel-fill gravels.

In summary, active alluvial fan-lake hybrid deposits are expected to vary laterally on scales of tens to hundreds of meters. Indicators of lake interactions can be very subtle, but commonly take the form of increased lateral persistence of beds, greater size sorting, and thin oncoid clast coats. Gently sloping fan-lake marginal deposits may preserve lacustrine fossils.

3. Stable fan-lake interface deposits

Away from active alluvial-fan channels, fan surfaces are relatively stable with little aggradation. On distal fan slopes, intermittent sheetwash flow may cause slow sedimentation such that sediments are generally somewhat oxidized throughout, but soil horizons are weak to absent. On medial to proximal parts of fans, surfaces are generally stable enough that soils form, with typical desert pavement. These different substrates influence the depositional environment of the fan-lake interface.

Distal fan-lake margins generally exhibit sharp transitions in color, induration, and sorting of sediments from the fan to the lake environment, but little change in particle-size distribution (Fig. 8). Distal fan deposits consist mainly of planar-bedded, moderately to poorly

sorted silt, sand, and fine-pebble gravel with scattered larger clasts, generally arranged in fining-upward sets and parallel to fan slope, with little or no channeling apparent. They are generally slightly to moderately indurated. Lake transgression across this surface may produce a single-clast-thick lag deposit (Fig. 8B) that locally has very thin oncoid carbonate coats. The transgressive surface is overlain by a generally fining-upward sequence of non-indurated, moderately to well-bedded and sorted pebble gravel, sand, and silt that is planar bedded and may be wave-ripple-laminated (Fig. 8A). The finer-grained sands commonly contain lacustrine ostracodes.

Medial to proximal fan-lake margins are easily recognized by abrupt contacts between oxidized, poorly to moderately sorted, coarse-grained gravelly fan deposits, typically with preserved soils, and overlying reduced, well-bedded, fine-grained lacustrine deposits (Fig. 9). In such cases, the sediment is likely supplied by long-shore drift along with minimal reworking of the local fan surface, such that the lacustrine sequence is constructed on top of an inactive fan. The transgressive surface typically has a concentration of clasts with prominent oncoid

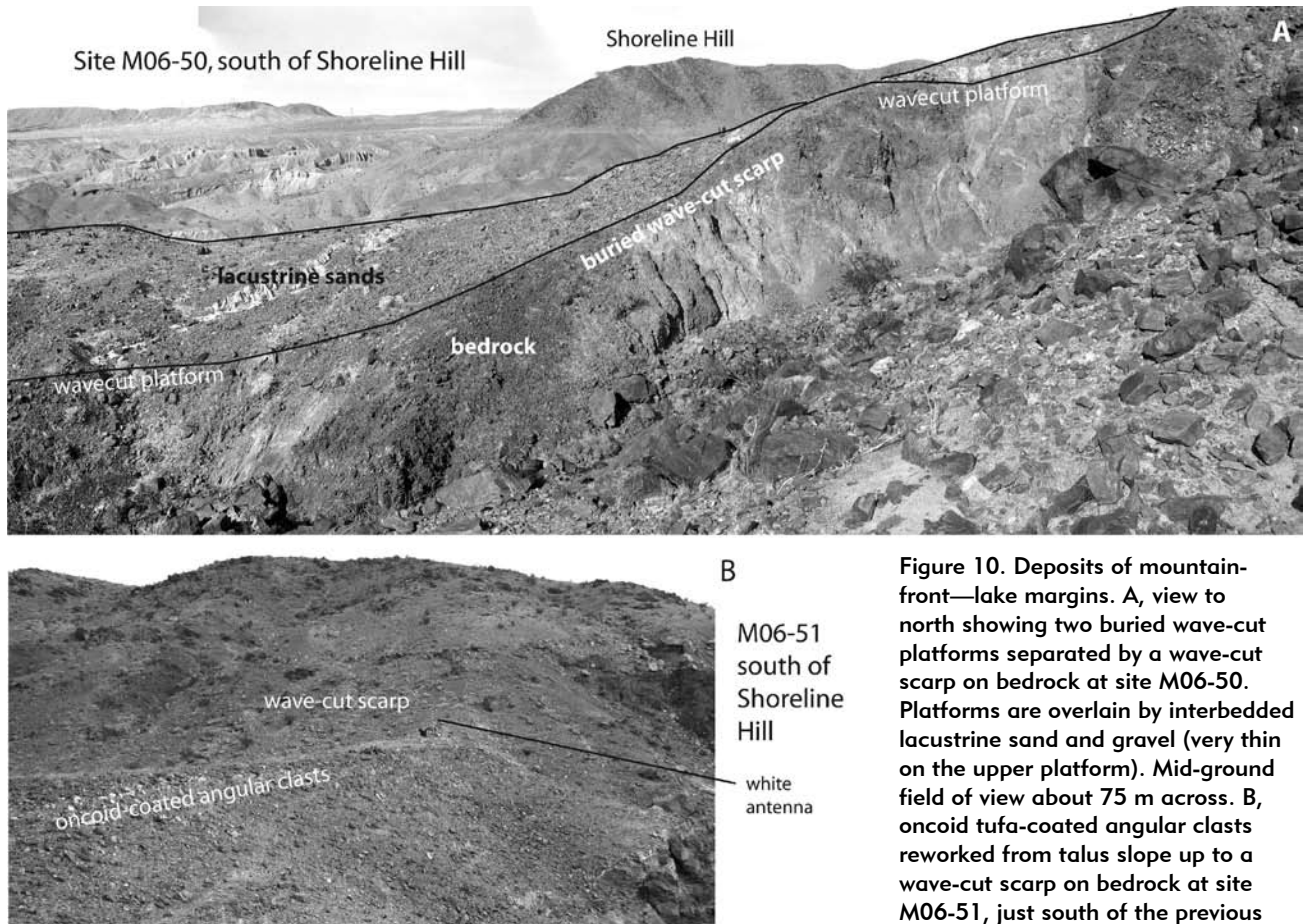


Figure 10. Deposits of mountain-front—lake margins. A, view to north showing two buried wave-cut platforms separated by a wave-cut scarp on bedrock at site M06-50. Platforms are overlain by interbedded lacustrine sand and gravel (very thin on the upper platform). Mid-ground field of view about 75 m across. B, oncoid tufa-coated angular clasts reworked from talus slope up to a wave-cut scarp on bedrock at site M06-51, just south of the previous site. 2-m white antenna and black dog for scale. Photographs by M. Reheis.

coats (Figs. 9A and B); oncoid carbonate coats tend to be thicker on larger clasts and where underlain by buried calcic soils. Overlying lacustrine deposits are typically well-bedded and sorted, fining-upward pebble gravel, sand, silt, laminated where finer-grained. Some locations (e.g., M05-71, Fig. 9C) exhibit dipping interbedded sand and gravel representing a transgressive beach setting, locally overlain by finer-grained green or brown muddy sand and mud that are interpreted as lagoonal deposits on the landward side of a beach berm.

4. Mountain-front deposits

Steep bedrock slopes are erosional environments, and lakes that abut such slopes have limited detritus to rework and deposit, as well as reduced longshore transport. In addition, sediments that are deposited have low potential for preservation and are probably largely absent from the rock record. At the eastern end of the Afton subbasin (Fig. 2), where wave energy was high due to the long fetch, a few lacustrine deposits that interfinger with talus and very steep small fans are preserved (Fig. 10). These deposits consist of interbedded, moderately bedded and sorted gravel and sand, commonly overlying a wave-eroded shoreline platform cut on rock (Fig. 10A) or fan deposits and locally with oncoid-coated clasts at the base (Fig. 10B). The angular clasts show little or no sign of rounding. Fossils are rare or absent due to the high-energy environment, and typically no beach-barrier bedding is preserved.

Conclusions

Nearshore settings of Lake Manix vary in sedimentary character and stratigraphic architecture, depending on proximity to the mouth of the Mojave River, to position on fringing alluvial fans, or to bedrock. Depositional features are also related to steepness of slope on the fan, presence of active channels, and length of fetch. These lake-marginal features are by no means unique to Lake Manix, but they have not been well characterized for lakes in low-desert settings, where the principal focus has been on interpreting lake fluctuations from cores taken in the deeper parts of the basins. Some of the most useful markers in Lake Manix deposits are the oncoid carbonate coats in transgressive settings. Tufa-coated clasts have been reported in Lake Mojave beach deposits (Ore and Warren, 1971; Wells et al., 2003) and in many other Great Basin lakes, but their association with lake transgression has not been previously noted. Although some of the settings we have described are enigmatic (i.e., the fluvial-mudflat setting), are initially difficult to recognize (the active fan-lake margin), or have low preservation potential (the mountain-front deposits), valuable information on response of desert lakes to past climate change can be

obtained by using such features for reconstruction of lake altitude.

References cited

- Adams, K.D., 2007, Late Holocene sedimentary environments and lake-level fluctuations at Walker Lake, Nevada, USA: *Geological Society of America Bulletin*, v. 119, p. 126-139.
- Adams, K.D., and Wesnousky, S.G., 1998, Shoreline processes and the age of the Lake Lahontan highstand in the Jessup embayment, Nevada: *Geological Society of America Bulletin*, v. 110, p. 1318-1332.
- Allen, B.D., and Anderson, R.Y., 2000, A continuous, high-resolution record of late Pleistocene climate variability from the Estancia basin, New Mexico: *Geological Society of America Bulletin*, v. 112, p. 1444-1458.
- Awramik, S.M., Buchheim, H.P., Leggett, L., and Woo, K.S., 2000, Oncoids of the late Pleistocene Manix Formation, Mojave Desert region, California, in Reynolds, R.E., and Reynolds, J., eds., *Empty basins, vanished lakes: San Bernardino County Museum Association Quarterly*, v. 47(2), p. 25-31.
- Bright, J., Reheis, M.C., and Miller, D.M., 2006, Ostracode faunal and initial stable isotope (O and C) analysis of a 45-m-long core from Manix basin, California: *Geological Society of America Abstracts with Programs*, v. 39, no. 6, p. 271.
- Ellsworth, E.W., 1932, *Physiographic History of the Afton Basin*: Palo Alto, Stanford, PhD thesis, 99 p.
- Enzel, Y., Brown, W.J., Anderson, R.Y., McFadden, L.D., and Wells, S.G., 1992, Short-duration Holocene lakes in the Mojave River drainage basin, southern California: *Quaternary Research*, v. 38, p. 60-73.
- Enzel, Y., and Wells, S.G., 1997, Extracting Holocene paleohydrology and paleoclimatology information from modern extreme flood events: An example from southern California: *Geomorphology*, v. 19, p. 203-226.
- Enzel, Y., Wells, S.G., and Lancaster, N., 2003, Late Pleistocene lakes along the Mojave River, southeast California, in Enzel, Y., Wells, S.G., and Lancaster, N., eds., *Paleoenvironments and paleohydrology of the Mojave and southern Great Basin Deserts: Geological Society of America Special Paper 368*, p. 61-77.
- Forester, R.M., Lowenstein, T.K., and Spencer, R.J., 2005, An ostracode based paleolimnologic and paleohydrologic history of Death Valley: 200 to 0 ka: *Geological Society of America Bulletin*, v. 117, p. 1379-1386.
- Gilbert, G.K., 1890, *Lake Bonneville*: U.S. Geological Survey Monograph 1, 438 p.
- Jayko, A.S., Forester, R.M., Kaufman, D.S., Phillips, F.M., Yount, J.C., McGeehin, J., and Mahan, S.A., 2008, Late Pleistocene lakes and wetlands, Panamint Valley, Inyo County, California, in Reheis, M.C., Hershler, R., and Miller, D.M., eds., *Late Cenozoic drainage history of the southwestern Great Basin and lower Colorado River region: Geologic and biotic perspectives: Geological Society of America Special Paper 439*, p. 151-184.
- Jefferson, G.T., 1987, The Camp Cady local fauna: Paleoenvironment of the Lake Manix basin, in Reynolds, J., ed., *Proceedings of 1987 Mojave Desert Quaternary Research Symposium: San Bernardino County Museum Association Quarterly*, v. 34 (3-4), p. 3-35.
- Jefferson, G.T., 2003, Stratigraphy and paleontology of the middle to late Pleistocene Manix Formation, and paleoenvironments of the central Mojave River, southern California, in Enzel, Y., and Wells, S.G., eds., *Paleoenvironments and paleohydrology of the Mojave and southern Great Basin Deserts: Geological Society of America Special Paper 368*, p. 43-60.
- Knott, J.R., Machette, M.N., Klinger, R.E., Sarna-Wojcicki, A.M., Lid-

- dicoat, J.C., Tinsley, J.C., III, David, B.T., and Ebbs, V.M., 2008, Reconstructing late Pliocene to middle Pleistocene Death Valley lakes and river systems as a test of pupfish (*Cyprinodontidae*) dispersal hypotheses, *in* Reheis, M.C., Hershler, R., and Miller, D.M., eds., Late Cenozoic drainage history of the southwestern Great Basin and lower Colorado River region: Geologic and biotic perspectives: Geological Society of America Special Paper 439, p. 1-26.
- Komar, P.D., 1998, *Beach Processes and Sedimentation* (2nd ed.): Englewood Cliffs, New Jersey, Prentice-Hall, 544 p.
- Lowenstein, T.K., Li, J., Brown, C., Roberts, S.M., Ku, T.-L., Luo, S., and Yang, W., 1999, 200 k.y. paleoclimate record from Death Valley salt core: *Geology*, v. 27, p. 3-6.
- Meek, N., 1990, Late Quaternary Geochronology and Geomorphology of the Manix Basin, San Bernardino County, California: Los Angeles, University of California, PhD thesis, 212 p.
- Meek, N., 1994, The stratigraphy and geomorphology of Coyote basin, *in* Reynolds, J., ed., *Calico, Coyote Basin & Lake Havasu Giants*: San Bernardino County Museum Association Quarterly, v. 41, no. 3, p. 5-13.
- Meek, N., 2000, The late Wisconsinan history of the Afton Canyon area, Mojave Desert, California, *in* Reynolds, R.E., and Reynolds, J., eds., *Empty Basins, Vanished Lakes: The Year 2000 Desert Symposium Field Guide: Redlands*, San Bernardino County Museum Association Quarterly, p. 32-34.
- Meek, N., 2004, Mojave River history from an upstream perspective, *in* Reynolds, R.E., ed., *Breaking Up--the 2004 Desert Symposium Field Trip and Abstracts*: Fullerton, Calif., California State University, Desert Studies Consortium, p. 41-49.
- Menges, C.M., 2008, Multistage late Cenozoic evolution of the Amargosa River drainage, southwestern Nevada and eastern California, *in* Reheis, M.C., Hershler, R., and Miller, D.M., eds., Late Cenozoic drainage history of the southwestern Great Basin and lower Colorado River region: Geologic and biotic perspectives: Geological Society of America Special Paper 439, p. 39-90.
- Metcalfe, S., Say, A., Black, S., McCulloch, R., and O'Hara, S., 2002, Wet conditions during the last glaciation in the Chihuahuan Desert, Alta Babicora basin, Mexico: *Quaternary Research*, v. 57, p. 91-101.
- Meyers, J.S., 1962, Evaporation from the 17 western states: U.S. Geological Survey Professional Paper 272-D, p. 71-100.
- Miller, D.M., and McGeehin, J.P., 2007, Depositional environment and radiocarbon chronology of the upper Lake Manix deposits, Mojave Desert, California: Geological Society of America Abstracts with Programs, v. 39, no. 6, p. 271.
- Morrison, R.B., 1991, Quaternary stratigraphic, hydrologic, and climatic history of the Great Basin, with emphasis on Lakes Lahontan, Bonneville, and Tecopa, *in* Morrison, R.B., ed., *Quaternary Nonglacial Geology: Conterminous U.S.: Boulder, Colorado*, Geological Society of America, *The Geology of North America*, v. K-2, p. 283-320.
- Ore, H.T., and Warren, C.N., 1971, Late Pleistocene-early Holocene geomorphic history of Lake Mojave, California: *Geological Society of America Bulletin*, v. 82, p. 334-346.
- Oviatt, C.G., Reheis, M.C., Miller, D.M., and Lund, S.P., 2007, Stratigraphy of Lake Manix beds in the Manix subbasin, Mojave Desert, CA: Geological Society of America Abstracts with Programs, v. 39, no. 6, p. 271.
- Phillips, F.M., 2008, Geological and hydrological history of the paleo-Owens River drainage since the late Miocene, *in* Reheis, M.C., Hershler, R., and Miller, D.M., eds., Late Cenozoic drainage history of the southwestern Great Basin and lower Colorado River region: Geologic and biotic perspectives: Geological Society of America Special Paper 439, p. 115-150.
- Reading, H.G., and Collinson, J.D., 1996, Clastic coasts, *in* Reading, H.G., ed., *Sedimentary Environments: Processes, Facies, and Stratigraphy*: (3rd ed.) Oxford, Blackwell Science, Inc., p. 154-231.
- Reheis, M., Lund, S.P., and Bright, J., 2009a, Middle to late Pleistocene fluctuations of Lake Manix, Mojave Desert, California: Geological Society of America Abstracts with Programs, v. 41, no. 7, p. 656.
- Reheis, M.C., Miller, D.M., and McGeehin, J.P., 2009b, Late Pleistocene shoreline fluctuations of Lake Manix, Mojave Desert: Paleoclimate implications: PACLIM Conference Abstracts, Asilomar, April 2009.
- Reheis, M.C., Miller, D.M., and Oviatt, C.G., 2007a, Pluvial Lake Manix, Mojave Desert, USA: Effects of subbasin integration on sedimentary record: Quaternary International, Abstracts, XVII INQUA Congress, Cairns, Australia, p. 339-340.
- Reheis, M.C., Miller, D.M., and Redwine, J.L., 2007b, Quaternary stratigraphy, drainage-basin development, and geomorphology of the Lake Manix basin, Mojave Desert: Guidebook for Fall Field Trip, Friends of the Pleistocene, Pacific Cell, October 4-7, 2007: U.S. Geological Survey Open-File Report 2007-1281, 31 p.
- Reheis, M.C., and Redwine, J.L., 2008, Lake Manix shorelines and Afton Canyon terraces: Implications for incision of Afton Canyon, *in* Reheis, M.C., Hershler, R., and Miller, D.M., eds., Late Cenozoic drainage history of the southwestern Great Basin and lower Colorado River region: Geologic and biotic perspectives: Boulder, Geological Society of America Special Paper 439, p. 227-260.
- Russell, I.C., 1885, Geological history of Lake Lahontan, a Quaternary lake of northwestern Nevada: U.S. Geological Survey Monograph 11, 288 p.
- Smith, G.I., Barczak, V.J., Moulton, G., and Liddicoat, J.C., 1983, Core KM-3, a surface to bedrock record of late Cenozoic sedimentation in Searles Valley, California: U.S. Geological Survey Professional Paper 1256, 24 p.
- Smith, R.S.U., 1976, Late-Quaternary Fluvial and Tectonic History of Panamint Valley, Inyo and San Bernardino Counties, California: Pasadena, California Institute of Technology, PhD thesis, 295 p.
- Steinmetz, J.J., 1987, Ostracodes from the late Pleistocene Manix Formation, San Bernardino County, California, *in* Reynolds, J., ed., *Quaternary History of the Mojave Desert: San Bernardino, San Bernardino County Museum Association Quarterly*, p. 47-47.
- Talbot, M.R., and Allen, P.A., 1996, *Lakes*, *in* Reading, H.G., ed., *Sedimentary Environments: Processes, Facies, and Stratigraphy*: Oxford, Blackwell Science, Inc., p. 83-124.
- Waldmann, N., Stein, M., Ariztegui, D., and Starinsky, A., 2009, Stratigraphy, depositional environments and level reconstruction of the last interglacial Lake Samra in the Dead Sea basin: *Quaternary Research*, v. 72, no. 1, p. 1-15.
- Wells, S.G., Brown, W.J., Enzel, Y., Anderson, R.Y., and McFadden, L.D., 2003, Late Quaternary geology and paleohydrology of pluvial Lake Mojave, southern California, *in* Enzel, Y., Wells, S.G., and Lancaster, N., eds., *Paleoenvironments and paleohydrology of the Mojave and southern Great Basin Deserts*: Boulder, Colo., Geological Society of America Special Paper 368, p. 79-114

Stratigraphy and paleontology of the middle to late Pleistocene Manix Formation, and paleoenvironments of the central Mojave River, southern California

George T. Jefferson

*Colorado Desert District Stout Research Center, Anza-Borrego Desert State Park
200 Palm Canyon Drive, Borrego Springs, California 92004, gjefferson@parks.ca.gov*

Abstract

The Manix Formation consists of lacustrine, fluvial and alluvial sediments that were deposited within and adjacent to Lake Manix. This middle to late Pleistocene pluvial lake was probably the largest along the Mojave River drainage system, which heads in the Transverse Ranges of southern California and runs northeast to Death Valley. During high lake stands, the lake covered ~236 km² of the central Mojave Desert. Lake Manix deposits record much of the early hydrological history of the upper Mojave River.

Attaining an exposed thickness of ~40 m, the Manix Formation overlies playa lake deposits of the Pliocene-Pleistocene Mojave River Formation, and is locally overlain by latest Pleistocene and Holocene fluvial deposits. The Manix Formation is mappable as four laterally equivalent members. Generally from oldest to youngest these were deposited as: basin margin fan conglomerates, alluvial and fluvial deposits, lacustrine and paralic deposits, and fluvial/deltaic deposits. The interfingering relationships of these sediments, especially between the lacustrine and fluvial systems, documents at least four major transgressive/regressive lacustrine events within the last 0.5 m.y. Transgressive lacustrine events are in phase with relatively cool/moist regional climatic conditions inferred to be present during even-numbered marine oxygen isotope stages. The chronology of these deposits is relatively well constrained by paleomagnetic data, U/Th series ages, tephra chronology, and ¹⁴C ages.

Molluscan, ostracode, and aquatic and terrestrial vertebrate assemblages have been recovered primarily from the fluvial, paralic and lacustrine deposits within the formation. Included are 55 taxa of fossil vertebrates that range between 350+ ka and ca. 20 ka, spanning possibly late Irvingtonian through late Rancholabrean North American Land Mammal Ages. The assemblage comprises extinct and extralocal extant taxa that reflect climatic and biogeographic conditions dramatically different from the present xeric environment.

from Jefferson, G.T., 2003, Stratigraphy and paleontology of the middle to late Pleistocene Manix Formation, and paleoenvironments of the central Mojave River, southern California, in Enzel, Y., Wells, S.G., and Lancaster, N., eds., *Paleoenvironments and paleohydrology of the Mojave and southern Great Basin Deserts*: Boulder, Colorado, Geological Society of America Special Paper 368, p. 43-60. © 2003 Geological Society of America.

Introduction

Lake Manix was probably the largest in a chain of Pleistocene lakes that extended across the central Mojave Desert region of California along the Mojave River drainage (Fig. 1). The lake was located about midway between the headwaters of the Mojave River in the San Bernardino Mountains to the southwest, and Death Valley to the northeast, where the ancestral Mojave River episodically joined the Amargosa River and ended in Pleistocene Lake Manly. Lake Manly also was the terminus of the Owens River system, which drained the east side of the southern Sierra Nevada Mountains and basins within the northern Mojave Desert region (see Blackwelder, 1954; Blanc and Cleveland, 1961a, 1961b; Snyder et al., 1964).

With a highstand of 543 m above mean sea level, Lake Manix occupied a cloverleaf-shaped area of ~236 km² (Meek, 1990) including the present Coyote Lake and Troy Lake playa basins and the Afton basin (Fig. 2). Well-developed wave-cut terraces and beach bars typify its eastern and southern shorelines (Meek, 2000). The ancestral Mojave River began to fill the Lake Manix basin ca. 0.5 Ma, and Lake Manix persisted with several, apparently short-lived low level stands until the late Wisconsin.

Geologic and paleontologic investigations in Lake Manix basin began with Buwalda's (1914) description of the Pleistocene "Manix Beds of the eastern Mojave Desert." Buwalda described the stratigraphy, depositional history, and recognized the climatic implications of such deposits. A year later, in his summary of the extinct faunas of the Mojave Desert, Merriam (1915) confirmed Buwalda's age assessment and compared the terrestrial vertebrate fossils from Manix to those from Rancho La Brea

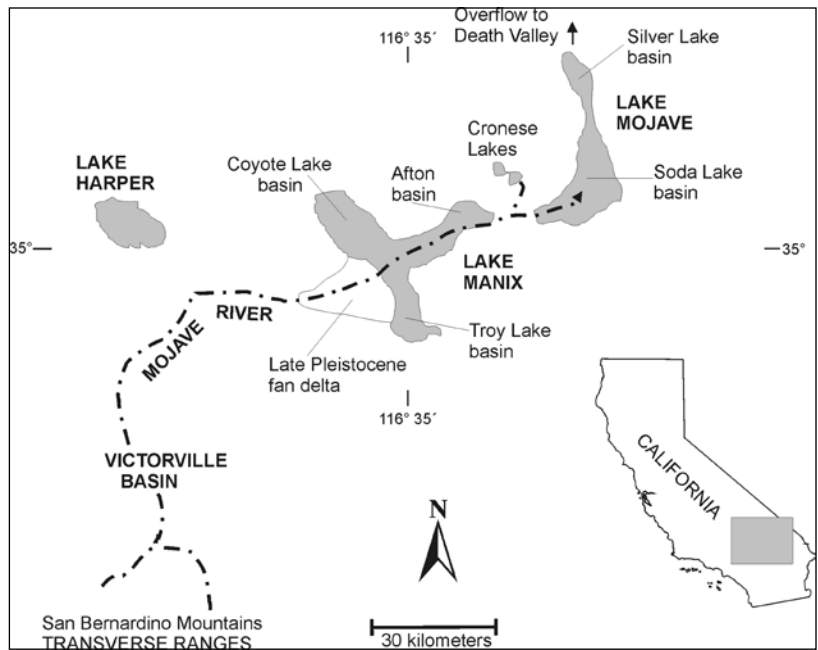


Figure 1. Regional map of Pleistocene lakes along the Mojave River drainage. Pleistocene lakes (shaded) are shown at highest stands. The modern Mojave River course is indicated by a long-short dashed line.

in Los Angeles, California. Since these early works, the geomorphology, stratigraphy, depositional history, and paleontology of the "Manix Beds" have been the focus of numerous studies. Manix basin and the long-lived lake it

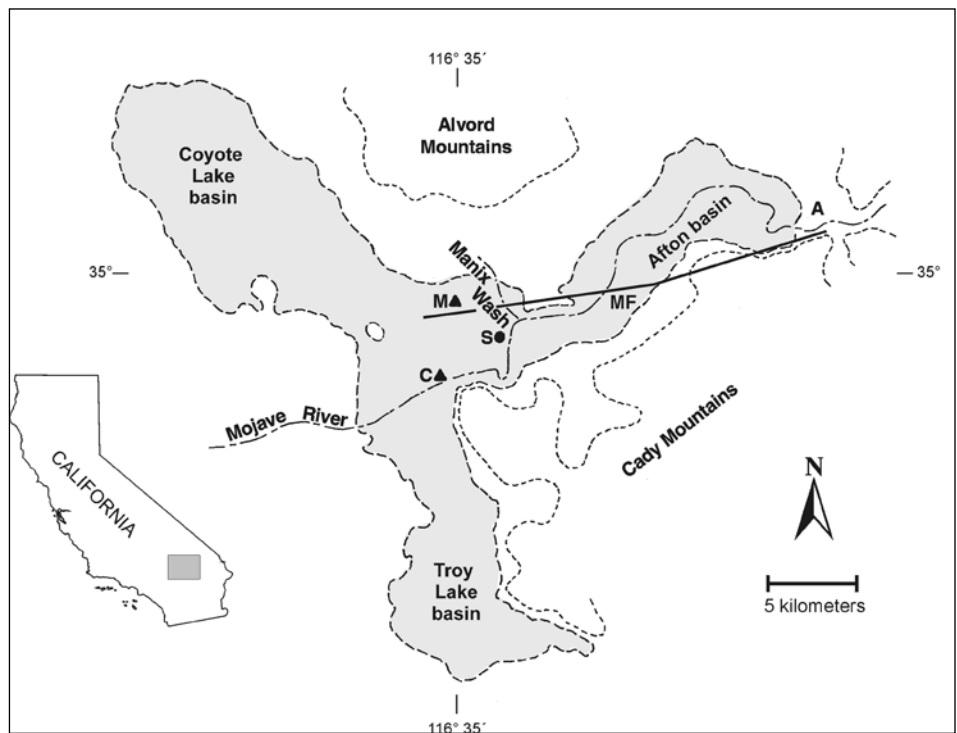


Figure 2. Generalized map of Lake Manix basin. The approximate maximum extent of the lake (shaded) is represented by a long dashed line at the 543 m topographic contour. Short dashed lines encompass mountains. The present Mojave River and Manix Wash drainages are represented by long-short dashed lines. Approximate trace of the Manix fault is shown as a solid line. A—Afton Canyon; C—Camp Cady; M—Manix railroad siding; MF—Manix fault; S—top of the reference stratigraphic section (see Figs. 6, 7)

held, derives its name from the Manix Union Pacific railroad siding (Fig. 2), 32 km east of the city of Barstow, San Bernardino County, California.

Latest Pleistocene and Holocene incision by the Mojave River and local tributaries such as Manix Wash, have exposed a 120+ m-thick section of alluvial fan, playa, lacustrine and fluvial sediments that records much of the Pleistocene history of the central Mojave River. Exposures in this area serve as the type localities for the Pliocene-Pleistocene Mojave River Formation (Nagy and Murray, 1991), the middle to late Pleistocene Manix Formation (Jefferson, 1985a, 1994, 1999), and the Camp Cady faunal assemblage (Winters, 1954; Howard, 1955; Jefferson, 1985b, 1987, 1991a; Seiple, 1994).

The upward-coarsening Pliocene-Pleistocene Mojave River Formation consists primarily of fine-grained sediments deposited in an internally drained playa basin (Nagy and Murray, 1991). These deposits rest on top of the late Miocene and Pliocene fanglomerates that initially filled Manix basin. Within the Manix Formation, which overlies the Mojave River Formation, locally derived alluvial and fluvial sediments, and lacustrine deposits largely transported by the ancestral Mojave River, appear to transgress/regress in consort with climate changes over the past 0.5 m.y. (Jefferson, 1985a, 1994, 1999).

The assemblage of invertebrate and vertebrate fossils recovered from eroded river bluffs and badlands of the Manix Formation dates from ca. 350 ka to 20 ka, and includes extinct and extralocal extant taxa that reflect a more equable climate and diverse biogeographic setting than at present (Buwalda, 1914; Merriam, 1915; Winters, 1954; Howard, 1995; Jefferson, 1985a, 1985b, 1987, 1991a; Jefferson and Steinmetz, 1986; Steinmetz, 1987, 1988; Seiple, 1994).

The following discussion is an amended synthesis of previous studies. It provides a summary of the stratigraphy and depositional history of the Manix Formation, and a brief discussion of deposits and events relevant to the history of Lake Manix and the paleohydrology of the Mojave River. Also included is an analysis of the fossil assemblage recovered from Lake Manix and its paleoenvironmental significance (Jefferson, 1987).

Pliocene-Pleistocene Stratigraphy

Basement rocks exposed around the margins of Manix basin are primarily pre-Cenozoic metamorphic rocks

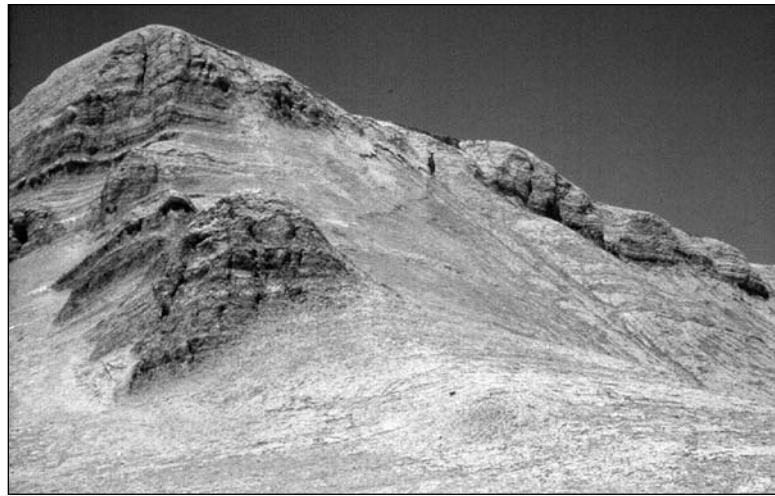


Figure 3. Mojave River Formation. The view is to the northwest from approximately 1 km east of the confluence of Manix Wash and the Mojave River. Gypsum-cemented pale gray-green siltstones and claystones form the more resistant horizons (also see exposures in mid-image of Fig. 5). Note geologist for scale.

and Tertiary volcanic and volcanoclastic deposits (Beyers, 1960; Bassett and Kupfer, 1964). The oldest fill within the basin consists of the late Miocene and Pliocene fanglomerates (older fanglomerate unit of Nagy and Murray, 1991; granitic fanglomerate of Meek and Battles, 1991). These fanglomerates are overlain by Pliocene-Pleistocene playa and fluvial sediments of the Mojave River Formation (Nagy and Murray, 1990, 1991). These units, in turn are usually overlain unconformably by fanglomerates, alluvial, fluvial, and lacustrine sediments of the middle to late Pleistocene Manix Formation (Buwalda, 1914; Ellsworth, 1933; Blackwelder and Ellsworth, 1936; Winters, 1954; Jefferson, 1985a, 1994, 1999; Budinger, 1992). Late Wisconsin and Holocene fluvial sediments (Hagar, 1966; Groat, 1967; Meek, 1990) both cap and are inset into the older sedimentary sequence.

Mojave River Formation, ca. 2.5–1 Ma

The Mojave River Formation consists primarily of gypsiferous, reddish-tan and light gray-green claystones and siltstones that total >80 m in thickness (Nagy and Murray, 1991) (Fig. 3). The base of this upward-coarsening sequence is not exposed in the type area, near the confluence of Manix Wash and the Mojave River (Fig. 2). However, 4 km to the east, these deposits overlie a 30+ m-thick section of fanglomerates (older fanglomerate unit of Nagy and Murray, 1991) that are exposed on the upthrown, north side of the Manix fault (Keaton and Keaton, 1977; McGill et al., 1988) (Fig. 4). To the south, toward the Cady Mountains, the deposits grade up section into fine to coarse-grained lithic arenites and granule to cobble conglomerates.

In the most northern exposures, these deposits are

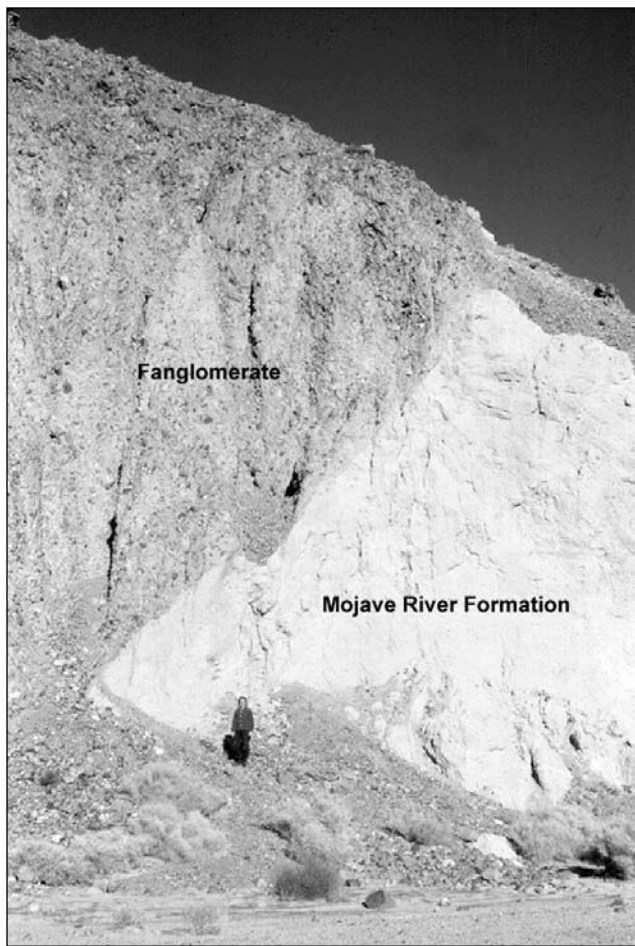


Figure 4. Fanglomerates and the Mojave River Formation. The view is to the northeast from the Mojave River bed about 4 km east of the mouth of Manix Wash. These units, medium gray fanglomerate (left) and pale reddish-tan siltstones and claystones of the Mojave River Formation (right), are in contact along one of several branches of the Manix fault.

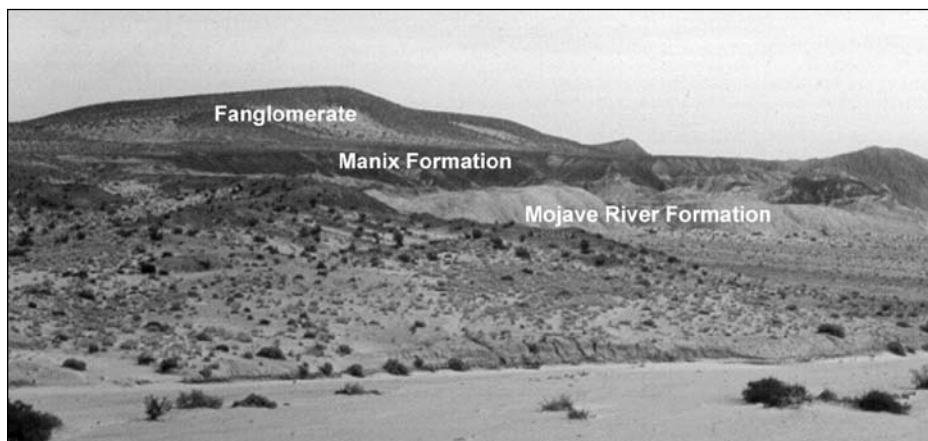


Figure 5. Mojave River Formation and Manix Formation. The view is to the east from Manix Wash (foreground). The gently folded Mojave River Formation appears mid-image and is unconformably overlain by the flat lying Manix Formation. Both units are in contact, along the Manix fault, with older fanglomerates (high left horizon).

severely deformed and upturned next to the Manix fault, and to the south they are folded into a broad east-west-trending syncline (Keaton and Keaton, 1977; McGill et al., 1988; Murray and Nagy, 1990; Nagy and Murray, 1991). Nagy and Murray (1991) maintain that the top of the unit grades up section through a transitional zone into the lowermost member of the Manix Formation (Member A of Jefferson, 1985a). Member A is lithologically similar to the upper Mojave River Formation, and some exposures exhibit this transitional relationship between the two formations (Jefferson, 1994, 1999). However, all other members of the mostly horizontal Manix Formation usually unconformably overlie the Mojave River Formation and all older units (Fig. 5).

Magnetostratigraphic correlations place the base of the Mojave River Formation at 2.48 Ma, and its top at either the Jaramillo (0.92–0.97 Ma) or the Cobb Mountain paleomagnetic events (1.1 Ma) (Pluhar et al., 1991). Sarna-Wojcicki et al. (1980) (Sarna-Wojcicki et al., 1984) correlate the lowermost fine-grained vitric gray ash that occurs in the top half of the Mojave River Formation with the Huckleberry Ridge ash bed (ca. 2.01 Ma, Izett, 1981), and the middle fine-grained vitric white ash with the slightly older Waucoba Beds sequence (ca. 2.3 Ma). The upper fine-grained vitric white ash of Sarna-Wojcicki et al. (1980) (Sarna-Wojcicki et al., 1984) remains to be positively correlated with dated tephra.

Nagy and Murray (1991) consider the presence of two sets of southwest-dipping cross strata, present in the top of the Mojave River Formation (Unit C2 at 19.1 m and 23 m, Nagy and Murray, 1991), as the earliest evidence of fluvial drainage in Manix basin, suggesting an open rather than a closed, internally drained system. However, these strata indicate a southwestward paleocurrent direction (Nagy and Murray, 1991). A change to the eastward flow

direction of the present Mojave River apparently postdates deposition of the Mojave River Formation.

Manix Formation, ca. 1–0.2 Ma

The middle to late Pleistocene Manix Formation consists of lacustrine, fluvial, and alluvial sediments that were deposited in and next to Lake Manix (Buwalda, 1914; Ellsworth, 1933; Blackwelder and Ellsworth, 1936; Winters, 1954; Jefferson, 1985a, 1994, 1999; Meek, 1990; Budinger, 1992).

Within the type area, near the confluence of Manix Wash and the Mojave River, these deposits unconformably overlie pre-Cenozoic basement rocks and Tertiary volcanic rocks, Miocene-Pliocene fanglomerates, and the Mojave River Formation. The Manix Formation is locally unconformably overlain by latest Pleistocene and Holocene lacustrine and fluvial sediments and Holocene eolian and slope deposits. The Manix Formation has a maximum exposed thickness of 41.5 m and is geologically mappable as four distinct Members A–D (Jefferson, 1985a, 1994, 1999; Budinger, 1992). Generally, from oldest to youngest these are: A, fanglomerates; B, alluvial/fluvial and lacustrine deposits; C, lacustrine and paralimnic deposits; and D, fluvial/deltaic deposits (Figs. 6, 7).

Member A. Member A is exposed south and southeast of Manix siding along both sides of the Mojave River (Fig. 2). The unit lies unconformably on metamorphic basement and volcanic rocks, and overlies the gravely, upper sediments of the Mojave River Formation. It is wedge-shaped, flanks the Cady Mountains, and dips gently to the north and northwest. To the southeast the deposit is >27 m-thick. The uppermost conglomerate beds of Member A interfinger with Member C to the west and north, and with Member B to the north and northeast (Fig. 7). The exposed base of the fanglomerate is estimated to be younger than ca. 0.9 Ma (Nagy and Murray, 1991).

Member A is composed primarily of dark to moderate brown, poorly sorted, subangular, cobble and boulder, clast-supported conglomerate interbedded with silty, fine to coarse-grained lithic arenites. Although locally cross stratified, the arenites are usually massive. Conglomerate beds are generally lenticular and exhibit lateral textural

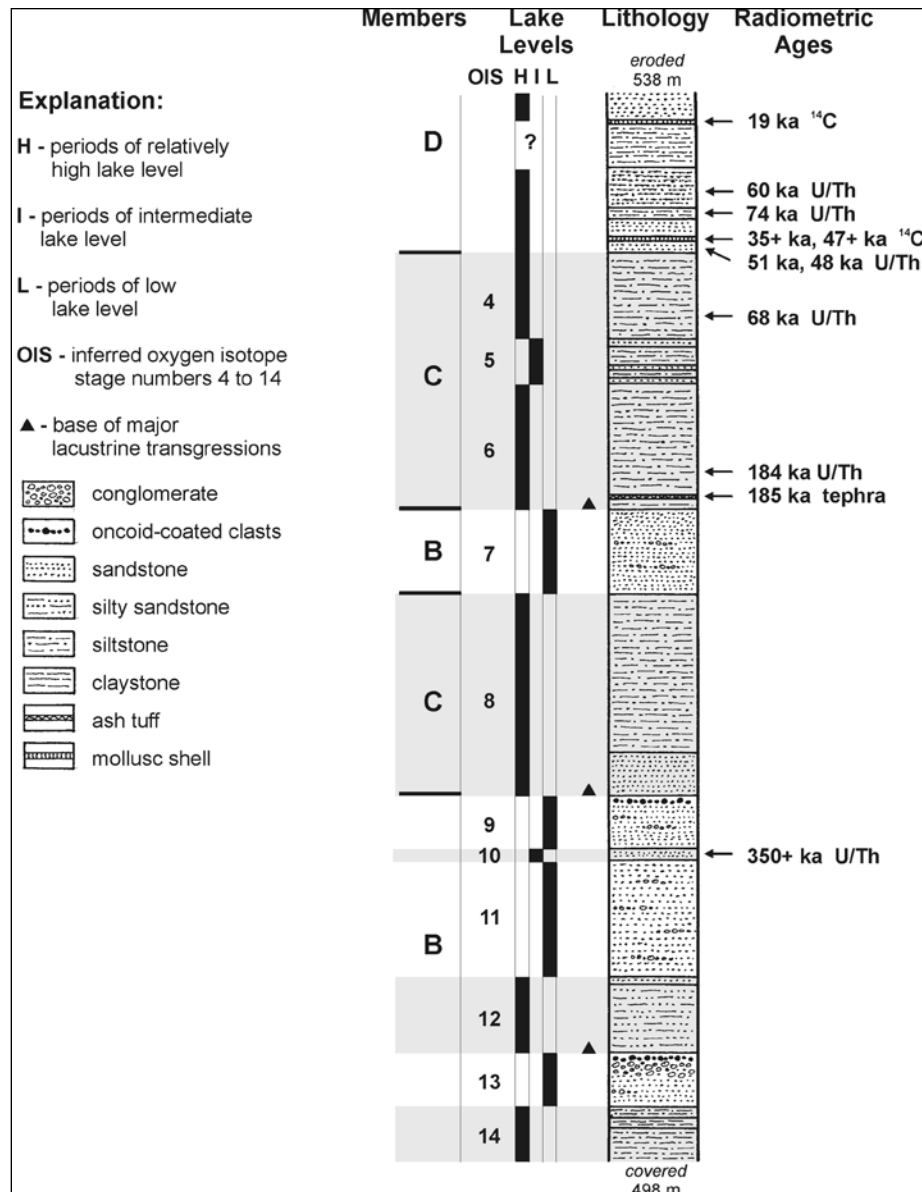


Figure 6. Reference stratigraphic section of the Manix Formation. Shown are Members B through D (see "S" Figs. 2, 7). The top of Member A is not exposed at this location, and only the last three of the four major lacustrine transgressions are indicated (□). The section totals 40 m, from 498 to 538 m above mean sea level. Lacustrine intervals within Members B and C are shaded. Periods of relatively high (H), intermediate (I), and low (L) lake levels are indicated by the vertical black bars. Radiometric determinations (Table 1) are discussed in the text

changes over very short distances. Clast imbrications generally indicate northwest paleocurrent directions, and clast lithologies are derived entirely from metamorphic and Tertiary volcanic and volcanoclastic rocks exposed to the south in the western Cady Mountains.

Member B. Member B crops out in Manix Wash, and south of Manix Wash along the west side of the Mojave River (Fig. 2). The unit conformably overlies Member A south of Manix Wash, and east of the southern part of Manix Wash, it unconformably overlies the Mojave River Formation. In Manix Wash, the unit is ~30 m thick. To

the south it bifurcates into upper and lower wedges (Fig. 7). The lower wedge averages 11 m in thickness and pinches out to the south between Members A and C. The upper wedge averages 2.5 m in thickness and interfingers with Member C and also pinches out to the south.

The base of the lower wedge consists of thinly bedded, pale olive claystones, gray-green siltstones, and fine-grained silty and platy micaceous arenites. These are overlain by coarse-grained arkose with lenses of granule to cobble conglomerate. Clast lithologies are predominately granitic. The uppermost clasts in this upward-coarsening sequence are coated with gray calcareous oncolid stromatolites, indicative of near-shore lacustrine environments typically formed during transgressive lacustrine events (Awramik et al., 2000) (Fig. 8). These sediments are overlain by light gray-green, thinly bedded silty arenite. The remaining upper part of the lower wedge is composed of grayish-orange to pale yellow-brown, silty coarse to medium-grained arkose and interbedded granule conglomerate.

Although both the lower and upper wedges of Member B thicken northward, to the south, the upper wedge generally occurs 13 m above the base of Member C (Fig. 7). It is composed of grayish-orange to pale yellow-

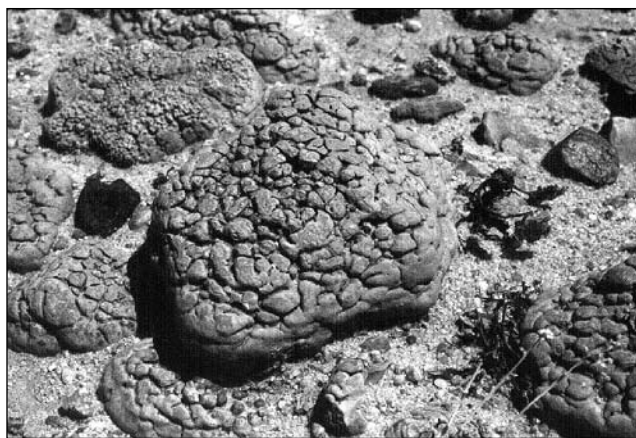


Figure 8. Oncolid-coated cobbles. The largest cobble is approximately 0.15 m in diameter. This exposed horizon occurs near the base of Member B, Manix Formation, at the inferred 13/12 marine oxygen isotope stage boundary, and represents the second major transgressive lacustrine event (Fig. 6). A similar horizon also is present at the base of Member C, 9/8 oxygen isotope stage boundary (Fig. 6). Highly rugose surfaces develop on the tops of the cobbles.

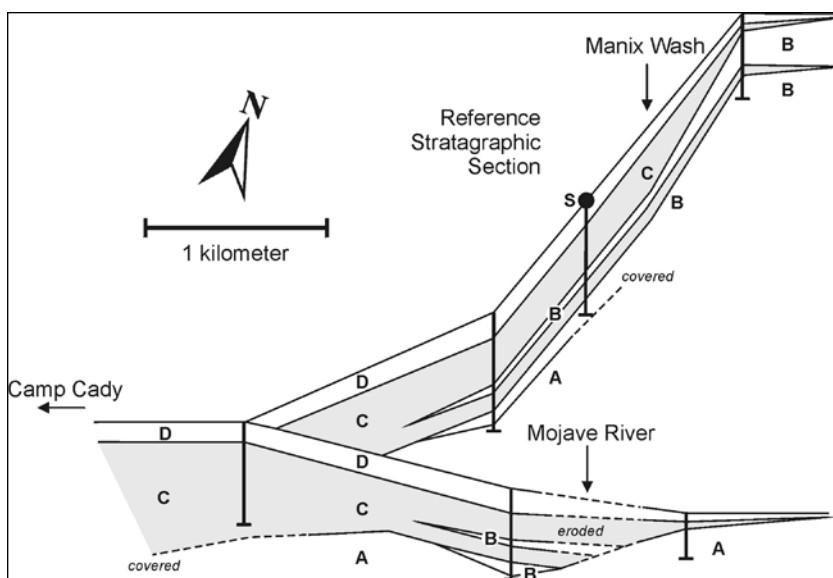


Figure 7. Panel diagram of the surficial exposures of the Manix Formation. Illustrated are the interdigitation and superposition of Members A through D. For location of the reference stratigraphic section (Fig. 6) see "S" in Figure 2. Other measured sections appear as vertical lines. Member C is shaded. The vertical scale is 10 × the horizontal scale. The area within the diagram extends from near Camp Cady, east then north 5 km along the Mojave River (see view in Fig. 9), and 3.5 km from the Mojave River north up Manix Wash drainage.

brown, silty coarse and medium sand-grained arkose and interbedded granule conglomerate. The conglomerates are cross stratified, indicating a southwest paleocurrent direction, and are scoured into the underlying siltstone of Member C. Clast lithologies are predominately granitic, and are probably derived from the east flank of the Alvord Mountains to the north. An erosional unconformity at the base of the upper part of Member C truncates the top of the wedge.

The lowest exposed beds of Member B are magnetically normal (Kirschvink, 1984, personal commun.), and, given their stratigraphic position, presumably fall within the midearly Brunhes paleomagnetic event. An infinite U/Th series age of 350+ ka (USGS 80-51) was obtained from an Equus ulna fragment (J.L. Bischoff, 1982, 1983, personal commun.) (Table 1) recovered from the top of the lower wedge of Member B, 5 m below the base of Member C (Fig. 6).

Member C. Member C crops out in bluffs along both sides of the Mojave River and Manix Wash (Fig. 2). Here it consists of >32 m of light gray, greenish-yellow and gray-green siltstone and claystone (Fig. 6). The unit is thickest in the middle of the basin and thins at the basin margin. The lowest exposures interfinger with both Members A (Fig. 9) and B. The upper part of Member C locally overlies and pinches out against Member A south of the Mojave River, and with Member B to the north along Manix Wash (Fig. 7).

Table 1. Radiometric and absolute dates from Lake Manix Basin

Age yr BP	Location	Material	Method	Laboratory	Source/s
11,810 ± 100	Coyote Lake basin	<i>Anodonta</i> shell	¹⁴ C	UCLA 2609C	Meek, 1990
12,900 ± 120	Coyote Lake basin	<i>Anodonta</i> shell	¹⁴ C	UCLA 2606	Meek, 1990
13,560 ± 145	Coyote Lake basin	<i>Anodonta</i> shell	¹⁴ C	UCLA 2609B	Meek, 1990
13,800 ± 600	Coyote Lake basin	<i>Anodonta</i> shell	¹⁴ C	La Jolla 958	Hubbs et al., 1965
14,230 ± 1,325*	Afton basin	<i>Anodonta</i> shell	¹⁴ C	UCLA 2601	Meek, 1989, 1990
15,025 ± 230	Troy Lake basin	<i>Anodonta</i> shell	¹⁴ C	UCLA 2605	Meek, 1990
15,125 ± 270	Coyote Lake basin	<i>Anodonta</i> shell	¹⁴ C	UCLA 2608	Meek, 1990
16,750 ± 1,000	central Manix basin	oncoïd stromatolite	¹⁴ C	UCLA 1079	Berger and Libby, 1967
17,950 ± 1,500	Coyote Lake basin	<i>Anodonta</i> shell	¹⁴ C	UCLA 6203	Meek, 1990
18,150 ± 400	Afton basin	<i>Anodonta</i> shell	¹⁴ C	UCLA 2607	Meek, 1990
19,100 ± 250	Manix basin D	<i>Anodonta</i> shell	¹⁴ C	QC 1467	R. Pardi, pers. comm., 1983; Jefferson, 1985a
19,300 ± 400	Afton basin	oncoïd stromatolite	¹⁴ C	UCLA 121	Fergusson and Libby, 1962
19,500 ± 500	Afton basin	oncoïd stromatolite	¹⁴ C	La Jolla 269	Hubbs et al., 1962
19,700 ± 260	central Manix basin	oncoïd stromatolite	¹⁴ C	UCLA 2600B	Meek, 1990
20,500 ± ?	Afton basin	<i>Anodonta</i> shell	¹⁴ C	Yale	Stuiver, 1969; Bassett and Jefferson, 1971
20,980 ± 345	central Manix basin	oncoïd stromatolite	¹⁴ C	UCLA 2602	Meek, 1990
21,300 ± 1,710	Afton basin	<i>Anodonta</i> shell	¹⁴ C	—	Meek, 1999
23,090 ± 445	central Manix basin	<i>Anodonta</i> shell	¹⁴ C	UCLA 2600A	Meek, 1990
28,960 ± 2,490	Afton basin	<i>Anodonta</i> shell	¹⁴ C	UCLA 2601C	Meek, 1999
29,310 ± 310	Afton basin	<i>Anodonta</i> shell	¹⁴ C	CAMS 1856	Meek, 1999
30,650 ± 890	Afton basin	<i>Anodonta</i> shell	¹⁴ C	UCLA 2604	Meek, 1990
30,950 ± 1,000	Afton basin	oncoïd stromatolite	¹⁴ C	La Jolla 895	Hubbs et al., 1965
35 + (infinite)	Manix basin D	<i>Anodonta</i> shell	¹⁴ C	UCLA	R. Berger, pers. comm., 1982; Jefferson, 1985a
47 + (infinite)**	Manix basin D	<i>Anodonta</i> shell	¹⁴ C	Yale	Stuiver, 1969; Bassett and Jefferson, 1971
47,700 ± 2.0	Manix basin D	<i>Mammuthus</i> bone	U/Th	USGS 81-48	J. Bischoff, pers. comm. 1982, 1983; Jefferson, 1985a
51,200 ± 2.5	Manix basin D	<i>Hemiauchenia</i> bone	U/Th	USGS 81-49	J. Bischoff, pers. comm. 1982, 1983; Jefferson, 1985a
60,300 ± ?	Manix basin D	Mammalia bone	U/Th	USGS	Budinger, 1992
68,000 ± 4.0	Manix basin C	<i>Camelops</i> bone	U/Th	USGS 81-51	J. Bischoff, pers. comm., 1982, 1983; Jefferson, 1985a
74,000 ± ?	Manix basin D	<i>Anodonta</i> shell	U/Th	USGS	Budinger, 1992
80,000 ± ?	Afton basin	oncoïd stromatolite	U/Th	—	Meek, 2000
183,800 ± 12.0	Manix basin C	<i>Camelops</i> bone	U/Th	USGS 81-30	J. Bischoff, pers. comm., 1982, 1983; Jefferson, 1985a
185,000 ± 15.0	Manix basin C	tephra	chemical	—	Bacon and Duffield, 1981; Izett, 1981
350 + (infinite)	Manix basin B	<i>Equus</i> bone	U/Th	USGS 80-51	J. Bischoff, pers. comm., 1982, 1983; Jefferson, 1985a

Notes: Ages are uncorrected. Manix Formation Members (B through D) are designated where known. Meek (1990) provides corrected ¹⁴C ages for analyses from Manix basin and a discussion of the reliability of both *Anodonta* and oncoïd stromatolite tufa determinations.

* This age has been revised, UCLA 2601C (Meek 1999). ** The previously reported age of 49 + ka for this sample (Jefferson, 1985a) is an error.

To the north, the lower and upper parts of Member C are separated by the upper wedge of Member B. These two lacustrine layers of Member C were attributed to two distinct lake phases by Buwalda (1914) and Win-

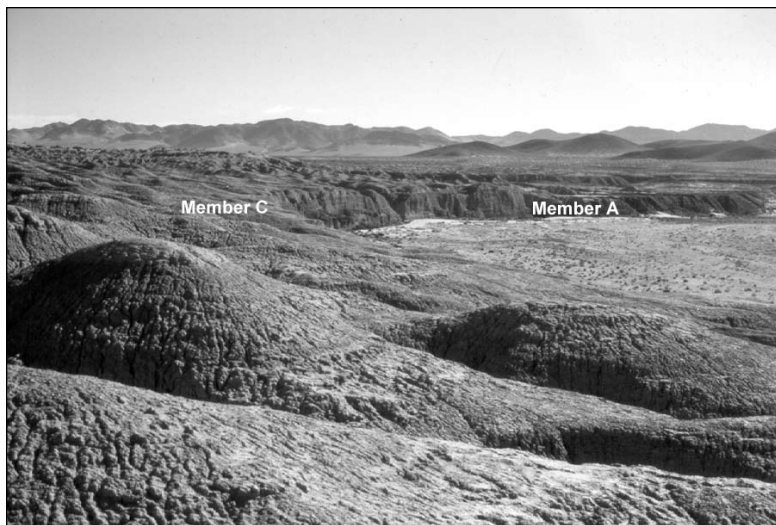


Figure 9. Eroded exposures of the Manix Formation. The view is to the east, from the north side of the Mojave River about 3 km east of Camp Cady. Light gray-green lacustrine silts and clays, Member C, are present in the foreground and to the left of the image. Fanglomerates of Member A form a 7 to 8 m-high bluff in the right mid-image. The Mojave River is to the right, and the Cady Mountains appear on the horizon.

ters (1954). The two layers are well exposed to the east throughout the Afton basin (Ellsworth, 1933; Blackwelder and Ellsworth, 1936). In eastern Afton basin, Member C interfingers with, and is locally overlain by alluvial fan deposits shed to the north from the eastern Cady Mountains (Blackwelder and Ellsworth, 1936; Meek, 2000).

A 0.5–1 m-thick pale yellowish-orange to grayish-orange, poorly bedded, moderately to well sorted, subrounded, medium-grained, quartz-rich arkose is commonly present where the base of lower Member C laps onto Member A and overlies the lower wedge of Member B. The arkose rests on light gray, rugose oncoïds that have grown on clasts at the top of Members A and B (Fig. 8). Much of the avian and terrestrial vertebrate assemblage has been recovered from this time-transgressive littoral horizon.

Three thin (0.1–0.15 m-thick), laterally persistent, grayish-orange, moderately well bedded, moderately well sorted, subrounded, medium-grained, quartz-rich arkose beds are present in the middle-upper part of Member C. These sandy horizons, like those at

the base of lower Member C, probably reflect littoral, near-shore depositional conditions (Fig. 6).

A 0.3–0.4 m-thick tephra of the Long Valley–Mono Glass Mountain family is present at the base of the upper part of Member C, 0.3 m above the top of the upper wedge of Member B (Fig. 6). It has been chemically correlated with tephra in the Long Canyon source area that date 185.0 ± 15.0 ka (Bacon and Duffield, 1981; Izett, 1981; Sarna-Wojcicki et al., 1984). This white, friable, well-sorted, fine-grained rhyolitic ash is composed of glass shards with minor amounts of biotite and quartz. A *Camelops* scapula fragment recovered 1.5 m above the ash provided a U/Th series age of 183.8 ± 12.0 ka (USGS 81–30) (J.L. Bischoff, 1982, 1983, personal commun.) (Table 1) (Fig. 6).

A U/Th series age of 68.0 ± 4.0 ka (USGS 81–51) (J.L. Bischoff, 1982, 1983, personal commun.) (Table 1), was obtained from a *Camelops* humerus fragment from the mid-upper part of Member C, which was recovered 1 m above the highest of the three near-shore arkosic sands (Fig. 6).

Member D. Member D conformably overlies Member C, and is overlain by late Pleistocene and Holocene deposits. The member pinches out to the east and north against Members A, B and C, and thickens from ~4.6 m in Manix Wash, to >7.7 m to the west near Camp Cady (Fig. 7). Along the Mojave River, east of Camp Cady, exposures of Member D are composed entirely of light brown, poorly to moderately sorted, silty medium to coarse-grained arkose, and granule to cobble conglomerate lenses. Clasts include a wide variety of igneous and metamorphic lithologies transported by the ancestral Mojave River.

South of Manix, Member D ranges in age from ca. 60 ka to 19 ka. U/Th series ages of 51.2 ± 2.5 ka (USGS 81-49) and 47.7 ± 2.0 ka (USGS 81-48) (J.L. Bischoff, 1982, 1983, personal commun.) were obtained from a *Hemiauchenia* radius/ulna and *Mammuthus* femur

Table 2. Camp Cady paleofauna taxonomic list

Order	Taxonomic identification	Common name	
Crustacea	<i>Limnocythere bradburyi</i>	ostracode, water flea	
	<i>L. ceriotuberosa</i>	ostracode, water flea	
	<i>L. platyforma</i>	ostracode, water flea	
	<i>L. robusta</i>	ostracode, water flea	
	<i>Heterocypris</i> sp.	ostracode, water flea	
Pelecypoda	<i>Anodonta californiensis</i>	freshwater clam	
	<i>Pisidium compressum</i>	freshwater clam	
Gastropoda	<i>Valvata humeralis</i>	freshwater snail	
	<i>Fossaria modicella</i>	freshwater snail	
	<i>Planorbella ammon</i>	freshwater snail	
	<i>P. subcrenata</i>	freshwater snail	
	<i>Planorbella</i> sp. ? <i>P. tenuis</i>	freshwater snail	
	<i>Carinifex newberryi</i>	freshwater snail	
	<i>Gyraulus vermicularis</i>	freshwater snail	
	<i>Gyraulus</i> sp.	freshwater snail	
	<i>Vorticifex effusa</i>	freshwater snail	
	Osteichthyes	<i>Gila bicolor mojaviensis</i>	tui (Mojave) chub
<i>Gasterosteus aculeatus</i>		threespine stickleback	
Reptilia	<i>Clemmys marmorata</i>	western pond turtle	
Aves	<i>Gavia</i> sp. cf. <i>G. arctica</i>	Arctic loon	
	<i>Podiceps</i> sp. cf. <i>P. nigricollis</i>	eared grebe	
	<i>Aechmophorus occidentalis</i>	western grebe	
	<i>Pelecanus</i> sp. aff. <i>P. erythrorhynchos</i>	American white pelican	
	<i>Phalacrocorax auritus</i>	double-crested cormorant	
	<i>Phalacrocorax macropus</i>	large-footed cormorant †	
	<i>Ciconia maltha</i>	stork †	
	<i>Phoenicopterus minutus</i>	small flamingo †	
	<i>Phoenicopterus copeii</i>	Cope's flamingo †	
	<i>Cygnus</i> sp. cf. <i>C. columbianus</i>	tundra swan	
	<i>Branta Canadensis</i>	Canada goose	
	<i>Anas</i> sp. cf. <i>A. crecca</i>	green-winged teal	
	<i>Anas</i> sp. cf. <i>A. platyrhynchos</i>	mallard	
	<i>Aythya</i> sp.	greater scaup or canvasback	
	<i>Mergus</i> sp. cf. <i>M. merganser</i>	common merganser	
	<i>Oxyura jamaicensis</i>	ruddy duck	
	<i>Haliaeetus leucophalus</i>	bald eagle	
	<i>Aquila chrysaetos</i>	golden eagle	
	<i>Fulica americana</i> cf. <i>F. a. minor</i>	small American coot †	
	<i>Grus</i> sp.	crane	
	cf. <i>Actitis</i> sp.	sandpiper	
	Phalaropodinae	phalarope subfamily	
	<i>Larus</i> sp. cf. <i>L. oregonus</i>	Oregon gull †	
	<i>Larus</i> sp.	gull (large-size)	
	<i>Bubo virginianus</i>	great horned owl	
	Mammalia	<i>Megalonyx</i> sp.	ground sloth (medium-size) †
		<i>Nothrotheriops</i> sp. cf. <i>N. shastensis</i>	Shasta ground sloth (small-size) †
<i>Paramylodon</i> sp.		ground sloth (large-size) †	
<i>Mammuthus</i> sp.		mammoth †	
<i>Lepus</i> sp.		jack rabbit	
Cricetidae		mice	
<i>Canis</i> sp. cf. <i>C. dirus</i>		dire wolf †	
<i>C. latrans</i>		coyote	
<i>Arctodus</i> sp.		short-faced bear †	
cf. <i>Ursus</i> sp.		black bear	
<i>Felis (Puma)</i> sp.		mountain lion	
<i>Homotherium</i> sp. cf. <i>H. crenatidens</i>		scimitar-tooth cat, robust †	
<i>Homotherium</i> sp. cf. <i>H. serum</i>		scimitar-tooth cat, gracile †	
<i>Equus conversidens</i>		horse (small-size) †	
<i>Equus</i> sp.		horse (large-size) †	
<i>Camelops</i> sp. cf. <i>C. hesternus</i>		yesterday's camel †	
<i>Camelops</i> sp. aff. <i>C. minidokae</i>		Minidoka camel †	
<i>Hemiauchenia macrocephala</i>		llama †	
Antilocapridae		prong bucks	
<i>Ovis canadensis</i>		mountain sheep	
<i>Bison</i> sp. cf. <i>B. antiquus</i>	antique bison †		

Note: Data are in part from Jefferson (1985a, 1987). † extinct taxon

fragment recovered from within 0.5 m above the base of Member D (Fig. 6). *Anodonta* shells from 0.7 m above the base of Member D have produced infinite ^{14}C ages of 47+ ka (Y-1993) (Stuiver, 1969; Bassett and Jefferson, 1971) and 35+ ka (R. Berger, 1982, personal commun.). A U/Th age of 74.0 ka on mammalian bone and a U/Th age of 60.3 ka on *Anodonta* shell are reported by Budinger (1992) 1.5 and 2.0 m above the base of Member D respectively (Table 1) (Fig. 6). Given its stratigraphic position, the former age appears to be too old. An *Anodonta* horizon, in the uppermost exposures of Member D south of Manix, has yielded a ^{14}C age of 19.1 ± 0.25 ka (QC-1467) (R. Pardi, 1983, personal commun.). Bivalve remains from stratigraphically correlative strata in eastern Afton basin have produced ^{14}C dates that range in age from ca. 31 ka to 28 ka and from ca. 21 ka to 18 ka (Table 1) (Bassett and Jefferson, 1971; Hastorf and Tinsley, 1981; Meek, 1990, 1999).

Paleontology of the Manix Formation

The aquatic invertebrate, and aquatic and terrestrial vertebrate taxa that compose the fossil assemblage from Lake Manix (Table 2) have been recovered from the fluvial sediments of both the lower and upper wedges of Member B, the littoral deposits at the base and middle of Member C, and the fluvial sediments of Member D of the Manix Formation. Invertebrate fossils are restricted to Members C and D, and not unexpectedly, the remains of fish and aquatic birds are restricted to Member C.

The fossiliferous part of the section ranges in age from >350 ka to ca. 20 ka, and spans late Irvingtonian through late Rancholabrean North American Land Mammal Ages (LMA). Most of the mammalian genera represented in the assemblage range through this entire period, and are found in Members B through D. However, the scimitar-tooth cat *Homotherium* and Antilocapridae occur only at the base of Member C, and *Bison* is restricted to the youngest deposits, Member D. Most vertebrate remains have been recovered from the base of Member C.

Jefferson (1968) named the mammalian fossil assemblage from the Manix Formation the "Camp Cady local fauna" after the historic Union Army post that was located on the Mojave River 3.7 km southwest of Manix siding (Fig. 2). Camp Cady was occupied during and shortly after the United States civil war (Chidester, 1965). In 1968, Lake Manix deposits were thought to encompass only the Wisconsin, and the entire assemblage was assigned to the Rancolabrean LMA (Winters, 1954; Howard, 1955; Jefferson, 1968). If retained, this name (see Walsh, 2000, p. 268–270) should apply to only those fossils recovered from the upper wedge of Member C and Member D.

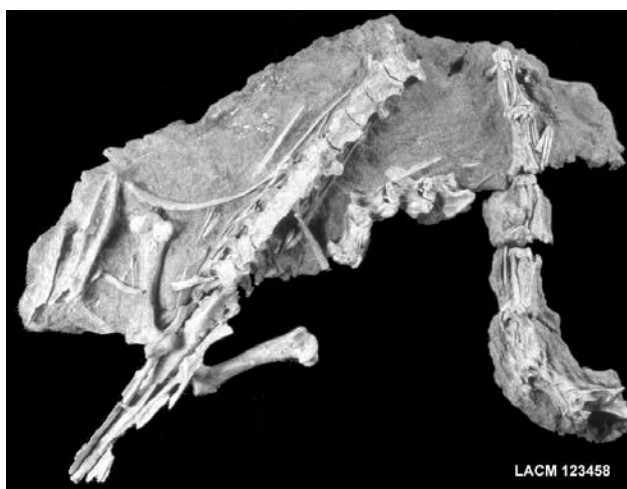


Figure 10. Partial skeleton of *Aechmophorus occidentalis*, western grebe (Natural History Museum of Los Angeles County specimen LACM 123458). The synsacrum and articulated left and right femora (46.1 mm in length) are to the lower left. Body and neck vertebrae occur to the right. A nearly complete skeleton of this animal was recovered from the base of Member C, Manix Formation (Jefferson, 1985b). Note the excellent preservation of delicate osteological structures that is typical of the avian remains from Member C.

The Lake Manix assemblage is significantly biased taphonomically toward large-sized mammals, although, excluding lower vertebrates, water fowl compose ~20% of the assemblage. Amphibians, lizards, snakes, insectivores, bats, small carnivores, and small ungulates are absent. However, these animals occur locally in other late Pleistocene assemblages like that from Schuiling Cave (Downs et al., 1959; Jefferson, 1983) ~10 km west of the southern edge of Troy Lake playa basin, or from the ancestral Mojave River deposits at Daggett and Yermo (Reynolds and Reynolds, 1985) ~20 km southwest of Camp Cady.

Avian remains from the lacustrine sediments are often partially articulated (Howard, 1955; Jefferson, 1985b) (Fig. 10). Those from the littoral deposits are not. Most mammalian fossils occur as single, isolated skeletal elements that often exhibit some abrasion due to fluvial transport and/or wave action. Few mammalian osteologic elements are articulated. Under fluvial or littoral high energy depositional conditions, estimates of relative taxonomic abundance within the assemblage (Table 3 and 4) are better represented as a percentage of the total number of identified specimens (NISP) for each taxon, rather than on a calculated minimum number of individuals (Horton, 1984; Badgley, 1986).

Large mammalian herbivores far outnumber carnivores in modern faunas (23:1, Mech, 1966; >250:1, Schaller, 1972). The herbivore/carnivore ratio for the larger mammals (larger than *Canis latrans*) in the Lake Manix assemblage is relatively well balanced with 672

NISP herbivores and 11 NISP carnivores (Table 4), or approximately a 61:1 ratio. All Carnivora in the assemblage are rare, totaling only 1.6% NISP of the mammalian assemblage. These large carnivores, like their modern counterparts, occupied a broad variety of habitats and geographic ranges of often continental scale.

Freshwater invertebrates and aquatic lower vertebrates

Fossil clams and snails have been recovered from several distinctive horizons within fluvial deposits near the base, middle and in the top of Member D (Winters, 1954; Jefferson, 1987, 1991a; Budinger, 1992), and in deposits of similar age in eastern Afton basin (Meek, 1999). Modern representatives of these fossils forms presently live in a variety of perennial freshwater habitats including streams, rivers, ponds, lakes and bogs or swamps. Some of the poorly sorted, coquina-like sediments in Member D that yield this ecologically diverse assemblage apparently were deposited during flood conditions on the Mojave River fan delta (Fig. 1).

Many of these molluscs are also known from other late Pleistocene localities in the western United States (Winters, 1954; Taylor, 1967), such as China Lake in Inyo County and Lake Cahuilla in Imperial County, California. Interestingly, the snail *Valvata humeralis* presently is endemic to lakes in the San Bernardino Mountains near the headwaters of the Mojave River.

Limnic ostracodes are common in the lacustrine silts and clays of Member C (Jefferson and Steinmetz, 1986; Steinmetz, 1987, 1988). Modern representatives of the five taxa identified (Table 2) indicate the following lake water conditions: mesotrophic, cool temperatures, well oxygenated, clear to muddy/turbid bottom, shallow to deep, and relatively fresh. In Lake Manix, ostracodes were apparently most abundant during mesotrophic conditions at high lake stands. Furthermore, the abundance of these ecologically sensitive organisms appears to peak with a periodicity of ca. 19 ka, suggesting a correlation of favorable habitat conditions with Milankovich-timed, even-numbered marine oxygen isotope stages (Steinmetz, 1987, 1988). The shaded bands on Figure 6 represent such lacustrine conditions.

The tui Mojave chub, *Gila bicolor mojavensis*, presently inhabits lakes and rivers along the Owens and Mojave River systems (Miller, 1973). This taxon is known from late Pleistocene deposits of the Death Valley river system, including the present Owens and Mojave River drainages (Miller, 1948, 1973). The Lake Manix fossils most closely resemble an endemic population living today at Zzyzx Spring (R.R. Miller, 1967, personal commun.) on the

Table 3. Relative abundance of avian taxa

Taxon	NISP	% NISP
<i>Gavia</i> sp. cf. <i>G. arctica</i>	1	0.7
<i>Podiceps</i> sp. cf. <i>P. nigricollis</i>	4	2.8
<i>Aechmophorus occidentalis</i>	41	29.5
<i>Pelecanus</i> sp. aff. <i>P. erythrorhynchos</i>	12	8.6
<i>Phalacrocorax auritus</i>	15	10.8
<i>Phalacrocorax macropus</i>	2	1.4
<i>Ciconia maltha</i>	6	4.3
<i>Phoenicopterus minutus</i>	14	10.1
<i>Phoenicopterus copei</i>	4	2.8
<i>Cygnus</i> sp. cf. <i>C. columbianus</i>	3	2.1
<i>Branta canadensis</i>	11	7.9
<i>Anas</i> sp. cf. <i>A. crecca</i>	1	0.7
<i>Anas</i> sp. cf. <i>A. platyrhynchos</i>	2	1.4
<i>Aythya</i> sp.	4	2.8
<i>Mergus</i> sp. cf. <i>M. merganser</i>	2	1.4
<i>Oxyura jamaicensis</i>	3	2.1
<i>Haliaeetus leucocephalus</i>	2	1.4
<i>Aquila chrysaetos</i>	2	1.4
<i>Fulica americana</i> cf. <i>F. a. minor</i>	1	0.7
<i>Grus</i> sp.	1	0.7
cf. <i>Actitis</i> sp.	1	0.7
Phalaropodinae	1	0.7
<i>Larus</i> sp. cf. <i>L. oregonus</i>	3	2.1
<i>Larus</i> sp.	2	1.4
<i>Bubo virginianus</i>	1	0.7

Note: Total number of identified specimens is 139, and total % NISP is 99.5. Data in part from Jefferson (1985b, 1987).
NISP = number of identified specimens
% NISP = number of identified specimens for each taxon divided by the total number of specimens
— = extinct taxon

southwestern side Soda Lake basin (Fig. 1).

The bones and scales of this fish are the most abundant vertebrate fossil remains from Lake Manix and probably represent many thousands of individuals. Their remains occur throughout the lacustrine clays and silts, and are locally concentrated in littoral sands at the base of Member C.

The stickleback, *Gasterosteus aculeatus*, is known from a single specimen recovered from a drill core in Coyote Lake basin (Roeder, 1985). This small fish presently occupies the Mojave River and rivers and streams in Los Angeles and Orange Counties, California.

Clemmys marmorata, the western pond turtle, presently inhabits fresh waters that occur in a narrow band along the West Coast of North America extending from British Columbia, Canada, to Baja California del Sur, Mexico. The eastern most margin of this range includes the Truckee and Carson River drainages of western Nevada and perennial portions of the Mojave River drainage including Afton Canyon (Lovich and Meyer, 2000). This turtle, which feeds mainly on aquatic plants and insects, typically is found in streams, ponds and lakes and prefers muddy bottoms and marshes (Stebbins, 1966). Although not abundant, plastron and carapace fragments of *C. marmorata* have been recovered from paralimnic deposits of Member C and fluvial deposits of Member D.

Aves

Most of the living species of birds in the assemblage (Howard, 1955; Jefferson, 1985b) presently range throughout southern California and are found seasonally

on inland lakes, such as the Salton Sea, Imperial County. The remaining species frequent coastal marine waters or inland areas from the San Joaquin Valley, central California northward (Pyle, 1961; Cogswell and Christman, 1977; Garrett and Dunn, 1981) (Table 2).

Extant species of migratory birds represented in the assemblage are presently absent from southern California during the summer (Pyle, 1961). They migrate northward in the spring following inland portions of the north-south Pacific Coast fly-way. During Pleistocene pluvial periods, this fly-way would have included the lakes of the Colorado and Mojave Deserts, those east of the Sierra Nevada Mountains, and the western part of Lake Lahontan (Snyder et al., 1964). Many of these species are also known from the late Pleistocene deposits of China Lake, Inyo County, California, and Fossil Lake, Lake County, southeastern Oregon (Jefferson, 1985b).

Two-thirds of the extant species (*Gavia arctica*, *Podiceps nigricollis*, *Aechmophorus occidentalis*, *Pelecanus erythrorhynchos*, *Phalacrocorax auritus*, *Mergus merganser*, *Haliaeetus leucophalus*, *Aquila chrysaetos*, and *Larus* spp.), represented by 80% NISP of the fossil avian specimens (Table 3), presently prefer or feed exclusively on small fish (Cogswell and Christman, 1977; Garrett and Dunn, 1981). *Gila bicolor mojavensis* represents an abundant food source for these predators. Modern representatives of the remaining taxa (*Cygnus columbianus*, *Branta canadensis*, *Anas crecca*, *Anas platyrhynchos*, *Aythya* sp., *Oxyura jamaicensis*, *Fulica americana*, *Grus* sp., *Actitis* sp.), feed on a variety of water plants and freshwater invertebrates (Cogswell and Christman, 1977; Garrett and Dunn, 1981). The owl, *Bubo virginianus*, is the only bird in the assemblage that feeds primarily on small mammals.

Mammalian herbivores

Among the larger herbivores, three taxa of ground sloth, *Megalonyx*, *Paramylodon*, and *Nothrotheriops*, are present. All are poorly represented in the assemblage (Table 4). *Megalonyx* is known from the West Coast and eastern two-thirds of the United States, and ranged from South America into Canada and Alaska. Stock (1925) suggested that this animal was adapted to forest or woodland habitats. The habitat and dietary preferences of *Paramylodon*, which also ranged through North and South America, are not well known. Given its association with other presumed grassland animals, Stock (1925, 1930) maintained that *Paramylodon* was a grazer. Based on analysis of the feeding mechanism in *Paramylodon*, Naples (1989) argued that it was a grazer-browser. Considering the varied environments within an extensive geographic range, both *Megalonyx* and *Paramylodon* were probably mixed feeders.

Nothrotheriops shastensis ranged throughout the

southwestern United States and northern Mexico. It is well represented in cave assemblages from northeastern California through southern Nevada, and Arizona where it browsed on desert shrubs and plants typical of a juniper-sage brush savannah habitat (Martin et al., 1961; Hansen, 1978). This selective folivore/browser (Naples, 1987) appears to have inhabited a broad spectrum of floral assemblages, but was limited latitudinally and altitudinally by minimum winter temperatures (McDonald et al., 1996).

Lepus sp. is the only small mammalian herbivore well represented in the assemblage. Jack rabbits typically inhabit grassy and brush covered areas. The order Rodentia is represented by a single cricetid humerus.

The relatively abundant but fragmentary remains of mammoth can not be identified to species. However, they probably represent the well-known middle and late Pleistocene form, *Mammuthus columbi* (= *M. imperator*). *M. columbi* ranged throughout the United States and likely browsed and/or grazed (Davis et al., 1984) in small herds similar to extant elephants. Evans (1961) suggested that juvenile mammoths were the favored prey of the scimitar-tooth cat, *Homotherium serum*. The American mastodon, *Mammot americanum*, is absent in the assemblage, and has not been identified from any Rancholabrean LMA sites within the Mojave Desert or Colorado Desert regions (Jefferson, 1991c).

Fossil horses are well represented in the assemblage (Table 4), and at least two forms, *Equus conversidens* and *Equus* sp. (large) have been identified. The occurrence

Table 4. Relative abundance of mammalian taxa, Manix Formation

Taxon	NISP	% NISP
<i>Megalonyx</i> sp.	1	0.1
<i>Nothrotheriops</i> sp. cf. <i>N. shastensis</i>	2	0.3
<i>Paramylodon</i> sp.	1	0.1
<i>Mammuthus</i> sp.	37	5.2
<i>Lepus</i> sp.	11	1.6
Cricetidae	1	0.1
<i>Canis</i> sp. cf. <i>C. dirus</i>	5	0.7
<i>Canis latrans</i>	5	0.7
<i>Arctodus</i> sp.	1	0.1
cf. <i>Ursus</i> sp.	1	0.1
<i>Felis</i> (<i>Puma</i>) sp.	3	0.4
<i>Homotherium</i> sp. cf. <i>H. crenatidens</i>	1	0.1
<i>Homotherium</i> sp. cf. <i>H. serum</i>	2	0.3
<i>Equus conversidens</i>	52*	7.4
<i>Equus</i> sp. (large-size)	67*	9.5
<i>Camelops</i> sp. cf. <i>C. hesternus</i>	371	54.7
<i>Camelops</i> sp. aff. <i>C. minidokae</i>	5	0.7
<i>Hemiauchenia macrocephala</i>	117	16.6
<i>Antilocapara</i> sp.	2	0.3
<i>Ovis canadensis</i>	17	2.5
<i>Bison</i> sp. cf. <i>B. antiquus</i>	1	0.1

Note: Total number of identified specimens is 703, and total % NISP is 101.5. Data are in part from Jefferson (1985a, 1987).
 NISP = number of identified specimens
 % NISP = number of specimens identified for each taxon divided by the total number of identified specimens
 * = may represent more than one species _ = extinct taxon

of extinct species of *Equus* has been used to infer the presence of grasslands (e.g., Jefferson, 1968). However, feral horses and burros are opportunistic feeders that prefer to graze, although, their diet may include up to 80% browse (Hansen, 1976; Ginnett, 1982; Ginnett and Douglas, 1982). Also, young horses from Rancho La Brea apparently were grazer-browsers (Akersten et al., 1988). These data suggest that extinct *Equus* may have periodically browsed as well as grazed, and was not restricted to grassland habitats.

Remains of the relatively small, stout-limbed extinct *Equus conversidens* have been recovered from Members C and D. The species ranged from northern Mexico through the central United States and the northern Great Plains, and has been identified in numerous sites in the Mojave Desert region (Jefferson, 1986, 1989, 1990, 1991b; Scott, 1997, 2000). Some small, specifically non-diagnostic specimens may represent an additional stilt-legged species of *Equus* in the assemblage (Scott, 1996, 1997).

Large fossil horse remains have been recovered from lower Member B through Member D. However, a lack of well-preserved cranial/dental specimens precludes assignment of this material to either the large Irvingtonian LMA form *Equus scotti* (Scott, 1998, 1999), or the large Rancho Labrean LMA form, *Equus* sp. cf. *E. occidentalis*. These horses are commonly recovered from middle and late Pleistocene sites respectively throughout the southwestern United States (Jefferson, 1986, 1989, 1990, 1991b; Scott, 1997, 1998, 1999). Both species are probably present in the assemblage; *E. scotti* from the older deposits and *Equus* sp. cf. *E. occidentalis* from the younger.

The extinct camel, *Camelops* sp. cf. *C. hesternus*, is by far the most abundant large herbivore in the assemblage (Table 4), ~55% NISP, and has been found in lower Member B through Member D. It is the most common large mammal in late Pleistocene assemblages of the Colorado Desert and Mojave Desert regions (Jefferson, 1986, 1989, 1990, 1991b). This is in marked contrast to its low abundance in late Pleistocene assemblages from intermontane or coastal southern California (Jefferson, 1988).

Camelops hesternus from Rancho La Brea, the only site where direct dietary data are available, seems to have fed on ~10% monocot (graze) and 90% dicot (browse) plants (Akersten et al., 1988). Although previously considered to be a grazer (e.g., Webb, 1965; Jefferson, 1968), *C. hesternus* was probably a browser or mixed feeder similar to the extant camel, *Camelus* (Gauthier-Pilters and Dagg, 1981). *Camelops minidokae* is represented by only a few specimens from upper Member B through Member D. Otherwise, this more northern form is unknown in

southern California Pleistocene assemblages (Jefferson, 1991c).

The llama, *Hemiauchenia macrocephala*, is moderately well represented in the assemblage (Table 4), and has been recovered from Members B through D. The taxon is a common member of late Pleistocene Mojave Desert (Jefferson, 1986, 1989, 1990, 1991b) and intermontane California assemblages like that from McKittrick, Kern County, California (Jefferson, 1988). However, it is very rare in coastal assemblages (Jefferson, 1988). This long-necked, cursorial llama was probably a grazer-browser adapted to open terrain.

Antilocaprids, either extinct *Tetrameryx* or extant *Antilocapra*, are represented by a single specimen recovered from the base of Member C. The other medium-sized browser, *Ovis canadensis*, is moderately well represented in the assemblage. It has been reported historically from the eastern Cady Mountains. Both antilocaprids and ovids are found in other late Pleistocene assemblages from the Mojave Desert (Jefferson, 1986, 1989, 1990, 1991b).

Bison antiquus is rare in the assemblage, and is positively known only from Member D. Pleistocene *Bison* remains are rare throughout the Mojave Desert region (Jefferson, 1986, 1989, 1990, 1991b), which is in contrast to coastal southern California where they are relatively common (Jefferson, 1988). Older bovid remains from the base of Member C, previously referred to this type Rancho Labrean LMA taxon (Jefferson, 1987, 1991a), probably represent a large *Camelops*.

Bison antiquus is closely related to the extant *B. bison*. *B. antiquus* is assumed to have had habits similar to the living form that is almost exclusively a grazer but will periodically browse (Meagher, 1973; Peden, 1976). However, analyses of chewed plant remains impacted into the fossettes and fossettids of dentitions from Rancho La Brea indicate that juvenile *B. antiquus* was a mixed feeder (Akersten et al., 1988).

Paleoenvironments

The Lake Manix assemblage includes extinct and extralocal extant forms that reflect ecological conditions dramatically different from the present xeric environment of the central Mojave Desert. Inferences based on the ecology of extant taxa that are closely related to the extinct forms allow paleoenvironmental reconstructions of the local lacustrine and terrestrial paleohabitats. Most of the fossil molluscs represent extant animals that live in a assortment of fluvial, lacustrine or paralic habitats. Some are extralocal, preferring cooler waters. However, others presently inhabit perennial waters in the Mojave Desert region. Although now rare, all lower vertebrates in the assemblage are known from the Mojave River drainage.

The fossil avians clearly suggest the presence of a variety of mildly saline or freshwater lake and lake margin environments. Judging from the food preferences, food procurement methods and nesting habits of extant bird species (Cogswell and Christman, 1977; Garrett and Dunn, 1981), open water, sandy beach flats, and extensive reedy marshlands were the dominant lacustrine habitats in Lake Manix (Jefferson, 1985b). The seasonally extralocal pattern of the migratory forms in the assemblage suggests an overall cooler or more equable climate.

Regional terrestrial vegetation patterns, reconstructed in part from packrat midden data (Spaulding et al., 1984; Spaulding, 1990), permit inferences about the local paleoflora. The alluvial slopes and low hills surrounding Lake Manix probably supported a juniper-sage brushland, and the nearby mountains most likely were covered with a pinyon-juniper woodland (Spaulding, 1980, 1990; Jefferson, 1987, 1991a). Local valley bottoms probably supported patchy semidesert grasslands and desert scrub. These floristic associations are consistent with the inferred browsing habits of the majority of the fossil mammals (Table 4) (Jefferson, 1987, 1991a).

Depositional history

Upper Mojave River drainage

Before the uplift of the Transverse Ranges, internal drainage typified most basins on the southern and central part of the Mojave Desert block. The southwestern margin of the block apparently drained across the San Andreas fault zone, west to the Pacific Ocean (Meisling and Weldon, 1989). About 3–2 m.y. ago, as the Transverse Ranges adjacent to this margin of the block were elevated along the San Andreas fault, drainage direction shifted to the northeast.

In the headwaters area of the ancestral Mojave River, ~100 km to the southwest of Manix, the Victorville Fan complex was shed off the rising Transverse Ranges northeast into the Victorville basin (Weldon, 1985; Meisling and Weldon, 1989; Kenny and Weldon, 1999) (Fig. 1). Here, magnetostratigraphic data provide a date of 1.95 Ma for the base of the ancestral Mojave River deposits (Cox and Tinsley, 1999; Cox et al., 2003). Cox et al. (2003) suggest that the river advanced from the Victorville basin northward 50 km to Lake Harper basin sometime after 0.78 Ma and probably between 0.57 and 0.47 Ma. They (Cox and Tinsley, 1999; Cox et al., 2003) argue that the ancestral Mojave River continued to advance eastward, overflowing Lake Harper basin, and (based on Jefferson, 1985a) reached the Manix basin, ~50 km to the east of the Harper basin, no earlier than ca. 0.5 Ma ago (Fig. 1). This places the appearance of the ancestral

Mojave River in Manix basin during the deposition of the lower wedge of Member B.

Lake Manix basin

A thick section of late Miocene and Pliocene fanglomerates that presumably reflects regional extensional tectonics, documents the initial formation of Manix basin. The top of the fanglomerates may be laterally equivalent with the base of the Mojave River Formation, which has an age of ca. 2.5 Ma. From prior to 2.5 Ma until ca. 1 Ma, Manix basin was internally drained (Nagy and Murray, 1991). Subaerial oxidization of silts and clays, bedded gypsum, and limestones in the Mojave River Formation reflect the presence of ephemeral saline lakes and/or playas. Fluvial deposits in the top of the upward-coarsening Mojave River Formation, estimated to be ca. 1 Ma (Nagy and Murray, 1991), record a westward-flowing drainage system. Given the drainage history of Victorville and Lake Harper basins (Cox and Tinsley, 1999; Cox et al., 2003), it is unlikely that these early fluvial deposits represent flow from an ancestral Mojave River as Nagy and Murray (1991) have suggested.

No lacustrine deposits ranging in age between ca. 1 and 0.5 Ma have been identified in Manix basin. During this period, Hale (1985) suggested that a large lake, Lake Blackwelder, filled Death Valley, Soda Lake basin (Hooke, 1999) and Manix basin (Hale, 1985). It is then argued that this lake overflowed the southern end of the Troy Lake basin arm of Manix basin southeast into the Colorado River drainage via Bristol and Danby Valleys (Hale, 1985). The existence of this lake has not been confirmed stratigraphically or by the presence of high elevation shorelines in the Manix basin region (Rosen, 1989; Brown and Rosen, 1995; Enzel et al., 2003). However, westward paleocurrents in the fluvial deposits of the upper Mojave River Formation (Nagy and Murray, 1991) are consistent with the flow direction of Hale's proposed drainage system.

A second pulse of alluvial fan development, Member A of the Manix Formation, which was probably tectonically induced (Nagy and Murray, 1991), separates the upper fluvial deposits of the Mojave River Formation from the base of Member B. Fluvial and lacustrine sediments that comprise the lowermost deposits in the lower wedge of Member B, estimated to be ca. 0.5 Ma (Jefferson, 1985a, 1994, 1999), document the appearance of the Mojave River system in Manix basin.

Fluvial/lacustrine deposition in Lake Manix records at least four major transgressive/regressive events over the past 500 ka (Jefferson, 1985a, 1994, 1991b) (the last three events appear on the reference stratigraphic section, Figure 6). Major lacustrine phases (shaded bands Figure

6) presumably were associated with high lake levels. Based on stratigraphic sequence and placement relative to dated horizons, some of these transgressions can be correlated with specific marine oxygen isotope stages (Morley and Hays, 1981; Martinson et al., 1987) (Fig. 6). Oncoid stromatolites also typically occur at the base of the transgressive lacustrine deposits (Awramik et al., 2000).

Although not constrained by radiometric dates, the two transgressive lacustrine events present in the lower wedge of Member B may correlate with the beginning of marine oxygen isotope stage 14 (505 ka, Morley and Hays, 1981) and stage 12 (421 ka, Morley and Hays, 1981). The lacustrine transgression at the base of the lower part of Member C, ~2 m above the U/Th series age of 350+ ka, is tentatively correlated with the beginning of marine oxygen isotope stage 8 (279 ka, Morley and Hays, 1981) (Fig. 6). Member C encompasses marine oxygen isotope stages 8 through 4.

The major lacustrine transgression at the base of the upper part of Member C, 0.3 m below the 185 ± 15 ka tephra (Fig. 6), is essentially coincident with the beginning of marine oxygen isotope stage 6 (189.6 ka, Martinson et al., 1987). Accordingly, the fluvial upper wedge of Member B that separates the two lacustrine phases of Member C (Figs. 6, 7), was deposited during marine oxygen isotope stage 7 (244.2–189.6 ka, Martinson et al., 1987).

Marine oxygen isotope stage 5 (129.8–73.9 ka, Martinson et al., 1987) occurred during the deposition of upper Member C. It may be represented by the three, littoral arkosic beds that appear in the section 1–3 m below a U/Th series age of 68 ± 4 ka (Fig. 6). If so, fluvial deposition within the central part of the lake basin was far less extensive during marine oxygen isotope stage 5 than during marine oxygen isotope stage 7, upper Member B (Figs. 6, 7).

Most of the lacustrine sediments preserved in Afton basin were deposited during marine oxygen isotope stage 6 (Meek, 2000). These are overlain by an extensive wedge of fanglomerate, presumably deposited during marine oxygen isotope stage 5. A U/Th series age of 80 ka, from an oncoid carbonate preserved atop the fanglomerate, indicates that Afton basin was filled by Lake Manix during marine oxygen isotope stage 4 (Meek, 2000).

Late in the deposition of upper Member C, ancestral Mojave River deltaic and fluvial sediments, in part represented by Member D, prograded eastward across the middle of the basin. This fan delta eventually divided the two western arms of the lake into separate Coyote and Troy Lake basins (Hagar, 1966; Groat, 1967; Meek, 1994, 1999; Cox et al., 2003; Enzel et al., 2003) (Fig. 1). Although dates from the base of Member D near Manix, suggest that the Member C/D contact may approximate

the age of marine oxygen isotope stage 4/3 boundary (58.9 ka, Martinson et al., 1987) (Jefferson, 1985a), this contact is time-transgressive and older to the west. Cox et al. (2003) suggest that deposition of the fan delta decreased the evaporative surface and the volume of Lake Manix, increasing overflow to the east into Soda Lake basin.

During the deposition of Member D, between ca. 60 and 19 ka, lake levels fluctuated (Meek, 1990). Detailed studies by Budinger (1992) at the reference section (Fig. 6), record three 0.5 m-thick lacustrine deposits in the lower, upper middle, and upper parts of Member D. These represent interfingering of lacustrine and fluvial/deltaic deposits along the western margin of the lake.

Based on dates obtained from Afton basin (Table 1), Lake Manix was at a highstand between ca. 31 ka and 28 ka (Meek, 1990). Between 28 ka and 21 ka, Meek (1990, 1999) has argued that Lake Manix was at a lowstand, and he further suggests that the ancestral Mojave River may have terminated in Lake Harper basin at this time. This assertion is based on a lack of dates within the 28–21 ka age range from Lake Manix, and the presence of lacustrine deposits in Harper basin of this age. However, evidence of a significant stratigraphic hiatus has not been identified within the fluvial/deltaic and lacustrine deposits in the upper half of Member D. Also, Enzel et al. (2003) suggest that there may have been sufficient Mojave River flow to sustain lakes in both Harper and Manix basins concurrently.

The youngest accepted age for Lake Manix deposits, 18.1 ka (Meek, 1999), was obtained in Afton basin. Ages of 17.9 ka and younger from Coyote Lake basin and 15.0 ka from Troy Lake basin (Table 1), post date deposition of the top of Member D south of Manix (Figs. 2, 6). By 19 ka, the Mojave River fan delta (Fig. 1) had prograded eastward dividing Coyote and Troy Lakes into separate basins, effectively restricting Lake Manix to Afton basin.

At this time, lake level was high. After 18 ka, probably during the late Wisconsin glacial maximum (17.8 ka, marine oxygen isotope event 2.2, Martinson et al., 1987), Lake Manix breached sill level at the east end of Afton basin (Ellsworth, 1933; Weldon, 1982; Meek, 2000; Cox et al., 2003). The formation of Afton Canyon, presently a >150 m-deep gorge leading east into the Pleistocene Lake Mojave basin (Figs. 1, 2), was initiated.

Meek (1989, 1990, 1999, 2000) has argued that this overflow resulted in catastrophic erosion of the upper part of Afton Canyon (Fig. 2) to the depth of the lake floor in Afton basin, >120 m. This assertion is based on analyses of erosion volumes in Afton basin and deeply buried Wisconsin surfaces in western Lake Mojave basin, a lack of recessional shorelines in Afton basin, a lack of fluvial

terraces within the upper walls of Afton Canyon, narrow deeply incised tributaries in Afton basin, and subsurface boulders overlying lacustrine clays downstream from the Canyon. Further incision of Afton Canyon apparently occurred much more slowly over the past <18 ka (Meek, 2000).

However, Wells and Enzel (1994) and Enzel et al. (2003) point out that an initial rapid incision is unlikely and suggest that the formation of Afton Canyon was time-transgressive. This argument is based on the slow westward, upstream migration of the Afton Canyon nick point as evidenced by the existence of lacustrine conditions on the Mojave River fan delta as late as 12–9 ka (Reynolds and Reynolds, 1985), extensive Holocene fluvial features within Afton Canyon (Wells and Enzel, 1994), and recessional shorelines in the Afton basin Enzel et al. (2003).

Clearly, the erosion of Afton Canyon and incision of the Mojave River through Manix basin has not been a simple process. Although the arguments of Meek (1989, 1999, 2000), Wells and Enzel (1994), and Enzel et al. (2003) appear well founded, an integrative hypothesis and resolution to this issue must await further field investigations.

Large meander channels, on the top of Member D and incised to varying depths through the Manix Formation east of Camp Cady, suggest that, at least in the central Manix basin, the Mojave River has gradually adjusted to Soda Lake basin base level over the past <18 ka. During latest Pleistocene through early Holocene time, after the disappearance of Lake Manix, the Mojave River continued to intermittently fill Coyote Lake basin (Hagar, 1966; Meek, 1999), Troy Lake basin (Groat, 1967; Meek, 1999) and possibly Afton basin (Ellsworth, 1933; Blackwelder and Ellsworth, 1936; Meek, 1999, 2000). Within historic time, the Mojave River has flowed periodically through Afton Canyon into Cronese Lakes and Soda Lake basin (Wells et al., 1989; Brown et al., 1990; Meek, 1999, 2000; Enzel et al., 2003; Wells et al., 2003) (Fig. 1).

Summary and conclusions

Middle to late Pleistocene deposits within the Manix Basin provide a relatively complete record of depositional environments for an important portion of the ancestral Mojave River. Fluvial and lacustrine deposits, which represent the first appearance of the Mojave River in the central Mojave Desert, are recognized near the base of the Manix Formation (Member B). These deposits confirm that the river terminated in the Victorville and/or Harper basins prior to 0.5 Ma (Fig. 1).

The Manix Formation is geologically mappable as four distinct, and largely laterally equivalent lithologic facies.

Sequential changes from primarily fluvial to primarily lacustrine deposition (like those represented by the interfingering of Member B and C, Figure 7), are largely climatically driven and not the consequence of motion on the Manix fault (McGill et al., 1988). Major transgressive lacustrine events and high lake levels were the result of increased flow along the ancestral Mojave River. These episodes occurred after marine oxygen isotope stage boundaries and are positively correlated with relatively cool/moist climatic conditions inferred to be present during even-numbered stages. Given the available dates and stratigraphic placement, such events occurred during marine oxygen isotope stages 14, 12, 8, 6, and 4 (Fig. 6).

Late in lake history, a fan delta (in part Member D), deposited by the ancestral Mojave River, spread across the west-central margin of the basin, separating Afton, Coyote Lake and Troy Lake basins (Fig. 1). Pulses of deposition along the prograding delta front were not necessarily climate driven or the result of changes in the upper drainage system of the ancestral Mojave River. About 18 ka, Lake Manix breached the east end of Afton basin, forming Afton Canyon. Erosion of the upper ~120 m of the Canyon may have been rapid, however, the lower part of Afton Canyon was cut more slowly. This has resulted in incision of the Mojave River along its present course, and exposure of the Manix Formation in eroded badlands that extend from Camp Cady to Afton Canyon (Fig. 2).

Fossils recovered from exposures of the fluvial and lacustrine sediments of the Manix Formation (Table 2) are now known to encompass late Irvingtonian through late Rancholabrean LMA time. The relative abundance and paleoecological character of taxa permit a limited but significant reconstruction of paleohabitats. A substantial portion of the extinct and extralocal fossil vertebrates were ecologically tied directly to the lacustrine and paralic environments of Lake Manix. These include essentially all lower vertebrate and avian species. Browsers or browser-grazers (75% NISP, Table 4) that take advantage of seasonally available forage were the dominant large mammalian herbivores at Lake Manix and in the central Mojave Desert region during middle to late Pleistocene time.

Acknowledgments

B.W. Cahill of California State Parks is kindly thanked for his assistance in preparation of the illustrations. The helpful reviews, comments and suggestions of Y. Enzel, D.P. Whistler, and M.O. Woodburne are greatly appreciated. The assistance and cooperation of the many individuals who have supported and participated in research in Manix basin over the past several decades is sincerely acknowledged.

References cited

- Akersten, W.A., Foppe, T.M., and Jefferson, G.T., 1988, A new source of dietary data for extinct herbivores: *Quaternary Research*, v. 30, p. 92–97.
- Awramik, S.M., Buchheim, H.P., Leggett, L., and Woo, K.S., 2000, Oncoids of the late Pleistocene Manix Formation, Mojave Desert region, California, in Reynolds, R.E., and Reynolds, J., eds., Empty basins, vanished lakes: San Bernardino County Museum Association Quarterly, v. 47(2), p. 25–31.
- Bacon, C.R., and Duffield, W.A., 1981, Late Cenozoic rhyolites from the Kern Plateau, southern Sierra Nevada, California: *American Journal of Science*, v. 281, p. 1–34.
- Badgley, C., 1986, Counting individuals in mammalian fossil assemblages from fluvial environments: *Palaios*, v. 1(3), p. 328–338.
- Bassett, A.M., and Jefferson, G.T., 1971, Radiocarbon dates of Manix Lake, central Mojave Desert, California: *Geological Society of America Abstracts to Meetings*, v. 3, no. 2, p. 79.
- Bassett, A.M., and Kupfer, D.H., 1964, A geologic reconnaissance in the southeastern Mojave Desert: *California Division of Mines and Geology Special Report 83*, 43 p.
- Beyers, F.M., Jr., 1960, Geology of the Alvord Mountain quadrangle, San Bernardino County, California: *U.S. Geological Survey Bulletin 1089-A*, 71 p.
- Blackwelder, E., 1954, Pleistocene lakes and drainage in the Mojave region, southern California, in Jahns, R.H., ed., *Geology of southern California*: California Division of Mines and Geology, *Bulletin 170*, p. 35–40.
- Blackwelder, E., and Ellsworth, E.W., 1936, Pleistocene lakes of the Afton basin, California: *American Journal of Science 5th Series*, v. 31, p. 453–463.
- Blanc, R.P., and Cleveland, G.B., 1961a, Pleistocene lakes of southeastern California, Part I: *California Division of Mines and Geology, Mineral Information Service*, v. 14(4), p. 1–8.
- Blanc, R.P., and Cleveland, G.B., 1961b, Pleistocene lakes of southeastern California, Part II: *California Division of Mines and Geology, Mineral Information Service*, v. 14(5), p. 72–82.
- Brown, W.J., and Rosen, R.M., 1995, Was there a Pliocene-Pleistocene fluvial lacustrine connection between Death Valley and the Colorado River?: *Quaternary Research*, v. 43, p. 286–296.
- Brown, W.J., Wells, S.G., Enzel, Y., Anderson, R.Y., and McFadden, L.D., 1990, The late Quaternary history of pluvial Lake Mojave: Silver Lake and Soda Lake basins, California, in Reynolds, R.E., et al., eds., *At the end of the Mojave: Quaternary studies in the eastern Mojave Desert*: Redlands, California, Special Publication of the San Bernardino County Museum Association, 1990 Mojave Desert Quaternary Research Center Symposium, May 18–21, 1990, p. 55–72.
- Budinger, F.E., Jr., 1992, Targeting early man sites in the western United States: An assessment of the Manix type section, central Mojave Desert, California [M.A. thesis]: Special Major, San Bernardino, California State University, 229 p.
- Buwalda, J.P., 1914, Pleistocene beds at Manix in the eastern Mojave Desert region: *Bulletin Department of Geology, Berkeley, University of California*, v. 7(24), p. 443–464.
- Chidester, D.H., 1965, A short history of Camp Cady [Senior thesis]: Department of History, La Verne College, 28 p.
- Cogswell, H.L., and Christman, 1977, *Water birds of California*: Berkeley, University of California Press, *Natural History Guides*, no. 40, 399 p.
- Cox, B.F., Hillhouse, J.W., and Owen, L.A., 2003, Pliocene and Pleistocene evolution of the Mojave River, and associated tectonic development of the Transverse Ranges and Mojave Desert, based on borehole stratigraphy and mapping of landforms and sediments near Victorville, California. In *Paleoenvironments and Paleohydrology of the Mojave and Southern Great Basin Deserts*, edited by Y. Enzel, S.G. Wells, and N. Lancaster, *Geological Society of America Special Paper 368*:1–42.
- Cox, B.F., and Tinsley, J.C., III, 1999, Origin of the late Pliocene and Pleistocene Mojave River between Cajon Pass and Barstow, California, in Reynolds, R.E., and Reynolds, J., eds., *Tracks along the Mojave*: San Bernardino County Museum Association Quarterly, v. 46(3), p. 49–54.
- Davis, O.K., Agenbroad, L., Martin, P.S., and Mead, J.I., 1984, The Pleistocene dung blanket of Bechan Cave, Utah, in Genoways, H.H., and Dawson, M.R., eds., *Contributions in Quaternary vertebrate paleontology: A volume in memorial to John E. Guilday*: Carnegie Museum of Natural History Special Publication 8, p. 267–282.
- Downs, T., Howard, H., Clements, T., and Smith, G.I., 1959, Quaternary animals from Schuiling Cave in the Mojave Desert, California: *Los Angeles County Museum of Natural History, Contributions in Science*, no. 29, 21 p.
- Ellsworth, E.W., 1933, Physiographic history of the Afton Basin of the Mojave Desert: Des Moines, Iowa, *Pan-American Geologist*, May 1933, v. 59(4), p. 308–309.
- Enzel, Y., S.G. Wells, and N. Lancaster, 2003, Late Pleistocene lakes along the Mojave River, southern California. In *Paleoenvironments and Paleohydrology of the Mojave and Southern Great Basin Deserts*, edited by Y. Enzel, S.G. Wells, and N. Lancaster, *Geological Society of America Special Paper 368*:61–77.
- Evans, G.L., 1961, The Friesenhan Cave: *Texas Memorial Museum Bulletin*, v. 2, p. 3–22.
- Fergusson, G.J., and Libby, W.F., 1962, UCLA radiocarbon dates, I: *Radiocarbon*, v. 4, p. 109–114.
- Garrett, K., and Dunn, J., 1981, The birds of southern California status and distribution: *Los Angeles Audubon Society, Los Angeles, California*, 408 p.
- Gauthier-Pilters, H., and Dagg, A.I., 1981, The camel, its behavior, evolution, and relationship to man: Chicago, Illinois, University of Chicago Press, 208 p.
- Ginnett, T.F., 1982, Comparative feeding behavior of burros and desert bighorn sheep in Death Valley National Monument: *Desert Bighorn Council 1982 Transactions*, p. 81–86.
- Ginnett, T.F., and Douglas, C.L., 1982, Food habits of feral burros and desert bighorn sheep in Death Valley National Monument: *Desert Bighorn Council 1982 Transactions*, p. 81–86.
- Groat, C.G., 1967, Geology and hydrology of the Troy Playa area, San Bernardino County, California [M.S. thesis]: New York, University of Rochester, 133 p.
- Hagar, D.J., 1966, Geomorphology of Coyote Valley, San Bernardino County, California [Ph.D. thesis]: Department Geology, Amherst, University of Massachusetts, 210 p.
- Hale, G.R., 1985, Mid-Pleistocene overflow of Death Valley toward the Colorado River, in Hale, G.R., ed., *Quaternary lakes of the eastern Mojave Desert, California*: Friends of the Pleistocene Pacific Cell Annual Field Trip Guidebook, p. 36–81.
- Hansen, R.M., 1976, Foods of free roaming horses in southern New Mexico: *Journal of Range Management*, v. 29(4), p. 347.
- Hansen, R.M., 1978, Shasta ground sloth food habits, Rampart Cave, Arizona: *Paleobiology*, v. 4(3), p. 302–319.
- Hastorf, C.A., and Tinsley, J.C., 1981, Maps and index of radiocarbon-dated samples from southern California: *U.S. Geological Survey Miscellaneous Field Studies Map MF-1294*, scale 1:500,000.
- Hooke, R.L., 1999, Lake Manly(?) shorelines in the eastern Mojave Desert, California: *Quaternary Research*, v. 52, p. 328–336.

- Horton, D.R., 1984, Minimum numbers; a consideration: *Journal of Archaeological Science*, v. 11, p. 255–271.
- Howard, H.H., 1955, Fossil birds from Manix Lake, California: U.S. Geological Survey Professional Paper 264-J, p. 199–205.
- Hubbs, C.L., Bein, G.S., and Suess, H.E., 1962, La Jolla natural radiocarbon measurements, II: *Radiocarbon*, v. 4, p. 204–238.
- Hubbs, C.L., Bein, G.S., and Suess, H.E., 1965, La Jolla natural radiocarbon measurements, IV: *Radiocarbon*, v. 7, p. 66–117.
- Izett, G.A., 1981, Volcanic ash beds: Recorders of upper Cenozoic silic pyroclastic volcanism in the western United States: *Journal of Geophysical Research*, v. 68(B11), p. 10,200–10,222.
- Jefferson, G.T., 1968, The Camp Cady local fauna from Pleistocene Lake Manix, California [M.A. thesis]: Department of Geology, Riverside, University of California, 106 p.
- Jefferson, G.T., 1983, A fragment of human skull from Schuiling Cave, Mojave Desert, California: *Southern California Academy of Sciences Bulletin*, v. 82(2), p. 98–102.
- Jefferson, G.T., 1985a, Stratigraphy and geologic history of the Pleistocene Lake Manix Formation, central Mojave Desert, California, in Reynolds, R.E., ed., *Cajon Pass to Manix Lake: Geological investigations along Interstate 15: Redlands, California*, San Bernardino County Museum Association Special Publication, p. 157–169.
- Jefferson, G.T., 1985b, Review of the late Pleistocene avifauna from Lake Manix, central Mojave Desert, California: *Natural History Museum Los Angeles County Contribution in Science*, no. 362, 13 p.
- Jefferson, G.T., 1986, Fossil vertebrates from late Pleistocene sedimentary deposits in the San Bernardino and Little San Bernardino Mountains region, in Kooser, M., and Reynolds, R.E., eds., *Geology around the margins of the eastern San Bernardino Mountains: Redlands, California*, Publications of the Inland Geological Society, no. 1, p. 77–80.
- Jefferson, G.T., 1987, The Camp Cady local fauna: Paleoenvironment of the Lake Manix Basin: *San Bernardino County Museum Association Quarterly*, v. 34(3 and 4), p. 3–35.
- Jefferson, G.T., 1988, Late Pleistocene large mammalian herbivores: Implications for big game hunters in southern California: *Southern California Academy of Sciences Bulletin*, v. 87(3), p. 89–103.
- Jefferson, G.T., 1989, Late Pleistocene and earliest Holocene fossil localities and vertebrate taxa from the western Mojave Desert, in Reynolds, R.E., ed., *The west-central Mojave Desert: Quaternary studies between Kramer and Afton Canyon: Redlands, California*, Mojave Desert Quaternary Research Center, San Bernardino County Museum Association Special Publication, p. 27–40.
- Jefferson, G.T., 1990, Rancholabrean Age vertebrates from the eastern Mojave Desert, California, in Reynolds, R.E., et al., eds., *At the end of the Mojave: Quaternary studies in the eastern Mojave Desert: Redlands, California*, Special Publication of the San Bernardino County Museum Association, 1990 Mojave Desert Quaternary Research Center Symposium, May 18–21, 1990, p. 109–115.
- Jefferson, G.T., 1991a, The Camp Cady local fauna: Stratigraphy and paleontology of the Lake Manix basin, in Woodburne, M.O., et al., eds., *Inland southern California: The last 70 million years: San Bernardino County Museum Association Quarterly*, v. 38(3,4), p. 93–99.
- Jefferson, G.T., 1991b, Rancholabrean Age vertebrates from the southeastern Mojave Desert, California, in Reynolds, R.E., ed., *Crossing the borders: Quaternary studies in eastern California and southwestern Nevada: Redlands, California*, Mojave Desert Quaternary Research Center, San Bernardino County Museum Association Special Publication, p. 27–40.
- Jefferson, G.T., 1991c, A catalog of late Quaternary vertebrates from California: Part two, mammals: *Natural History Museum of Los Angeles County Technical Reports*, no. 7, 129 p.
- Jefferson, G.T., 1994, Stratigraphy and Pliocene-Pleistocene history of the Lake Manix basin, in McGill, S.F., and Ross, T.M., eds., *Geological investigations of an active margin: Geological Society of America Guidebook, 27th Annual Meeting, San Bernardino, California*, p. 175–177.
- Jefferson, G.T., 1999, Age and stratigraphy of Lake Manix basin, in Reynolds, R.E., and Reynolds, J., eds., *Tracks along the Mojave: San Bernardino County Museum Association Quarterly*, v. 46(3), p. 109–112.
- Jefferson, G.T., and Steinmetz, J.J., 1986, Ostracode biostratigraphy and paleoecology of the late Pleistocene Manix Formation: *Current Research in the Pleistocene*, v. 3, p. 55–56.
- Keaton, J.R., and Keaton, R.T., 1977, Manix fault zone, San Bernardino County, California: *California Division Mines and Geology, California Geology*, v. 30(8), p. 177–186.
- Kenny, M.D., and Weldon, R.J., 1999, Timing and magnitude of mid to late Quaternary uplift of the western San Bernardino and northeastern San Gabriel Mountains, southern California, in Reynolds, R.E., and Reynolds, J., eds., *Tracks along the Mojave: San Bernardino County Museum Association Quarterly*, v. 46(3), p. 33–46.
- Lovich, J., and Meyer, K., 2000, Aspects of the ecology of the Western Pond Turtle in the Mojave River, in Reynolds, R.E., and Reynolds, J., eds., *Empty basins, vanished lakes: San Bernardino County Museum Association Quarterly, Abstracts from the Year 2000 Desert Symposium*, v. 47(2), p. 79.
- Martin, P.S., Sables, E.B., and Shutler, D., Jr., 1961, Rampart Cave coprolite and ecology of the Shasta ground sloth: *Science*, v. 259, p. 102–127.
- Martinson, D.G., Pisias, N.P., Hays, J.D., Imbrie, J., Moore, T.C., Jr., and Shackleton, J.N., 1987, Age dating and the orbital theory of the ice ages: Development of a high-resolution 0–300,000-year chronology: *Quaternary Research*, v. 27, p. 1–29.
- McDonald, H.G., Jefferson, G.T., and Force, C., 1996, Pleistocene distribution of the ground sloth *Nothrotheriops shastensis* (Xenarthra, Megalonychidae), in Reynolds, R.E., and Reynolds, J., eds., *The 1996 Desert Symposium, Abstracts of Papers Submitted to the Meetings, San Bernardino County Museum Association Quarterly*, v. 43(2), p. 151–152.
- McGill, S.F., Murray, B.C., Maher, K.A., Lieske, J.H., Jr., and Rowan, L.R., 1988, Quaternary history of the Manix fault, Lake Manix basin, Mojave Desert, California: *San Bernardino County Museum Association Quarterly*, v. 35(3 and 4), p. 3–20.
- Meagher, M.M., 1973, The bison of Yellowstone National Park: *National Park Survey Scientific Monograph Series 1*, 171 p.
- Mech, L., 1966, The wolves of Isle Royale: *Fauna of the National Parks of the United States: Washington, Fauna Series 7*, 210 p.
- Meek, N., 1989, Geomorphic and hydrologic implications of the rapid incision of Afton Canyon, Mojave Desert, California: *Geology*, v. 17, p. 7–10.
- Meek, N., 1990, Late Quaternary Geochronology and geomorphology of the Manix basin, San Bernardino County, California [Ph.D. thesis]: Department of Geography, Los Angeles, University of California, 212 p.
- Meek, N., 1994, The stratigraphy and geomorphology of Coyote basin, central Mojave Desert, California: *San Bernardino County Museum Quarterly*, v. 41(3), p. 5–13.
- Meek, N., 1999, New discoveries about the late Wisconsin history of the Mojave River system, in Reynolds, R.E., and Reynolds, J., eds., *Tracks along the Mojave: San Bernardino County Museum Association Quarterly*, v. 46(3), p. 113–118.

- Meek, N., 2000, The late Wisconsin history of the Afton Canyon area, in Reynolds, R.E., and Reynolds, J., eds., Empty basins, vanished lakes: San Bernardino County Museum Association Quarterly, v. 47(2), p. 33–34.
- Meek, N., and Battles, D.A., 1991, Displacement along the Manix fault: California Division Mines and Geology, California Geology, February, v. 44(2), p. 33–38.
- Meisling, K.E., and Weldon, R.J., 1989, Late Cenozoic tectonics of the northwestern San Bernardino Mountains, southern California: Geological Society of America Bulletin, v. 101, p. 106–128.
- Merriam, J.C., 1915, Extinct faunas of the Mojave Desert, their significance in a study of the origin and evolution of life in America: Popular Science Monthly, March 1915, p. 245–264.
- Miller, R.R., 1948, The cyprinodont fishes of the Death Valley system of eastern California and southwestern Nevada: Miscellaneous Publications of the Museum of Zoology, University of Michigan, no. 68, 155 p.
- Miller, R.R., 1973, Two new fishes, *Gila bicolor snyderi* and *Catostomus fumeiventris*, from the Owens River basin, California: Occasional Papers of the Museum of Zoology, University of Michigan, no. 667, 19 p.
- Morley, J.J., and Hayes, J.D., 1981, Toward a high-resolution, global deep-sea chronology for the last 750,000 yr: Earth and Planetary Science Letters, v. 53, p. 279–295.
- Murray, B.C., and Nagy, E.A., 1990, The relationship of the Manix Formation to the "Mojave River Formation": Mojave Desert Quaternary Research Center Symposium, San Bernardino County Museum Association Quarterly, v. 37(2), p. 32.
- Nagy, E.A., and Murray, B.C., 1991, Stratigraphy and intrabasin correlation of the Mojave River Formation, central Mojave Desert, California: San Bernardino County Museum Association Quarterly, v. 38(2), p. 5–30.
- Naples, V.L., 1987, Reconstruction of cranial morphology and analysis of function in the Pleistocene ground sloth, *Nothrotheriops shastense* (Mammalia, Megatheriidae): Natural History Museum of Los Angeles County Contributions in Science, no. 398, 21 p.
- Naples, V.L., 1989, Feeding mechanism in the Pleistocene ground sloth, *Glossotherium*: Natural History Museum of Los Angeles County Contributions in Science, no. 415, 23 p.
- Peden, D.S., 1976, Botanical composition of bison diets on shortgrass plains: American Midland Naturalist, v. 96(1), p. 225–229.
- Pluhar, C.J., Kirschvink, J.L., and Adams, R.W., 1991, Magnetostratigraphy and clockwise rotation of the Pliocene-Pleistocene Mojave River Formation, central Mojave Desert, California: San Bernardino County Museum Association Quarterly, v. 38(2), p. 31–42.
- Pyle, R.L., 1961, Annotated field list birds of southern California: Los Angeles Audubon Society, revised by A. Small, O. Wade Publisher, Los Angeles, California, 61 p.
- Reynolds, R.E., and Reynolds, R.L., 1985, Late Pleistocene faunas from Daggett and Yermo, San Bernardino County, California, in Reynolds, R.E., ed., Cajon Pass to Manix Lake: Geological investigations along Interstate 15: Redlands, California, San Bernardino County Museum Association Special Publication, p. 175–191.
- Roeder, M.A., 1985, Late Wisconsin records of *Gasterosteus aculeatus* (threespine stickle-back) and *Gila bicolor mojaviensis* (Mojave tui chub) from unnamed Mojave River sediments near Daggett, San Bernardino County, California, in Reynolds, R.E., ed., Cajon Pass to Manix Lake: Geological investigations along Interstate 15: Redlands, California, San Bernardino County Museum Special Publication, p. 171–174.
- Rosen, M.R., 1989, Sedimentological, geochemical and hydrological evolution of an intercontinental closed-basin playa (Bristol Dry Lake): A model for playa development and its implications for paleoclimate [Ph.D. thesis]: University of Texas.
- Scott, E., 1996, The small horse from Valley Wells, San Bernardino County, California, in Reynolds, R.E., and Reynolds, J., eds., Punctuated chaos in the northeastern Mojave Desert: San Bernardino County Museum Association Quarterly, v. 43(1, 2), p. 85–89.
- Scott, E., 1997, A review of *Equus conversidens* in southern California, with a report on a second, previously unrecognized species of Pleistocene small horse from the Mojave Desert: Journal of Vertebrate Paleontology, v. 17(sup. 3), p. 75A.
- Scott, E., 1998, *Equus scotti* from southern California: Journal of Vertebrate Paleontology, Abstracts of Papers, v. 18(sup. 3), p. 76A.
- Scott, E., 1999, The *Equus (Plesippus)*-*Equus scotti* transition in western North America: Journal of Vertebrate Paleontology, Abstracts of Papers, v. 19(sup. 3), p. 74A.
- Scott, E., 2000, Fossil horses at Fort Irwin, the paleontology of Bitter Springs Playa: Environmental Division of the Directorate of Public Works, Fort Irwin, California, Natural and Cultural Resources Series No. 2, 15 p.
- Sarna-Wojcicki, A.M., Bowman, H.R., Meyer, C.E., Russell, P.C., Asaro, F., Michael, H., Rowe, J.J., Jr., Baedeker, P.A., and McCoy, G., 1980, Chemical analyses, correlations, and ages of late Cenozoic tephra units of east-central and southern California: U.S. Geological Survey Open-File Report 80-231, 52 p.
- Sarna-Wojcicki, A.M., Bowman, H.R., Meyer, C.E., Russell, P.C., Woodard, M.J., McCoy, G., Rowe, J.J., Jr., Baedeker, P.A., Asaro, F., and Michael, H., 1984, Chemical analyses, correlations, and ages of upper Pliocene and Pleistocene ash layers of east-central and southern California: U.S. Geological Survey Professional Paper 1293, 40 p.
- Schaller, G.B., 1972, The Serengeti lion, a study of predator-prey relations: Wildlife Behavior Series, Chicago, Illinois, University of Chicago Press, 480 p.
- Seiple, E., 1994, Lake Manix: California Division Mines and Geology, California Geology, March/April, v. 47(2), p. 50–57.
- Snyder, C.T., Hardman, G., and Zdenek, F.Z., 1964, Pleistocene lakes in the Great Basin: U.S. Geological Survey Miscellaneous Geological Investigations Map I-416, scale 1:1,000,000.
- Spaulding, W.G., 1980, The presettlement vegetation of the California desert: Manuscript on file, U.S. Bureau of Land Management, Riverside, California, 97 p.
- Spaulding, W.G., 1990, Vegetational and climate development of the Mojave Desert: The last glacial maximum to the present, in Betancourt, J.L., et al., eds., Packrat middens: The last 40,000 years of biotic change: Tucson, Arizona, University of Arizona Press, p. 166–199.
- Spaulding, W.G., Leopold, E.B., and Van Devender, T.R., 1984, Late Wisconsin paleoecology of the American southwest, in Spaulding, C.P., ed., Late-Quaternary environments of the United States, Volume 1: The late Pleistocene: Minneapolis, University of Minnesota Press, p. 259–293.
- Stebbins, R.C., 1966, A field guide to western reptiles and amphibians: The Peterson Field Guide Series: Boston, Houghton Mifflin Company, 279 p.
- Steinmetz, J.J., 1987, Ostracodes from the late Pleistocene Manix Formation, San Bernardino County, California, in Reynolds, J., ed., Quaternary history of the Mojave Desert, San Bernardino County Museum Association Quarterly, v. 34(3 and 4), p. 46–47.
- Steinmetz, J.J., 1988, Biostratigraphy and paleoecology of limnic ostracodes from the late Pleistocene Manix Formation [master's thesis]: Department of Biological Sciences, Pomona, California State Polytechnic University, 64 p.

- Stock, C., 1925, Cenozoic gravigrade edentates of western North America with special reference to the Pleistocene Megalonychinae and Mylodontidae of Rancho La Brea: Washington, D.C., Carnegie Institute of Washington, Publication no. 331, p. 206 p.
- Stock, C., 1930, Rancho La Brea, a record of Pleistocene life in California: Los Angeles Museum Publication, no. 1, 82 p.
- Stuiver, M., 1969, Yale natural radiocarbon measurements, IX: Radiocarbon, v. 11 (2), p. 545–658.
- Taylor, D.W., 1967, Late Pleistocene molluscan shells from the Tule Springs area, in Wormington, H.M., and Ellis, D., Pleistocene studies in southern Nevada: Nevada State Museum Anthropological Papers, no. 13, p. 395–399.
- Walsh, S.L., 2000, Eubiostatigraphic units, quasibiostratigraphic units, and “assemblage zones”: Journal of Vertebrate Paleontology, v. 20(4), p. 761–775.
- Webb, S.D., 1965, The osteology of *Camelops*: Natural History Museum of Los Angeles County, Science Series, no. 1, 54 p.
- Weldon, R.J., II, 1982, Pleistocene drainage and displaced shorelines around Manix Lake, in Cooper, J.D., ed., Geologic excursions in the California desert: Geological Society of America, Cordilleran Section Guidebook, p. 77–81.
- Weldon, R.J., II, 1985, Implications of the age and distribution of late Cenozoic stratigraphy in Cajon Pass, southern California, in Reynolds, R.E., ed., Cajon Pass to Manix Lake: Geological investigations along Interstate 15: Redlands, California, San Bernardino County Museum Association Special Publication, p. 59–68.
- Wells, S.G., and Enzel, Y., 1994, Fluvial geomorphology of the Mojave River in the Afton Canyon area, eastern California—implications for the geomorphic evolution of Afton Canyon, in McGill, S.F., and Ross, T.M., eds., Geological investigations of an active margin: Geological Society of America, Cordilleran Section Guidebook, p. 177–182.
- Wells, S.G., Anderson, R.Y., McFadden, L.D., Brown, W.J., Enzel, Y., and Miossec, J., 1989, Late Quaternary paleohydrology of the eastern Mojave River drainage, southern California: Quantitative assessment of the late Quaternary hydrologic cycle in large arid watersheds: New Mexico Water Resources Research Institute, Technical Completion Report, Project 14–08–0001-G1312, no. 242, 253 p.
- Wells, S.G., W.J. Brown, Y. Enzel, R.Y. Anderson, and L.D. McFadden, 2003, Late Quaternary geology and paleohydrology of pluvial Lake Mojave, southern California. In Paleoenvironments and Paleohydrology of the Mojave and Southern Great Basin Deserts, edited by Y. Enzel, S.G. Wells, and N. Lancaster, Geological Society of America Special Paper 368:79–114.
- Winters, H.H., 1954, The Pleistocene fauna of the Manix Beds in the Mojave Desert, California [master’s thesis]: Department of Geology, California Institute of Technology, Pasadena, California, 58 p.

An interview with Elmer Ellsworth, the first researcher to study the Pleistocene deposits in Afton basin

Norman Meek

Department of Geography, California State University, San Bernardino, CA 92407, nmeek@csusb.edu

Abstract

This paper reports an interview with Elmer Ellsworth recorded in 1988, where he recalled his work conducted in the Afton basin of Lake Manix more than 55 years earlier. His memory of the events, along with some rare photos of Afton station, help elucidate how field research was conducted in the 1930s, and provides some interesting information about the early scientific publications in the central Mojave Desert.

Introduction

Elmer Ellsworth (see Fig. 1) was the first Pleistocene researcher to study the Afton sub-basin of the Lake Manix basin. As I was completing my own dissertation there in the late 1980's (Meek, 1990), I began to wonder what had happened to Elmer. Part of my interest was generated by several important discrepancies I found (Meek, 1988) between Elmer's dissertation (Ellsworth, 1932a), and what was reported in Blackwelder and Ellsworth (1936), which purported to be a summary of his dissertation.

I was able to locate Elmer with the help of the Stanford University alumni association, the institution that ap-

proved his doctoral work on the Afton basin. He invited me to his residence in San Francisco for the interview. I met Elmer and his wife on February 27, 1988, and conducted a recorded interview that lasted for more than two hours. He was 80 years old at the time and his memory would sometimes fade regarding events more than 55 years earlier. Nevertheless, he still remembered some information about the area and related numerous stories about what it was like to do academic geological work in the late 1920s and early 1930s, both in Wisconsin and the Mojave Desert.

My primary goal was to figure out why the discrepancies existed. I never intended to publish the interview,

but two decades later I have realized that this taped interview may be the last opportunity to shed some light on the first field research in Afton basin. Moreover, the pictures Elmer graciously provided show how Afton basin and Afton station looked in 1932 when he completed his fieldwork there (see Figs. 2, 3, 4 and 5).

I have omitted most portions of the interview where Elmer digressed into topics such as his career with the Association of American Geologists, his World War II service, and similar topics (for this information, see Ellsworth, 1999). I have not omitted any parts of the interview that focus on his geological background or his experiences in Afton basin. In some spots, I cannot figure out what he said, and because the intervening two decades prevent me from recalling specifics, I have simply written [inaudible] where this happens.



Figure 1. Photo of Elmer Ellsworth and the author. This photo was taken at Ellsworth's residence in San Francisco, CA, on February 27, 1988.

The Interview

Q: When and where were you born?

A: June 17, 1907. About the time of Halley's Comet. In Norfolk, Virginia.

Q: Would you describe your schooling prior to Stanford?

A: I was living in Madison, Wisconsin, on University Avenue. It was only two blocks to Science Hall where the geology department was, so I had no problem communicating.

Q: Did you go to Madison [the University of Wisconsin] for your Bachelor's degree?

A: I lived in Madison. Right across from the university, right there. I went on to get my Masters there with no problems, but they advised me--Twenhofel^[1] was the faculty man--he said "you shouldn't take all of your work here." Well that was a new idea to me. So he said, "Why don't you write to Harvard, Columbia and Stanford and apply for a fellowship?" And all three of them gave me fellowships.

Q: And this was what year?

A: 1930. So I came out here and started right in.^[2]

Q: You said you lived in Madison. Had your family moved there from Norfolk?

A: I moved there about 1912, I'd say.

Q: Was there any reason you moved to Madison?

A: Her mother [referring to his wife] lived there. She was a widow and had a three story rooming house. And so we just fit right in there.

Q: What did your parents do?

A: My father served in the army in the Philippines. He came back and was discharged right here at the presidio—at the medical building which is still here—with a medical discharge. He wrote about his experiences in the Philippines and the volume is in the Library of Congress. He was just an enlisted man, a sergeant. My mother was a registered nurse. She graduated from the Boston city hospital and was the head nurse at the Providence [inaudible] hospital.

Q: Was your family from New England?

A: No. The family was from Evansville, Indiana.

Q: When were you married?

A: We were married in '46, after the army and World War II. I served in the Natural Resources Section. That was in Tokyo. We started out in the Arctic, Desert, and Tropical Information Center. It was part of the army air corps. I was a special air force intelligence officer. I was

the chief of one of the sections of information collecting and records.

Q: How many children do you have?

A: Well, I think we can count about three—two boys and one daughter. My daughter is a nurse, and one of my sons is the Chief of Seismology at the USGS, Menlo Park. My other son has a Ph.D. from Duke. He is an oral historian at the Smithsonian. He works for the National Museum of American History in the archives. His name is Scott, and my other son is William.

Q: Did you study geology for your Bachelor's and Master's at the University of Wisconsin?

A: I wanted to be a geologist because a friend of mine sat next to me in high school. His older brother was a geologist. He was a smart geologist and I wanted to be a geologist too. But I knew that I was weak in geography. So I wanted to start geology in my freshman year. I postponed a year so that I could take a course that I needed to fill a vacancy I had in my working knowledge. So I took geology.

Q: Do you recall any of the names of the professors you had at Wisconsin?

A: C. K. Leith,^[3] head of the department. I worked for him one summer after my Master's degree up in Crystal Falls, Michigan for McCann's Math[?] Iron Mining Company. Also Dr. William H. Twenhofel. He was a sedimentologist. And he got me interested in... well, my theses were written under him. The first one was on the varved clay deposits of Wisconsin. The annual rates, just like tree rings. And I sent my graph material to Baron Gerard de Geer^[4] in Stockholm and he worked it into the dated, [the] year-dated system that he had developed.

Q: And is that what you worked on for your thesis then?

A: That was my undergraduate thesis and also my Master's thesis. For my Master's I concentrated on a... sedimentology. I guess you would say... classification of sediments.

Q: Did you publish anything from that back then?

A: Let's see. I'm pretty sure. Yes, it was published in the Wisconsin Academy of Sciences,^[5] I believe.

Q: Now, let's start in with the Manix basin study. Why did you choose Stanford University?

A: Well, I had never been to California. That was a time when Los Angeles was a small area and they had a P.R. department of Pete Morrill and they had these ads all about the beaches, you know, and boy I wanted to go to the sunshine and beach. I came out here late in

the fall and it wasn't very warm and there weren't many people lying around on the beach.

Q: And how old were you at this point?

A: Well I was born in 1907, and went out there in 1930, in a Model T Ford.

Q: How were you introduced to the Manix basin?

A: Dr. Blackwelder^[6] was a student of the basins. Professor Buwalda,^[7] at CalTech, had written the first paper, and I think he kind of wanted someone to go down there and check it out.

Q: And were you with Blackwelder when you first visited the basin?

A: Oh no. He didn't come down there with a graduate classmate of mine, to check out, to see what we were doing. He was (inaudible) of the detail of where we actually worked or what we were doing. I presented a paper at the GSA meeting at the Pacific Coast Section in Los Angeles.^[8] He wanted me to publish it, but I didn't have any time to go into the detail on that sort of thing. When I got done with school I got a job back in Florida and never got back to it, so he published it.

Q: So your first sight of the Manix basin was on your own? He just said, "Here it is. Go out and find it and study it"?

A: Well, he referred me to the paper by Buwalda, so I had something to start with at that point. So Addimas [?] and me went down there and made as thorough of an examination as we could. We could read it. We didn't have any equipment to drill boreholes or anything like that. It was all a surface examination.

Q: Were your trips there by train or Model T?

A: Model T Ford. My friend Harold Kirchen^[9] and I shared the driving. We drove at night. It was hot. And even at night we had to close the windows to this car because it was still too hot in the desert. It is hard to believe.



Figure 2. Original caption: "Taken during Feb. trip to the Afton basin, in the Mohave Desert. Typical desert landscape. Almost no vegetation at all. We are here [Elmer on left, Harold W. Kirchen on right] seen loaded with shovels, picks, etc., on way to exposure of sediments which I measured and sampled."

Q: Can you name the towns that were on the route that you took back then?

A: We went through Los Angeles to San Bernardino.

Q: OK, so you went to Los Angeles from San Francisco first? I was wondering if you came out through Bakersfield, and then over Tehachapi Pass.

A: We had a rooming house where we lived at 1601 Wilshire Boulevard [Los Angeles]. And from there we would drive. To get this Kirchen thing in perspective, I was the representative of the manufacturer of a new magnetometer called the Hotchkiss, named after the chief geologist of Wisconsin—the state geologist.^[10] The Hotchkiss Superdip Magnetometer. I worked on the summer survey—the state of Wisconsin survey, and we used magnetometers for that survey. We used sundial compasses and handheld type of equipment. And I, in fact, was introducing it commercially out here. Harold Kirchen, as we were both classmates and stuck together, we would go around to the headquarters of Union Oil Company, and another oil company, and for a demonstration run a cross-section across their lot and show what indications the magnetic profile would give. And compare it with the standard German-made Askania Balance Magnetometer. And we ran a profile across [inaudible]. And he would just follow with us just to make sure we weren't monkeying around. We got the same picture, basically. And we did some work—not too much, but enough to earn some bread.



Figure 3. Original caption: "My first desert home; Afton, California. An isolated outpost. You are looking across the Mohave River, which is almost bone dry. Some of the sediments which I am studying are exposed on the far side of the canyon. Harold standing beside the building. Taken on Feb. 19 [probably 1932]."

Q: *I was wondering how you could afford a car as a graduate student.*

A: It carried \$100 a day, so just a couple of days would last us for quite a while.

Q: *And that is why you were in L.A., then, and not Stanford?*

A: It had nothing to do with the university. But the thesis actually was written in Los Angeles, and I remember on December 31st of 1932 we went to the post office together and mailed it.

Q: *And so is that why some people write 1932 on your thesis, and some people write 1933? I think it was actually filed formally in 1933.*

A: Well, the university, of course, didn't get the copies until 1933.

Q: *You wrote 1932 on it, and so I have always cited it as 1932. When you were studying the Manix basin, did you do reconnaissance work anywhere else, or did you confine your study pretty much just to the Afton area?*

A: The Afton area. In other words, there was a bar marking the shoreline north of Afton and that was as far as—because we knew it was all part of the Great Basin system, and that was our basin.

Q: *My question, more directly, is did you ever travel into Coyote basin, which was part of the lake, or Troy basin, which was part of the lake?*

A: Well. I don't... Can I get a map of it? I'm lost for a moment.

Q: *[Pointing to a map.] Afton is here and you worked at the town and canyon.*

A: No, we didn't get there.

Q: *You never went into Coyote or Troy?*

A: We were doing detailed work. Sure, this was not a reconnaissance survey.

Q: *[Examining the maps together.] I was just wondering if you confined your study mainly to the Afton area—which is what your thesis suggests, but at the same time you could have gone to the other areas and just not written about them.*

A: No, this wasn't the main part.

Q: *Would you describe the surveying methods and techniques that you used?*

A: We used a plane table. An alidade. A tripod. I've been doing that kind of work all along, so there wasn't anything unique about it.



Figure 4. Original caption: "Here is the house in which Pete Pederson and I are batching it. The Mexicans seen in front are living in the other half of the building. This is just across the track from the Afton 'station.' Pete works on the [Union Pacific] section crew (four men), while I am away at work. He and I get along O.K., and he is a pretty fair cook. I am a wow. Note the especially imported trees."

Q: *In a lot of those areas when you were doing profiles you had to go up some pretty steep grades with your survey lines. Do you recall how you were doing that?*

A: [He was confused by this line of questioning.] We weren't doing any reconnaissance.



Figure 5. Original caption: "This is the interior of Pete's and my present desert home. Looks better inside than out, don't you think so? A two room 'suite.' Good beds, mine is the far one. My 'office and laboratory' table, with accompanying soap-box chair, is seen next to the wall on the right. I have everything in the line of excellent field and laboratory equipment [sic], from microscopes to telescopic alidades. Not a bad camp headquarters at all."

Q: *Do any of the field notes, or surveying notes or data you collected back then still exist?*

A: I don't have any idea where they were. It doesn't register. I don't remember that. [After some questions to try and help with his memory.] We had the alidade and the rod, so we must have done something with it.

Q: *Do any other items or pictures exist from that study?*

A: I don't know. One reason is that my work—I didn't go in this direction. When I was through there, that was the end, you might say. Because I had to start making a living. This is fine. This is science and it is for a great cause. But a young one has to make a living, and so that was basically

it. This thing I presented at Los Angeles, I just took a great big sheet of grocery paper or something like that, and stretched it across the front of the place there at an auditorium where the GSA was meeting and just showed that type of—so they could get a rough idea.

Q: *So you said a Mr. Kirchen helped you?*

A: Yes. He was just doing that as a friend. Because we had this common interest in the magnetometer, and of course, whenever we got any money from that, which wasn't very often, why that helped pay our rent and food.

Q: *What was his full name?*

A: Harold W. Kirchen.

Q: *Did you know the big beach ridge north of Afton was there prior to your study?*

A: Oh no. Well, as far as I was concerned. It was just fantastic. It was like yesterday the water had just left. And if you see them anywhere else—the rains come and there is grass, weeds and trees—there was nothing there.

Q: *But when you went out there you didn't know it existed?*

A: Of course not.

Q: *I know Buwalda never mentioned it, but at the same point somebody could have discovered it in the intervening time like Dr. Blackwelder.*

A: No one was interested in the place—that we knew about. I never pumped anybody that knew anything about it.

Q: *So you were the first to discover it?*

A: It was on the aerial photographs—Fairchild, beautiful photos. It was all known to somebody. But they were not people that were in our profession

Q: *And where did you pick up the Fairchild photographs?*

A: Los Angeles. Here, the Santa Fe railway^[11] went out of the way to give me a place to live and everything, so Fairchild, I figure, did their part too, I'm sure.

Q: *What impact did any previous work, like Thompson's work that described the basin real generally, and Buwalda's work—did that have much influence on you when you were studying?*

A: Well, Buwalda was the thing, because as I told you, Blackwelder wanted me to double-check Buwalda. Use that kind of soft—don't make it a harsh thing. He just wanted someone to go down there, and you know.

Q: *Was there any animosity between the two?*

A: I'm not aware of anything like that, but you know all scientists can be a little touchy when another fellow comes into your territory, and starts to work on the thing you are working on. He was glad to get somebody he could send down there to check it out.

Q: *Recalling that your field work was done in the early summer months, what do you recall about your daily field schedule? Did you work during the heat of the day, or did you try to confine your work to the mornings and evenings?*

A: No, we just worked right on. One day it was so hot, and the night was hot, I just took my cot that I normally slept on at the place and took it outside there, just outside there. And Pete stayed there. After the oil[?] when I woke up the next morning, blood was running all of the way down one of these toes. Some insect had bitten it during the night and the blood had run down and dried. But it would keep you from sleeping. It was strange to look in the morning and to discover that something had been chewing at you, but you had to sleep there.

Q: *You said Pete. Who was Pete?*

A: He was the man that worked for the railroad, the section crew, and it was his apartment. And he was very kind to allow me, and when Harold was with me, we both stayed; although Harold wouldn't stay there too long because he was working long hours in Los Angeles.

Q: *Do you recall Pete's last name?*

A: Pederson. It might have been "en."

Q: *So you worked right through the day, then?*

A: Oh yeah. We did the same for the Wisconsin survey. It didn't make any difference about the weather or rain. [He then digressed to talk about the conditions in Wisconsin for a few minutes].

Q: *Do you recall any other unusual or exciting events in the Afton area when you were working?*

A: I was impressed by the fact that the sidewinders slept on the rails because they were so hot at night. There were sidewinders there, and they just stretched right out on the rails.

Q: *Did you find very many of them?*

A: Just one on a rail.

Q: *Were there a lot of snakes back then?*

A: They were quicker to get away from me. I don't remember. We weren't taking tallies of snakes [A digression follows.]

Q: *Did you spend much time in Palo Alto when you were in the program at Stanford?*

A: I lived in a fraternity house here, Alpha Kappa Lambda. In fact, I waited on tables there.

Q: *Do you recall any classes that you took?*

A: Well, I saw the need to take some more mineralogy. I had some unfinished work I wanted to do on the mineral contents of the summer layers of the varved clays that I had written about for my Bachelor's and Master's. I still have the slides here, but I couldn't do anything with them because I had no way of making a detailed analysis. [Then there is a digression about the foreign language requirement and how hard it was to pass.]

Q: *Did you have any classes with Blackwelder?*

A: Yes. He taught sedimentation, I believe. His classes were pretty broad and general. I have to have someone tell me what he was supposed to be teaching. I can't recall what it was, really. He caught on to me because I worked on the varved clay deposits when the GSA met in Madison.^[12] We had new dormitories, just built for

the season, and he had housing there. I carried his bags for him, and mentioned that I had just got through studying his paper on the glacial geology of the Big-horns. Well that made a very good contact as it turned out later. You see, he remembered that when I came out there to start a thesis, you know. People are flattered when some guy reads your paper you've written years ago. So he was very kind, and always has been. [This was followed by an irrelevant question.]

Q: Now I'd like to talk about the 1936 paper: Blackwelder and Ellsworth, that was written from your thesis work.

A: He gave up on me because I just couldn't do any survey work in Florida [inaudible] work with the magnetometer. I had no opportunity to sit down and do this kind of thing. It was just way beyond anything I could possibly do.

Q: So you didn't help write it?

A: No. No way. In fact, he just gave up and wrote it himself. That's what it amounted to. And I appreciated him doing that. I just couldn't handle it.

Q: Did you review it before it went to press?

A: Oh no.

Q: Were there any noteworthy discussions or debates between you and Professor Blackwelder concerning the content?

A: No. He just gave up and kindly published the paper, and I appreciate it.

Q: And you didn't mind that he became lead author on it?

A: Of course not. He thought enough of it to publish it.

Q: Was there any response to the paper by any others that you were ever told about?

A: I don't know. When you get into the oil business these things fade out.

Q: Let's talk about Professor Blackwelder for a couple minutes.

A: He was a fine gentleman—very fine.

Q: Was he a good thesis advisor? Did he give you much guidance in the writing of it?

A: [A laugh] Well, let me give you an answer this way. One day he called me into his office and we had a discussion and he said, "What is your minor?" Minor, what's that? I don't have a minor. He hadn't told me about any minor. So there wasn't any minor. I mean, he was a fine gentleman—a fine gentleman—ethical and everything else. But he didn't have the fullest contact with the T-As [teaching assistants?] or that sort of

thing. And I was as far as anybody when he asked me, "What is your minor?" I said I hadn't heard anything about that, so [inaudible].

Q: Did he provide you much assistance in the way of financing, or lab or field help, or anything like that?

A: No. But of course I used the equipment at the university.

Q: How many times did he read your thesis? Did he read it and then give it back to you for comment, for corrections, or...

A: Oh no. He didn't even see it because I was down in Los Angeles and published it down there.

Q: But it didn't go through a review through him or anything like that?

A: Oh no.

Q: So you wrote it completely on your own and then just gave it to him for approval?

A: Well, sure. I didn't give it to him. I sent it to where you send them for the university. I remember writing it, but [can't] find out for sure.

Q: Were there many other graduate students working with Professor Blackwelder while you were at Stanford?

A: Well. Thompson. I'm not sure whether he's the same one there or not.

Q: No, it couldn't be. So there was just maybe only one other student working with Professor Blackwelder at the time?

A: Well, I'm not even sure of that. Our paths didn't cross. I was busy. I waited on tables so I didn't have a lot of time foolin' around. And I spent so much time getting ready for these language exams—that was a bugaboo. I was taking all of that French [a lengthy digression].

Q: Now, in your thesis you report that Dr. Blackwelder visited you at least once in the field.

A: Yes, with my friend Thompson. We both were examined for our doctorate at the same time, so we were rather close in that regard. He brought him down there. Let's see. I think it was Blackwelder's car. I remember Blackwelder saw the condition of my tires, retreads, and he said, "Well, let me loan you money so you can get a couple of new tires here." Well, I had never had any money to pay him back. So we just got along the best we could.

Q: But he came there with one other student?

A: With Mr. Thompson, the same level as myself.

Q: Do you recall Mr. Thompson's first name?

A: Didn't I say A.R. Thompson?^[13] He was from Colorado. 95 Gilbert Street, but I can't remember the town.

Q: So, in total, how much time did he spend in the Manix basin with you?

A: None. He just came down on one little trip—just an afternoon or whatever it was—or part of a day. He wanted to see the evidence of the shorelines, you know, and he was satisfied with [inaudible].

Q: Did he lead any field trips to the basin while you were a student?

A: Oh hell no. Well there weren't that many students. One man that was real interested was the new state geologist.^[14] His name slips my mind. And about the time of the GSA annual meeting in Los Angeles before I gave this paper he said, "You're working down there in that desert area, Afton basin, how about my driving you down there and we go to the GSA meeting and you take me out to the place," and I'd show him. And then we went up to Death Valley, which is an exciting place to go with him. So, I said, "Well I [inaudible] taking anybody down there to see it." [This is followed by a lengthy digression about Death Valley Scotty and related topics.]

Q: Did you have much contact with Professor Blackwelder in the years between your graduation and your filing of your thesis and when he wrote that paper?

A: No. [Another lengthy digression.]

Q: Have you visited Afton basin since the 1930's?

A: No. [The lengthy interview continues into his work in Florida during the 1930s, his World War II service, and his post-war activities.]

Commentary

Looking back at this interview twenty two years later, a few points are noteworthy. First, Elmer did not have any background on desert processes, but he was an expert on the ice ages and lake sedimentation in Wisconsin when he arrived in California. This might explain why he repeatedly described features such as the well-cemented pedogenic carbonate horizons as mudflows (e.g., Ellsworth, 1932a, p. 14-16; 34), although this might have been partly influenced by his advisor (Blackwelder, 1928). Moreover, Elmer Ellsworth completed his doctoral studies from 1930 to 1932, apparently without much help from the Stanford faculty or other desert experts. In his dissertation, (Ellsworth, 1932a, p. 4) reports completing his fieldwork during April, May and June, 1931, with

follow-up visits in November, 1931, and March, 1932. He also reports two visits by Eliot Blackwelder (p. 4), but he could only recall one during the interview. To the best of his knowledge, Elmer was the first scientist to discover and realize the significance of the Afton basin beach ridges, since they had not been described previously, and he was unaware of their existence when he began his studies. Finally, Elmer completed his work without ever visiting Coyote basin, or studying other parts of the Manix basin, and he took no part in the preparation of the Blackwelder and Ellsworth (1936) manuscript, which explains why some important discrepancies exist between his dissertation and the paper.

In this interview Ellsworth reported that Eliot Blackwelder, along with another graduate student [probably W.O. Thompson], visited him at Afton on an afternoon, presumably in March, 1932, as part of his graduation exam process. Because two visits by Blackwelder have been reported (Ellsworth, 1932a, p. 4), and because most of the photos in Ellsworth's dissertation are attributed to Blackwelder (which probably would have taken more than an afternoon visit), Blackwelder must have made an earlier, somewhat lengthier visit. Because both of these visits preceded the July 11, 1933, submission of Blackwelder's manuscript "Origin of the Colorado River" (Blackwelder, 1934), where Blackwelder invoked a trunk prolongation model for the Colorado River, I have long believed that the Afton basin was the likely inspiration behind his model for the origin of the Colorado River landscape, just like it was for me 60 years later (Meek, 1992; Meek and Douglass, 2001). However, further research to confirm the validity of the statements in this interview has revealed that Blackwelder had already developed the trunk prolongation model for the Colorado River and presented it at a regional GSA meeting in Pasadena, CA during the first week of March, 1931 (Blackwelder, 1931), apparently before visiting the Afton basin with Ellsworth.

Dr. Elmer Ellsworth had a long life. After completing his work in Afton basin, he worked in the oil industry in the Midwest and southeastern U.S. before serving in a variety of military intelligence positions during World War II. After spending some additional time in postwar Japan, Elmer was the head convention organizer for the American Association of Petroleum Geologists for 25 years. He died at the age of 90 on September, 22, 1997 after a long struggle with Alzheimer's disease.

Acknowledgments

I would like to thank Joy Barta, George Jefferson, and David Miller for editing assistance; Patricia White of Stanford University for archival research assistance; and the hospitality and photos generously provided by Helen and Elmer Ellsworth.

References Cited

- Blackwelder, E., 1928. Mudflow as a geologic agent in semiarid mountains: *Geological Society of America Bulletin*, 39(2), p. 465-483.
- Blackwelder, E., 1931. Physiographic history of the Colorado River: *Geological Society of America Bulletin*, 43, p. 229.
- Blackwelder, E., 1934. Origin of the Colorado River: *Geological Society of America Bulletin*, 45, p. 551-566.
- Blackwelder, E., and Ellsworth, E.W., 1936. Pleistocene lakes of the Afton basin, California: *American Journal of Science*, 231, p. 453-463.
- Buwalda, J.P., 1914. Pleistocene beds at Manix in the eastern Mohave Desert region: *University of California Publications in Geology*, 7(24), p. 443-464.
- Ellsworth, E.W., and Wilgus, W.L., 1930. The varved clay deposit at Waupaca, Wisconsin: *Transactions of the Wisconsin Academy of Sciences, Arts and Letters*, 25, p. 99-111.
- Ellsworth, E.W., 1932a. Physiographic history of the Afton basin. Palo Alto, CA (Stanford): unpublished Ph.D. dissertation, 99 p.
- Ellsworth, E.W., 1932b. Varved clays of Wisconsin, *Transactions of the Wisconsin Academy of Sciences, Arts and Letters*, 27, p. 47-58.
- Ellsworth, E.W., 1933. Physiographic history of the Afton basin of the Mohave Desert: *Proceedings of the Geological Society of America* for 1933, p. 306-307.
- Ellsworth, S.A., 1999. Memorial to Elmer William Ellsworth, 1907-1997. *Geological Society of America Memorials*, v. 30.
- Meek, N., 1988. The Lake Manix shorelines: pseudohistory as a result of inadvertent errors, *Friends of Calico Newsletter*, 10(2); see also the follow-up article after significant elevation errors were later recognized: Meek, N., 1989, *Eating Crow*, *Friends of Calico Newsletter*, 11(1), p. 3-4.
- Meek, N., 1990. Late Quaternary geochronology and geomorphology of the Manix basin, San Bernardino County, California. Los Angeles, CA (UCLA): unpublished Ph.D. dissertation, 212 p.
- Meek, N., 1992. Episodic forward prolongation of trunk channels in the western United States: unpublished manuscript, 34 pp.
- Meek, N., and Douglass, J., 2001. Lake overflow: an alternative hypothesis for Grand Canyon incision and development of the Colorado River, p. 199-204 in Young, R.A., and Spamer, E.E., 2001, *Colorado River, Origin and Evolution*; Grand Canyon, AZ: Grand Canyon Association, 280 pp.
- Myrick, D.F., 1992. Railroads of Nevada and Eastern California, Volume II, the Southern Roads. Reno: University of Nevada Press, 933 pp.
4. Swedish geologist that pioneered the study of proglacial varves (1858-1943)
5. Ellsworth and Wilgus (1930); and Ellsworth (1932b)
6. Eliot Blackwelder (1880-1969), chairperson of the geology department at Stanford University, 1922 to 1945.
7. John P. Buwalda (1886-1954), founder of the geological division at the California Institute of Technology in 1925, and chairperson from 1926-1947. He wrote the first scientific paper about the Manix basin (Buwalda, 1914), before completing his dissertation at Berkeley in 1915.
8. Ellsworth (1933); Elmer was introduced by Blackwelder, and W.M. Davis was a discussant
9. Harold W. Kirchen (1907-2002). Harold was born in Montana, and educated at the Montana School of Mines and University of Wisconsin. He worked for the Corps of Engineers and Bureau of Reclamation in Denver, CO, before retiring in 1967. He died in Denver on December 4, 2002, age 95.
10. William O. Hotchkiss (1878-1954), state geologist of Wisconsin, 1909-1925.
11. In 1905, the San Pedro, Los Angeles, and Salt Lake Railroad, which was eventually controlled by the Union Pacific Railroad, built the tracks through Afton basin (Myrick, 1992). Today, the Salt Lake Route is a part of the Union Pacific Railroad, as it has always been.
12. 39th annual meeting of the Geological Society of America, Madison, WI, Dec 27-29, 1926.
13. "I did not find any A.R. Thompson in the alumni or student directories. I did find a Warren Osborne Thompson, Ph.D. in geology, 1935. The 1955 Alumni Directory lists his address as 985 Gilbert Street, Boulder, Colorado." Written communication with Patricia E. White, Archives Specialist, Department of Special Collections, Stanford University Libraries, 21 January 2010.
14. This was probably Walter W. Bradley (1878-1950), California State Mineralogist from 1928 to 1946.

Endnotes

1. W.H. Twenhofel (1875-1957), University of Wisconsin sedimentologist W.H. Twenhofel (1875-1957), University of Wisconsin sedimentologist
2. Ellsworth (1932b, p. 48) reports collecting sediments in Wisconsin as late as Spring, 1929.
3. Charles K. Leith (1875-1956), Department of Geology, University of Wisconsin (chairperson 1903-1934)

A brief comparison of central Mojave Desert lakes: Miocene Barstow Formation and Pleistocene Lake Manix

Robert E. Reynolds¹ and George T. Jefferson²

¹Redlands CA 92373, rreynolds220@verizon.net

²Colorado Desert District Stout Research Center, Borrego Springs, CA 92004

History and Setting

Definitions

The Barstow Formation consists of lacustrine and fluvial sedimentary facies deposited on alluvial fans in an actively forming Miocene basin. It is apparently the largest deposit of Miocene lacustrine sediments in the central Mojave Desert. The contained fossil faunas are the basis for the Barstovian North American Land Mammal Age (NALMA). Sediments lower in the section contain taxa attributed to the Hemingfordian NALMA. Early in Barstovian time, the Barstow Basin drained to the Pacific Ocean (Bell and Reynolds, 2010).

Lake Manix was one of the largest pluvial late Pleistocene lakes in the central Mojave Desert. Fed by the ancestral Mojave River (Reynolds and Cox, 1999; Cox et al., 2003) and largely represented by lacustrine sediments of member C of the Manix Formation (Jefferson, 1985a, 2003), Lake Manix deposits yield the Manix Local Fauna of invertebrate and latest Irvingtonian through Rancho-labrean NALMAs vertebrate fossils (Jefferson 1985b, 2003).

Age and duration

At its type section in the Mud Hills, the lacustrine portion of the sediments form the upper part of the Barstow Formation, and range in age from about 16 Ma to 13.3 Ma (Woodburne et al., 1990). On eastern Daggett Ridge, the Barstow Formation lies below the 18.7 Ma Peach Spring Tuff (Reynolds et al., this volume). Farther east in the Harvard Hill area, the lake beds are older than 19.2 Ma (Leslie et al., this volume). The duration of lake facies of the Barstow Formation starts earlier than 19.1 Ma in a partial section at Harvard Hill and extends to 13.3 Ma in the western Mud Hills, a span of 5.8 Ma.

Lake Manix is middle and late Pleistocene in age. Exposures of lacustrine and paralic deposits date from about 400 to 20 kyr BP based on U/Th, tephra correlation, and ¹⁴C (Jefferson, 2003), a time span of 380 ky. The Manix Formation is magnetically normal and contains a 185 ka-old ash (Jefferson, 2003). From ca. 2.5 to 1 Ma,

prior to the formation of Lake Manix, the basin held ephemeral playa lakes of the Mojave River Formation (Nagy and Murray, 1991).

Size and shape

The Barstow Formation appears to have been deposited in an elongate west-northwest to east-southeast basin. Outcrops are scattered over 1550 km² (600 square miles). Although outcrop distribution is, in part, a result of left-lateral and right-lateral strike slip faulting, this distribution does not represent the original, probably smaller, size of the basin.

Relatively continuous outcrops of Lake Manix deposits are found along the sides of the incised modern Mojave River channel and along local tributary drainages, such as Manix Wash. These exposures extend from near Camp Cady east to Afton Canyon, approximately 32 km (20 mi). The early Lake Manix probably lay west of Buwalda Ridge and occupied a broad basin west to Yermo, north to Coyote Lake, and southeast to Newberry Springs. Catastrophic integration by flood united the Afton arm with the main lake about 185 ka (Reheis et al., 2007). After that integration, Lake Manix had three arms or lobes, presently Coyote Lake playa basin to the north, Troy Lake playa basin to the south, and the Afton arm to the east. The west side of the lake, between the Coyote and Troy arms, was filled by the prograding fluvial delta of the ancestral Mojave River; drill records studied by D.M. Miller (pers. comm. 2010) show that lake sediments once extended west of the town of Daggett. At 543 m above sea level, the prominent high lake stand, Lake Manix episodically occupied about 236 km² (91 mi²) (Meek, 1990, 2004). The oldest and highest shoreline features are poorly preserved and sit at 547–549 m (Reheis and Redwine, 2008).

Drainage integration and source areas

Barstow Formation sediments contain interbedded fluvial gravels that provide paleogeographic information. In northern sections (e.g. Mud Hills, Calico Mountains,

Yermo Hills) gravels have a northern source, judging by distinctive clasts of metamorphic rock similar to modern exposures in the Goldstone area. The lowest units at Harvard Hill appear to be derived from the south (Leslie et al., this volume). Basin sediments in Stoddard Valley and on the north side of Daggett Ridge contain clasts shed from the rising of that ridge in the early Miocene. Stickleback fish found in the Barstow sediments at the Toomey Hills (Bell and Reynolds, this volume) suggest that the basin drained to the Pacific Ocean at 16 Ma.

Lake Manix was fed largely by the ancestral Mojave River which arrived in Manix basin ca. 500 kyr BP. The catchment of the Mojave River presently includes the Transverse Ranges, San Gabriel and San Bernardino Mountains, and portions of the western and central western Mojave Desert that immediately flank the river course. As the ancestral Mojave River progressed north and eastward over a period of about 1.0–1.5 my (Cox et al., 2003), once-isolated basins like the Victorville basin were integrated into a through-flowing drainage system

Depositional drivers

Miocene drainage systems from the central Mojave Block that reached the Pacific Ocean probably had a large catchment basin. A pre-Miocene erosional surface and pre-Barstow basin sediments were apparently disrupted by early Miocene extensional faulting, causing drainage system fragmentation and local ponding as early as early as 19.1 Ma (Fillmore and Walker, 1996; Leslie et al., this volume) for Barstow lake beds. Tectonic forces continued and caused the depocenter to migrate westward until it reached the western Mud Hills at 16 Ma.

The three, primarily lacustrine episodes of the Manix Formation (member C) appear to be climate driven and follow principal Milankovitch cycles. These lacustrine phases have been correlated with even-numbered Oxygen isotope stages (12, 8, and 6), and were deposited during pluvial climatic conditions (Jefferson, 2003). Newly recognized soils and shallow-water facies in these intervals are causing re-evaluations of the synchronicity of deep lakes and climate cycles (Reheis and Miller, 2010).

Lake disappearance

As the Barstow Basin depocenter migrated westward, lobes or sub-basins may not have been able to receive contributions of water due to structural dams. The Stoddard sub-basin may have been cut off from the main drainage system by the 19 Ma year uplift of Daggett Ridge, since few sediments are found there after the deposition of the 18.7 Ma PST. Harvard basin contains 19–18 Ma lacustrine sediments, but erosion has removed any record of further deposition. Deposition of sediments continued in

the portion of the Barstow Basin in the western Mud Hills from 16 Ma to less than 13 Ma.

After about 25 kyr ago, Lake Manix apparently catastrophically overflowed at a low point in the sill near Afton Canyon (Wells and Enzel, 1994; Enzel et al., 2003; Wells et al., 2003; Meek, 1989, 2000, 2004, Reheis and Redwine, 2008). Flowing east into Soda Lake basin, the upper part of Afton Canyon was relatively rapidly cut. Subsequent fluvial erosion has incised the present Mojave River channel into the former floor of the lake in the Afton arm and Manix Wash areas, and into basin bedrock in Afton Canyon.

Geologic evidence

Sections and sediments

The Barstow Formation is a 1 km-thick sequence of fluvial and lacustrine sediments that has been divided into three members or five faunal divisions (Woodburne et al., 1990). At the type section alluvial fanglomerates of crystalline rocks and Pickhandle volcanics are overlain by members that are lacustrine and fluvial. Each member consists of facies that change laterally, as do the included marker beds and air-fall tuffs. Recently, a sequence of marker units (Reynolds and Woodburne, 2002) has been recognized that unites the Barstow Formation in the separate ranges and hills. The marker sequence is time transgressive and ranges from approximately 19.5 Ma in southern lobes of the basin to 16.5 Ma in the northwestern basin in the Mud Hills.

The 40+ m-thick exposed section of the Manix Formation has been divided into four members. These represent basin flanking alluvial fans (member A), alluvial apron and fluvial sediments (member B), lacustrine and paralic deposits (member C), and ancestral Mojave River deltaic deposits (member D) (Jefferson, 2003). The Manix Formation locally sits unconformably on the Barstow Formation, and eroded Miocene deposits have contributed sediment to Lake Manix. The Manix Formation also sits disconformably or unconformably on the Mojave River Formation.

Structural settings

Excellent exposures of the Barstow Formation occur in mountain ranges (Mud Hills, Calico Mountains, Daggett Ridge) and isolated hills (Agate, Lime, Harvard). Uplift of these rocks apparently was initially caused by post-extensional isostatic rebound, and more recently by interactions of forces at the junctions of left-lateral and right-lateral strike slip faults (Singleton and Gans, 2008). Transpressive bends in the northwest-trending right-lateral strike slip faults may have also been contributed to the uplift.

The Lake Manix basin is largely intact, and the Pleistocene sediments are relatively undeformed and essentially flat lying. Along the Manix Fault, both underlying Mojave River Formation playa sediments and members of the Manix Formation have been gently folded and sharply upturned next to the Manix Fault east of Manix Wash (Nagy and Murray, 1991). Farther west, sand beds and intercalated beach gravels of Lake Manix are folded and thrust faulted in hills adjacent to the Manix Fault (Harvard Hill and Toomey Hills; Miller et al., 2009).

Paleo-lake conditions and aquatic organisms

Miocene lacustrine sediments contain 3–10 m-thick deposits of silicified limestone and lake-margin transgressive deposits of tufa over plants. These different types of carbonate deposits reflect, respectively, high bicarbonate content of shallow lake waters near the center of the basin and low bicarbonate content in shallow lake waters expanding over fanglomerates at the lake margins (Reynolds et al., this volume). Stickleback fish (Bell and Reynolds, this volume), conispiral and planispiral gastropods, and ostracodes all suggest shallow, warm, well oxygenated lake water. Hot spring exhalites (Schuiling, 1999) may have added minerals to the water over a period of 2.5 my, but water-dwelling insects preserved by silica and sulfates show that percentages of dissolved minerals were tolerable by complex life-forms (Park, 1995; Park and Downing, 2001; Spencer, 2005).

As Lake Manix periodically grew during pluvial periods, paralimnic and lacustrine sediments prograded over the basin margin alluvial apron. Layers of oncoid tufa at these horizons suggest that paleo-lake conditions were relatively warm, fresh and well oxygenated. Ostracodes recovered from the lacustrine sediments (member C) reflect cooler, meso-trophic, fresh, well oxygenated water (Steinmetz, 1987). A variety of fresh water mollusca, bivalves and snails, from the paralimnic, lacustrine and deltaic deposits indicate perennial fresh water habitats including streams, rivers, ponds, lakes and bogs or swamps. Fish (*Gasterosteus aculeatus*, *Gila bicolor mojavensis*) and pond turtle (*Actinemys [Clemmys] marmorata*) represented by fossil materials presently occupy lower elevation lakes and rivers in California (Jefferson, 1985a, 2003).

Summary

Deposition in the Miocene Barstow Basin was initiated by central Mojave Desert tectonic basin formation that dammed an extensive drainage system that once reached the Pacific Ocean. Deposition in a variety of freshwater lake environments within the basin continued for 6 my

during active uplift of the basin margins and as fanglomerates prograded from north and south highlands. Water chemistry differed between basin center and its margins producing contrasting forms of carbonate deposition. Tectonic disruption of the Miocene drainage apparently finally reduced the volume of water entering a once long term basin. Subsequent tectonic activity over the last 12 my left dismembered basin sections in isolated ranges that crop out over 1550 km².

Late Pliocene tectonism (Miller et al., 2009) developed a basin in the central Mojave Desert that was subsequently filled by saline playa sediments of the Mojave River Formation, followed by late Pleistocene lacustrine, basin margin alluvial and fluvial, and fluvial deltaic sediments of the Manix Formation. The Manix Basin was initially a single catchment downstream from the glaciated Transverse Ranges. As the Mojave River became an integrated drainage system, the single basin was divided into three lobes by ancestral Mojave River delta deposits. These reduced the basin capacity at the point of maximum late Wisconsinan pluviation and maximum runoff from sources in the Transverse Ranges, causing breaching eastward through Afton Canyon. Latest Pleistocene and Holocene fault activity have deformed the 20+ ka Lake Manix sediments at several locations.

Acknowledgements

The authors thank Dr. David Miller for thorough review and knowledgeable comments that helped to make this a better manuscript.

References

- Bell, M. A. and R. E. Reynolds. 2010. Miocene and Late Pleistocene stickleback spines from the Mojave Desert, California, in *Overboard in the Mojave: 20 million years of lakes and wetlands*, R.E. Reynolds and D. M. Miller (eds). Desert Studies, California State University Fullerton, this volume.
- Cox, B.E., J.W. Hillhouse, and L.A. Owen. 2003. Pliocene and Pleistocene evolution of the Mojave River, and associated tectonic development of the Transverse Ranges and Mojave Desert, based on borehole stratigraphy and mapping of landforms and sediments near Victorville, California. In *Paleoenvironments and Paleohydrology of the Mojave and Southern Great Basin Deserts*, edited by Y. Enzel, S.G. Wells, and N. Lancaster, Geological Society of America Special Paper 368:1-42.
- Enzel, Y., S.G. Wells, and N. Lancaster. 2003. Late Pleistocene lakes along the Mojave River, southern California, pp 61-77 in *Paleoenvironments and Paleohydrology of the Mojave and Southern Great Basin Deserts*, edited by Y. Enzel, S.G. Wells, and N. Lancaster, Geological Society of America Special Paper 368.
- Fillmore, R.P., and J.D. Walker. 1996. Evolution of a supradetachment extensional basin: the early Miocene Pickhandle Basin central Mojave Desert, California. Geological Society of America Special Paper 303:107-126.
- Jefferson, G.T. 1985a. Stratigraphy and geologic history of the Pleistocene Lake Manix Formation, central Mojave Desert, California, p. 157-179 in *Cajon Pass to Manix Lake: Geological Investigations*

- Along Interstate 15, compiled by R.E. Reynolds, San Bernardino County Museum Publication, Redlands, California.
- Jefferson, G.T. 1985b. Review of the late Pleistocene avifauna from Lake Manix, central Mojave Desert, California. *Natural History Museum Los Angeles County Contribution in Science* 362:1-13.
- Jefferson, G.T. 2003. Stratigraphy and paleontology of the middle to late Pleistocene Manix Formation, and paleoenvironments of the central Mojave River, southern California, pp 43-60 *in* *Paleoenvironments and Paleohydrology of the Mojave and Southern Great Basin Deserts*, edited by Y. Enzel, S.G. Wells, and N. Lancaster, Geological Society of America, Special Paper 368.
- Leslie, S.R., D.M. Miller, J.L. Wooden, and J.A. Vazquez. 2010. Stratigraphy, age, and depositional setting of the Miocene Barstow Formation at Harvard Hill, central Mojave Desert, California, in *Overboard in the Mojave: 20 million years of lakes and wetlands*, R.E. Reynolds and D. M. Miller (eds). *Desert Studies*, California State University Fullerton, this volume.
- Meek, N. 1989. Geomorphic and hydrologic implications of the rapid incision of Afton Canyon, Mojave Desert, California. *Geology* 17:7-10.
- Meek, N. 1990. Late Quaternary Geochronology and geomorphology of the Manix basin, San Bernardino County, California. Doctoral Dissertation, Department of Geography, University of California, Los Angeles 212 p.
- Meek, N. 2000. The late Wisconsinan history of the Afton Canyon area, pp. 33-34 *in* *Empty Basins, Vanished Lakes*, edited by R.E. Reynolds and J. Reynolds, San Bernardino County Museum Association Quarterly 47(2).
- Meek, N. 2004. Mojave River history from an upstream perspective, pp. 41-49 *in* *Breaking Up*, edited by R.E. Reynolds, California State University, Desert Studies Consortium, Desert Symposium Field Trip and Abstracts.
- Miller, D.M., S.R. Leslie, M.C. Reheis, K.A. Schmidt, L. Amoroso, S.L. Dudash, and S.A. Mahan. 2009. Late Quaternary faulting and distributed deformation in the central Mojave Desert, California. *Geological Society of America Abstracts with Programs*, v. 41, http://gsa.confex.com/gsa/2009AM/finalprogram/abstract_166956.htm
- Nagy, E.A., and B.C. Murray. 1991. Stratigraphy and intra-basin correlation of the Mojave River Formation, Central Mojave Desert, California. *San Bernardino County Museum Association Quarterly* 38(2):5-30.
- Reheis, M.C. and D. M. Miller. 2010. A half-million years of paleoclimate and effects on landscape, Mojave Desert, California, *in* *Abstracts of proceedings, the 2010 Desert Symposium: this volume*.
- Reheis, M.C. and D.M. Miller. 2010. Environments of nearshore lacustrine deposition in the Pleistocene Lake Manix basin, south-central California, in *Overboard in the Mojave: 20 million years of lakes and wetlands*, R.E. Reynolds and D. M. Miller (eds). *Desert Studies*, California State University Fullerton, this volume.
- Reheis, M.C., D.M. Miller, and J.L. Redwine. 2007. Quaternary stratigraphy, drainage-basin development, and geomorphology of the Lake Manix basin, Mojave Desert: Guidebook for Fall Field Trip, Friends of the Pleistocene, Pacific Cell, October 4-7, 2007: U.S. Geological Survey Open-File Report v. 2007-1281, 31 p.
- Reheis, M.C., and J.I. Redwine. 2008. Lake Manix shorelines and Afton Canyon terraces: implications for incision of Afton Canyon. In *Late Cenozoic Drainage History of the Southwestern Great Basin and Lower Colorado River Region: Geologic and Biotic Perspectives*, edited by M.C. Reheis, R. Hershler, and D.M. Miller, Geological Society of America Special Paper 439:227-259.
- Park, L.E. 1995. Geochemical and paleoenvironmental analysis of lacustrine arthropod bearing concretions of the Barstow Formation, southern California. *Palaios* 10:44-57.
- Park, L.E., and K.F. Downing. 2001. Paleocology of an exceptionally preserved arthropod fauna from lake deposits of the Miocene Barstow Formation, southern California, U.S.A. *Palaios* 16:175-184.
- Reynolds, R. E. and B. F. Cox. 1999. Tracks along the Mojave: a field guide from Cajon Pass to the Manix Basin and Coyote Lake. *San Bernardino County Museum Association Quarterly* 46(3):1-26.
- Reynolds, R.E., and M.O. Woodburne. 2002. Marker bed correlations between the Mud Hills, Calico Mountains, and Daggett Ridge, central Mojave Desert, California, *in* *Between the Basins: Exploring the Western Mojave and Southern Basin and Range Province*, edited by R.E. Reynolds, California State University, Desert Studies Consortium p. 82.
- Reynolds, R.E., D. M. Miller, M. O. Woodburne, and L. B. Albright, 2010. Extending the boundaries of the Barstow Formation in the central Mojave Desert, in *Overboard in the Mojave: 20 million years of lakes and wetlands*, R.E. Reynolds and D. M. Miller (eds). *Desert Studies*, California State University Fullerton, this volume.
- Schuiling, W. T. 1999. A Miocene hot spring exhalite in the southern Calico Mountains, Redlands. *San Bernardino County Museum Association Quarterly* 46(3):89-94.
- Singleton, J. S., and P. B. Gans. 2008. Structural and stratigraphic evolution of the Calico Mountains: implications for early Miocene extensions and Neogene transpression in the central Mojave Desert, California. *Geosphere* 4:459-479.
- Spencer, R. S. 2006. Arthropod-bearing lacustrine deposits, Black Canyon, Barstow Formation. Master of Science, Geology, Loma Linda University, California 100 p.
- Steinmetz, J. J. 1987. Ostracodes from the late Pleistocene Manix Formation, San Bernardino County, California. *Mojave Desert Quaternary Research Center Proceedings*, San Bernardino County Museum Association Quarterly 34(3, 4):46-47.
- Wells, S.G., and Y. Enzel. 1994. Fluvial geomorphology of the Mojave River in the Afton Canyon area, eastern California—implications for the geomorphic evolution of Afton Canyon, *in* *Geological Investigations of an Active Margin*, edited by S.F. McGill and T.M. Ross, Geological Society of America, Cordilleran Section Guidebook p. 177-182.
- Wells, S. G., W. J. Brown, Y. Enzel, R. Y. Anderson, and L. D. McFadden. 2003. Late Quaternary geology and paleohydrology of pluvial Lake Mojave, southern California, *in* Y. Enzel, S.G. Wells, and N. Lancaster (eds), *Paleoenvironments and paleohydrology of the Mojave and southern Great Basin Deserts: Boulder, Colo.*, Geological Society of America Special Paper 368, p. 79-114.
- Woodburne, M. O., R. H. Tedford, and C. C. Swisher III. 1990. Lithostratigraphy, biostratigraphy, and geochronology of the Barstow Formation, Mojave Desert, southern California. *Geological Society of America Bulletin* 102:459-477.

Reconnaissance geochronology of tuffs in the Miocene Barstow Formation: implications for basin evolution and tectonics in the central Mojave Desert

D. M. Miller¹, S. R. Leslie¹, J. W. Hillhouse¹, J. L. Wooden¹, J. A. Vazquez¹, and R. E. Reynolds²

¹U.S. Geological Survey, 345 Middlefield Road, Menlo Park, CA 94025

²Redlands, California, 92373

Abstract

Early to middle Miocene lacustrine strata of the Barstow Formation are well dated in just a few places, limiting our ability to infer basin evolution and regional tectonics. At the type section in the Mud Hills, previous studies have shown that the lacustrine interval of the Barstow Formation is between ~16.3 Ma and ~13.4 Ma. Elsewhere, lake beds of the Barstow Formation have yielded vertebrate fossils showing the Hemingfordian/Barstovian transition at ~16 Ma but are otherwise poorly dated. In an attempt to clarify the age and depositional environments of the lake deposits, we are mapping the Barstow Formation and dating zircons from interbedded tuffs, as well as testing ash-flow tuffs for the distinctive remanent magnetization direction of the widespread Peach Spring Tuff.

Thus far, our new U-Pb zircon ages indicate that the Barstow lake beds contain tuff beds as old as 19.1 Ma and as young as 15.3 Ma. At Harvard Hill, Barstow lake beds contain a thick tuff dated at 18.7 Ma. On the basis of zircon ages, mineralogy, zircon chemistry, and paleomagnetic results, we consider the thick tuff to be a lacustrine facies of the Peach Spring Tuff. We have identified the Peach Spring Tuff by similar methods at eight localities over a broad area, providing a timeline for several fluvial and lacustrine sections.

The new dates indicate that long-lived lacustrine systems originated before 19 Ma and persisted to at least 15 Ma. The onset of lacustrine conditions predates the Peach Spring Tuff in most Barstow Formation sections and may be older than 19.5 Ma in some places. The new data indicate that the central Mojave Desert contained narrow to broad lake basins during and after extension, and that Barstow lacustrine deposits did not exclusively postdate extensional tectonics. At present, it is unclear whether several separate, small lake basins coexisted during the early to middle Miocene, or if instead several small early Miocene basins gradually coalesced over about 6 million years to form one or two large middle Miocene lake basins.

Introduction

As part of an investigation of active and neotectonic faults in the Mojave Desert part of the Eastern California Shear Zone (ECSZ), we are studying the Barstow Formation and its basins to establish a pre-ECSZ template that could be used for reconstructing faults and distributed deformation in the central Mojave Desert. The Barstow Formation has been described as a post-extension basinal deposit (e.g., Fillmore 1993; Fillmore and Walker, 1996) and therefore should not require restoration for early Miocene extension associated with the central Mojave metamorphic core complex (CMMCC) and extension of similar timing in other areas.

This report provides preliminary results of our geochronology and geologic mapping of the Barstow Formation, summarizes these results along with the findings of Leslie et al. (this volume) and Hillhouse et al. (this volume), and provides preliminary interpretations. Based

on these results, the timing and depositional setting of Barstow Formation sections appear to be much more complex than heretofore appreciated.

Previous studies of the Barstow Formation

The sedimentary facies and depositional environments of the Barstow Formation were described by Link (1980). The Barstow Formation has been well dated at its type section in the Mud Hills (Fig. 1), where six tuff and tuffaceous sandstone beds have been dated by $^{40}\text{Ar}/^{39}\text{Ar}$ methods and magnetostratigraphy (MacFadden et al., 1990b). The lower part of the formation lies above the 19.3 Ma Red Tuff and the uppermost part is dated by the 13.4 Ma Lapilli Sandstone (Fig. 2). This sequence is well-characterized for vertebrate fossils and serves as the reference section for the Barstovian land mammal stage (e.g., Woodburne et al., 1990) of the upper Miocene. Woodburne et al. (1990) and Woodburne (1991) describe the

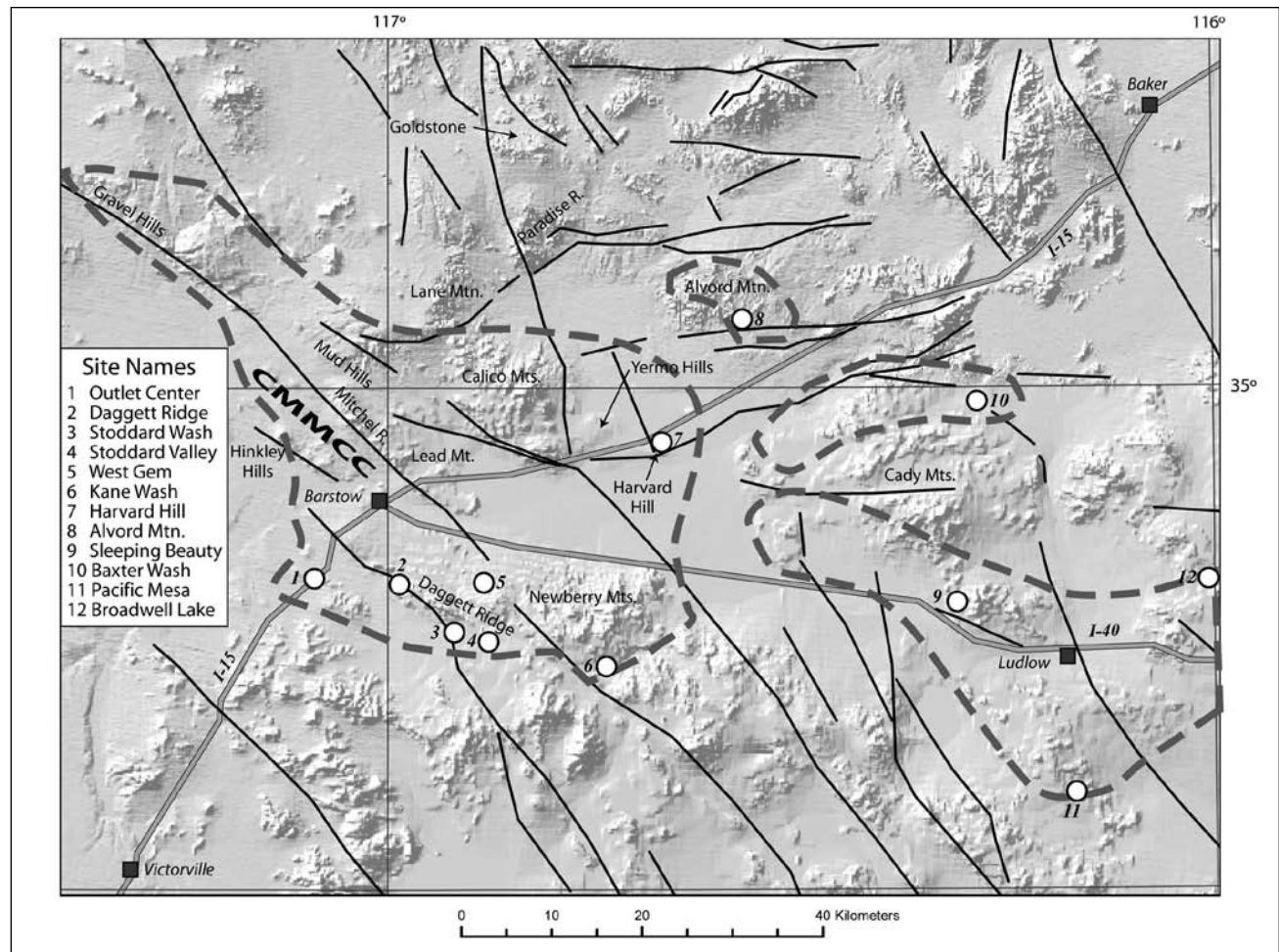


Figure 1. Location map for central Mojave Desert study area, showing locations referred to in the text, major Quaternary faults (black) from Miller et al. (2007). Numbered sites are Peach Spring Tuff localities described by Hillhouse et al. (this volume) and referred to in this paper. Gray dashed lines: Approximate outline of Miocene basinal deposits in the region. CMMCC, central Mojave metamorphic core complex, is centered on main culminations of mylonitized Miocene granodiorite and older rocks (after Glazner et al., 2002).

type section as a lower unit of fluvial conglomerate succeeded by a thick lacustrine interval. The lacustrine part is approximately bracketed by the 16.3 Ma Rak Tuff and the 13.4 Lapilli Sandstone.

Singleton and Gans (2008) studied the Barstow Formation and volcanic rocks in the Calico Mountains, establishing that the underlying Pickhandle Formation is as young as 19.0 Ma and that volcanic and intrusive rocks dated at 17.1 to 16.8 Ma intrude and are deposited on the Barstow Formation. They considered the Barstow Formation to be primarily lacustrine and about 19.0 to 16.9 Ma (Singleton and Gans, 2008). However, study of vertebrate fossils in the Calico section of the Barstow Formation indicates that the upper part of the section ranges from older than 16 Ma to 14.8 Ma (Reynolds and others, this volume), and therefore must have continued to be deposited after the 17.1 to 16.8 Ma volcanic episode in the Calico Mountains. One possible resolution of this discrepancy is to call on the volcanic breccia within the Barstow sequence to be an avalanche deposit.

Woodburne (1991, 1998; Woodburne and Reynolds, this volume) established the locations and characteristics of Miocene strata across the central Mojave, demonstrating that the Hector Formation in the Cady Mountains (Fig. 1) is partly age-equivalent to the Barstow Formation and also that it contains the Peach Spring Tuff. The Peach Spring Tuff is a regional distinctive ash-flow tuff (Gusa et al., 1987) that emanated from a caldera in the Colorado River corridor (Pearthree et al., 2009), flowing east onto the Colorado Plateau and west to the region of Barstow (Glazner et al., 1986).

Reynolds (2000, 2004, this volume) established a sequence of marker units in the Barstow Formation at the type section and in many other exposures, providing an opportunity to correlate strata across the basin. A key uncertainty in correlating the marker units is knowing to

what extent they are time-transgressive. Because tuffs are commonly altered in places other than the type section and not amenable to K-Ar dating methods, and paleontologic ambiguities persist, this chronologic uncertainty has hampered interpretations.

Tectonic setting

The Barstow Formation is well documented at its type section in the Mud Hills and the nearby Gravel Hills, Calico Mountains, and Lead Mountain (Woodburne, 1991; Fillmore and Walker, 1996). Dokka et al. (1988) used the term to describe lacustrine strata in the Daggett Ridge area and Byers (1960) termed the upper part of a thick Miocene sequence at Alvord Mountain the Barstow Formation. Glazner (1980) suggested that the Barstow Formation may extend to the southern margin of the Cady Mountains on the basis of lacustrine deposits that overlie the Peach Spring Tuff. This is the area we will discuss, and henceforth will call it the "Barstow area."

The Barstow area is underlain by pre-Miocene rocks of many kinds: Triassic, Jurassic, and Cretaceous granitoids; Mesozoic metavolcanic rocks; and Paleozoic metamorphic rocks (e.g., Walker et al., 2002; Miller and Walker, 2002). The area was apparently uplifted and beveled by erosion during the early Cenozoic (Glazner et al., 2002), setting the stage for Neogene tectonism.

Miocene extensional tectonics characterized by NE-directed extension and accompanying normal faults, basin development, and magmatism began about 24 Ma and continued to at least 19 Ma (e.g., Dokka, 1989; Fletcher et al., 1995; Glazner et al., 2002). The cessation of extensional tectonics is a point of disagreement, with various methods of inferring this time (end of basin sedimentation, end of magmatism, cooling rates in footwall rocks) giving results ranging from 19 Ma to 17 Ma. The extensional tectonics resulted in an exposed core complex in

the Mitchel Range and Hinkley Hills along with footwall granitoid plutons that are dated loosely at about 24 to 21 Ma (Glazner et al., 2002). Northeast of the core complex, 1- to 2-km-thick sequences of early and middle Miocene sediment formed in northwest-elongate basins above an

Gravel Hills	Mud Hills	Calico Mts	Yermo Hills	Harvard Hill	Alvord Mtn	N. Daggett R	S Daggett R	N. Cady Mts
		Yermo Gvl	Yermo Gvl		Langford Gravels			
Barstow Fm.	Barstow Fm.	Barstow Fm.	Barstow Fm.	Barstow Fm.	Barstow Fm.	Barstow Fm.	Barstow Fm.	Hector Fm.
	~19.3-~13.5	19.0-14.8	>16-15.3	>19.1-<18.5	~18-<13.5	>19-<18.5	>19.5-18.5	
Gravel & sand	Lake sed above gvl	Lake sed	Lake sed	Lake sed w/ breccias	Lake sed w/ volcanics	Lake sed	Lake sed	Lake and fluvial
Pickhandle Fm.	Pickhandle Fm.	Pickhandle Fm.			Spanish Canyon Fm.	volcanic breccia	volcanic breccia	Hector Fm.
Volcanic & Clastic	Volcanic & clastic	Volcanic & clastic			volcanic			Fluvial and lake
		Jackhammer Fm.			Clews Fm.			volcanics
		Clastic			Clastic and Lake			
Color codes: blue = Pliocene yellow = early, middle and late Miocene green = early Miocene								

Figure 2. Stratigraphic terminology for the locations mentioned in the text. Data from sources cited in text as well as Miller and Yount (2002).

east-dipping detachment fault. The sedimentary rocks are typically grouped as the lower volcanogenic Pickhandle Formation and the upper lacustrine and alluvial Barstow Formation. Although the Early Miocene strata in places have undergone vertical axis rotations (Ross et al., 1989; Hillhouse, this vol.), it has been argued that the rotations were accomplished by small blocks rotating in the hangingwall during detachment faulting, rather than regional crustal blocks rotating (Glazner et al., 2002), a conclusion that is supported by studies of similar phenomena in the Colorado River extended terranes (Wells and Hillhouse, 1989).

Starting no earlier than 11 to 12 Ma (Schermer et al., 1996), strike-slip faulting of the Eastern California Shear Zone (ECSZ) began across the region of former extensional basins. Driven by nearly north-south maximum stress, northwest-striking dextral faults and conjugate northeast-striking sinistral faults formed across the Barstow area (Fig. 1). In the west, the dextral faults partly reactivated former normal faults (Singleton and Gans, 2008); whereas, in the east, sinistral faults formed. Studies of vertical-axis rotations of the ECSZ tectonic blocks show that blocks associated with dextral faults have not rotated since Peach Spring Tuff time (~18.5 Ma; Nielson et al., 1990) but that blocks associated with sinistral faults have rotated as much as ~45 degrees (Schermer et al., 1996; Ross et al., 1989; Hillhouse et al., this volume). A consequence of this rotation is that early Miocene extensional tectonic elements have been rotated in these sinistral blocks but not in areas of dextral faulting. ECSZ faulting accumulated a few tens of km of dextral shear across the Mojave Desert, with faults in the Barstow area typically assigned total offsets of about 3 to 9 km (Jachens et al., 2002). As a result, Miocene extensional basins have been shuffled and in some cases rotated but not displaced great distances (see Hillhouse et al., this volume, for a summary of offsets).

Methods

Field studies employed geologic mapping by standard methods supplemented by use of LiDAR-generated shaded relief maps and custom aerial photography. We mapped limited areas in detail (parts of the Calico Mountains, Yermo Hills, Toomey Hills, and Harvard Hill) along with reconnaissance study of many areas.

We attempted to date tuffs within altered lacustrine sequences by use of U-Pb geochronology, using the joint Stanford-USGS SHRIMP-RG ion microprobe. Tuff samples were crushed, ground, and heavy minerals separated by use of heavy liquids, sieving, and magnetic sorting. Zircons were hand-picked under a binocular microscope,

embedded in epoxy mounts along with standards, and polished and imaged with cathodoluminescence. The images were used as guides for siting the SHRIMP probe spots for analysis; individual analyses used a 5-6 nA primary beam of O_2^- to sputter an approximately 15 to 30 micrometer pit, allowing for separate zones in zircon crystals to be analyzed. In addition to radiogenic daughter products of the U-Pb system, we collected data on several trace and rare-earth elements. Concentrations for U, Th and all of the measured trace elements are standardized against well-characterized, homogeneous zircon standard MAD, and U-Pb ages are referenced to zircon standard R33 (419 Ma, Black et al., 2004). Data were reduced using programs SQUID 1 and Isoplot (Ludwig, 2001, 2002).

Results

Our early explorations of tuffs yielded the unexpected result of a possible Peach Spring Tuff correlative deposited in lacustrine rocks at Harvard Hill (Leslie et al., this volume), leading us to expand the study to include subaerial Peach Spring Tuff ignimbrite deposits and related rhyolites. As a result, a significant amount of work was performed on the Peach Spring Tuff. This work is described first in this section, followed by studies of other tuffs.

Peach Spring Tuff

The Peach Spring Tuff is a regional ignimbrite formed by explosive eruption of rhyolitic magma from the Colorado River area near Oatman, AZ (Pearthree et al., 2009). Pyroclastic density currents that formed the Peach Spring Tuff traveled east on to the Colorado Plateau and west to at least Barstow (Glazner et al., 1986; Wells and Hillhouse, 1989), resulting in the ability to characterize remanent magnetization in the tuff on the stable Colorado Plateau and contrast that with values obtained farther west in several tectonic settings. The Peach Spring Tuff is about 18.5 Ma based on Ar dating of sanidine in pumice blocks in two places (Nielson et al., 1990; Miller et al., 1998). Where previously known in the Barstow area, the Peach Spring Tuff is interbedded with lacustrine and alluvial deposits that unconformably overlie faulted volcanic strata: in the Sleeping Beauty area (fig. 1, location 9) of the southern Cady Mountains (Glazner, 1980), Newberry Mountains (fig. 1, location 6; Cox, 1995), and south of Daggett Ridge (fig. 1, location 3; Dokka et al., 1988; Wells and Hillhouse, 1989). We have identified the Peach Spring Tuff in four additional locations by U-Pb dating of zircon, zircon chemistry, and mineralogy of the deposit. These locations are described in the following paragraphs along with a newly dated site that was earlier correlated by

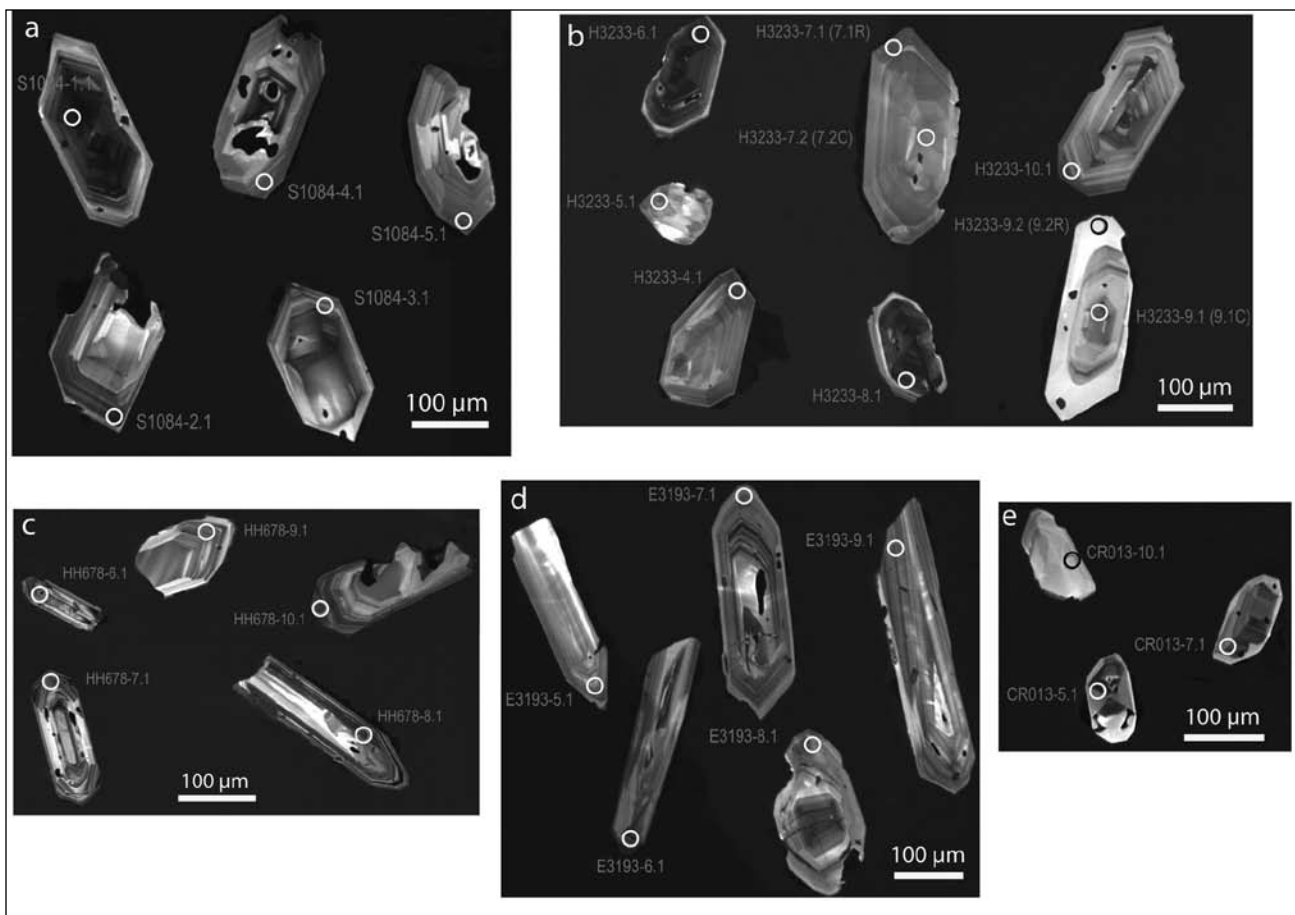


Figure 3. Cathodoluminescence images of representative zircons from samples studied. Circles represent sample sites.

magnetic properties.

Stoddard Wash

The site of Wells and Hillhouse (1989) lies south of Daggett Ridge in a section of thin-bedded lacustrine sandstone and siltstone (Dibblee, 1970) considered by Dokka et al. (1988) to be the Barstow Formation (fig. 1, location 3). Thin interbeds of fine gravel ~10 m below the tuff indicate that the location was near the shore of the lake somewhat before the ignimbrite was emplaced. The ignimbrite deposit is 3 to 4 m thick, slightly welded, and carries lithic and pumice fragments. Down-section in the lacustrine beds are the marker units of Reynolds (2004), suggesting that a significant lake accumulation occurred in this area before the Peach Spring Tuff was erupted. Zircon crystals are complexly zoned and commonly contain melt inclusions (Fig. 3). U-Pb ages (sample SL09NS-1084) are strongly bimodal as shown by the histogram (Fig. 4c); modes are at about 18.6 Ma and 18.3 Ma. The probability distribution demonstrates that a single broad peak is best supported given the errors for individual analyses. However, large MSWD (Table 1) indicates that the population may be significantly non-Gaussian. A significant population of grains is younger

than the accepted age of 18.5 $^{40}\text{Ar}/^{39}\text{Ar}$ age Ma for the Peach Spring Tuff. Zircons from this sample possess high Th/U, compared to many other tuffs we sampled (Fig. 6). This distinctive zircon chemistry is found in samples studied from Peach Spring Tuff outcrops near Kingman (C.F. Miller, written commun., 2009), apparently confirming it as a Peach Spring Tuff characteristic.

Shamrock Tuff (Harvard Hill)

The Shamrock tuff, informally named by Leslie et al. (this volume), is a thick green waterlain tuff that contains sanidine, biotite, and sphene throughout, along with rare hornblende and common pumice and lithic fragments. It is greater than 7 m thick at the north end of Harvard Hill (fig. 1, location 7), where sampled. Glass shards are aligned parallel to overlying limestone beds of the Barstow Formation. Zircons from sample M09NS-3233 yield a strong mode at 18.5 to 18.6 Ma (fig. 4a), with many older grains to 19.8 Ma, as well as a significant group of younger grains, several about 18.0 Ma (Leslie et al, this volume). A second sample (SL09HH-871), from bedded, probably wave-reworked tuff overlying the massive Shamrock tuff yielded similar results with a strengthened emphasis on younger zircons about 18.1 Ma (fig. 4b)

and an overall peak at about 19.1 Ma. Together, the two samples from the Shamrock yield a broad distribution of zircon ages from about 18.0 to 20.0 Ma. Based on these age data and other similarities, Leslie et al. (this volume) correlated the Shamrock with the Peach Spring Tuff.

Gem Columbus Mine

On the northeast side of Daggett Ridge (fig. 1, location S), a thin, pale pink tuff that contains sanidine and quartz, as well as abundant lithic fragments, lies in sequence stratigraphically above lacustrine beds that include the

marker units of Reynolds (2004). The lacustrine beds apparently lie on granitic alluvial fan beds, which lie on Mesozoic granite basement (Dibblee, 1970; Reynolds, 2004). The tuff is weakly welded. Zircons included a significant detrital fraction identified partly by their Mesozoic and Proterozoic ages. Restricting this discussion to the Miocene zircons, the result is a unimodal population centered on roughly 18.7 Ma (Fig. 4d), with most grains yielding ages between 18.6 and 19.0 Ma, and the youngest grains at about 17.0 Ma. Zircon chemistry matches

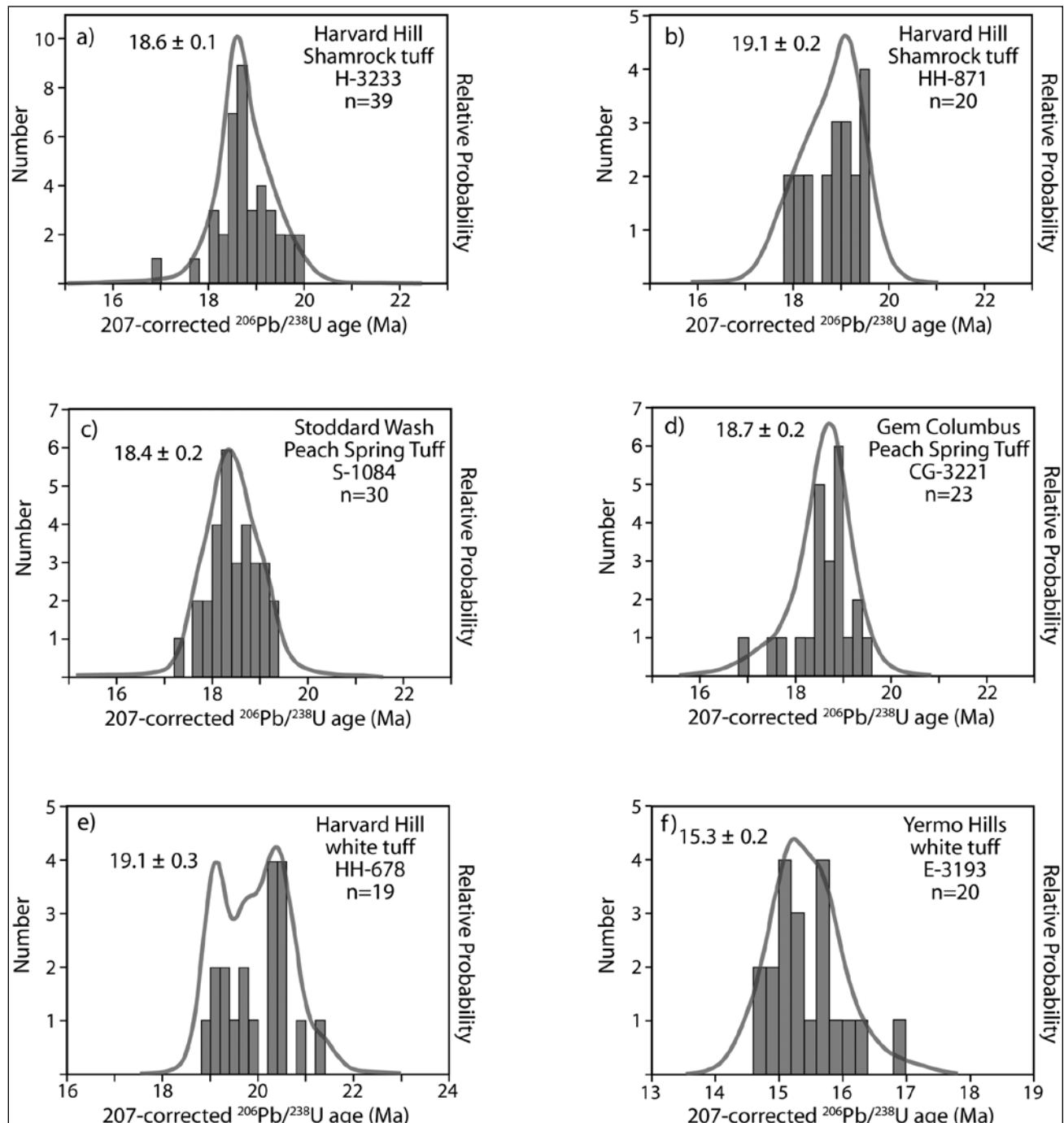


Figure 4. Plots of zircon age data. For each sample, histogram of non-detrital zircons is presented as well as a probability density distribution. Age values for peaks in probability density distribution are labeled.

that of other Peach Spring Tuff samples studied (Fig. 6).

Baxter Wash

The northeastern Cady Mountains were studied by Miller (1980), who mapped and K-Ar dated (at 17.9 Ma) a prominent pink ignimbrite in the middle of the Hector Formation (fig. 1, location 10). Although Miller later considered the tuff to be the Peach Spring Tuff (oral commun. 1984), it was not described as such by MacFadden et al (1990a) who studied the magnetostratigraphy of the Hector Formation. Woodburne (1998) later used the Peach Spring Tuff for this section and others in the Cady Mountains. The Peach Spring Tuff is about 10 m thick and lies below vaguely bedded muddy fine sandstone, tuffaceous sandstone, and calcite-cemented conglomerate, all apparently lacustrine. It is pink and contains visible sanidine, sphene, and minor biotite. The tuff is moderately welded. Zircon data from sample M09NS-2684 support the paleomagnetic (Hillhouse et al., this volume) and mineralogic correlation with the Peach Spring Tuff: four grains confirm an age of about 18.7 Ma and have zircon chemistry similar to that for Stoddard Wash (Table 1).

Summary of Peach Spring Tuff dates

Many of the samples we studied yield zircon ages spread across nearly 2 million years, from about 17.5 Ma to 19.5 Ma. Several samples yielded two modes of ages, one near 18.6 Ma and another near 18.2 Ma. In general, a 'tail' of ages extends from the 18.6 Ma mode to as old as 19.5 Ma to 19.8 Ma. All of the zircons included for the plots of Fig. 6 shared the high Th/U described for the Stoddard Wash tuff. The combined probability distribution curves for the four fully analyzed Peach Springs samples (fig. 5a) shows significant overlap in the curves, supporting the interpretation from Th/U and mineralogy that all samples represent a single tuff. We grouped all Peach Spring Tuff zircon data and calculated statistics for the result (fig. 5b). The weighted mean age for all Peach Spring Tuff zircons ($n=116$) is 18.66 ± 0.08 Ma (2σ) with a MSWD of 2 (Table 1). The grouped data have a distinct age peak at 18.6 to 18.7 Ma. Although many older and younger zircons exist, they do not define discrete peaks. This result may indicate that some samples contained too few zircons to fully describe the population, but it may also indicate that there is sample-to-sample variation in zircon age characteristics.

The Peach Spring Tuff was also identified by paleomagnetic methods in several locations (Fig. 1) that are briefly described as follows. At Stoddard Valley, a location (fig. 1, location 4) about 7 km east of the Stoddard Wash site, a pink tuff similar to that at Stoddard Wash lies in a section of parallel-bedded sandstone. The slightly welded

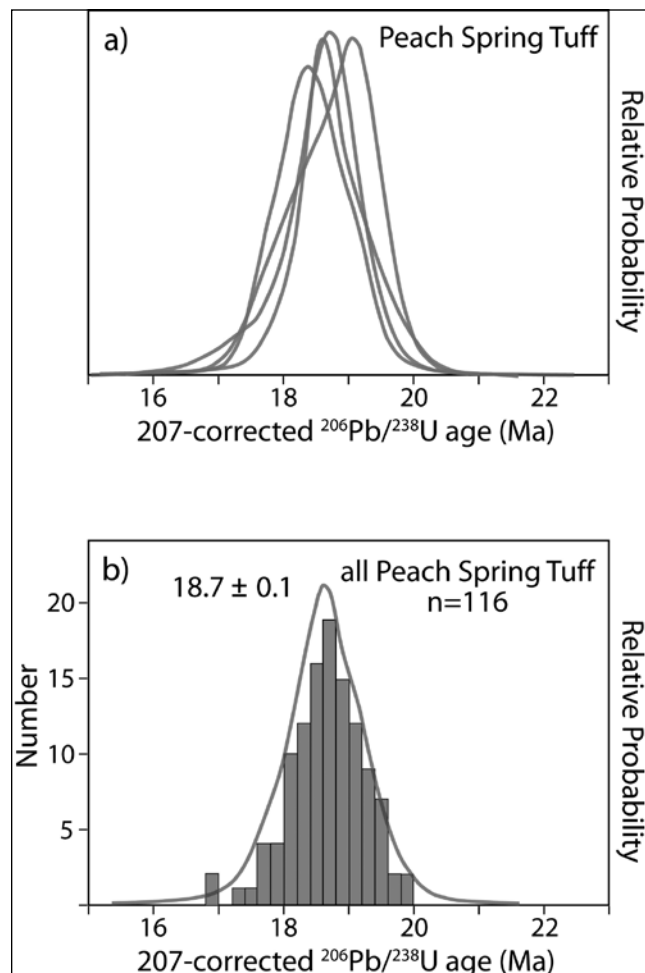


Figure 5. Summary plots of zircon age data for the Peach Spring Tuff samples. A) Composite of probability density distribution curves from plots in Fig. 4. B) Histogram and probability density distribution plot for all zircon grains.

tuff contains sanidine, as well as minor sphene, quartz, and biotite grains. Lithic fragments are common. The unit is about 4 m thick. Paleomagnetic study (Hillhouse et al., this volume) demonstrates that the tuff is the Peach Spring Tuff by virtue of low inclination. At West Gem, a pink tuff similar in most respects to the tuff at Columbus Gem Mine lies about 2 km to the west of the mine (fig. 1, location 5). It occurs stratigraphically above distal alluvial fan and thin lacustrine beds and below lacustrine sandstone, siltstone, and mudstone. The tuff carries biotite and hornblende as well as minor quartz. Paleomagnetic study at this site strongly suggests correlation with the Peach Spring Tuff (Hillhouse et al., this volume), as does similarity to the zircon-dated pink tuff at the Columbus Gem Mine locality. At Outlet Center (fig. 1, location 1), brownish-pink tuff exposed in quarries carries the characteristic remanent magnetization of the Peach Spring Tuff (Hillhouse et al., this volume). The tuff is 3 m thick, weakly to moderately welded, and carries sanidine, quartz, hornblende, and biotite. The stratigraphic setting

is uncertain due to poor exposures, but this outcrop is the farthest west exposed Peach Spring Tuff.

At the Northern Cady Mountains, Moseley (1978) mapped a pink ignimbrite in the Hector Formation and correlated it to a similar ignimbrite in the nearby Baxter wash section of Miller (1980), which we correlated (see above) with the Peach Spring Tuff. The Peach Spring Tuff as mapped by Moseley is a westward-thinning unwelded tuff that locally overlies a thick white tuff, probably comagmatic and similar to deposits described by Valentine et al. (1989), in a section of distal alluvial fan sandstone and thin beds of lacustrine sandstone. At Alvord Mountain (fig. 1, location 8), the Peach Spring Tuff was mapped by Fillmore (1993) as part of his “two tuffs” marker unit above the Clews Formation. The Peach Spring Tuff there is reported on by Hillhouse et al. (this volume), who describe it as filling channels in an alluvial sequence.

Other dated tuffs. We dated three additional tuffs that are interbedded with lake deposits interpreted to be the Barstow Formation at Harvard Hill, the Yermo Hills, and Calico Mountains. All of these structural blocks lack the Peach Spring Tuff.

South Harvard Hill (fig. 1, location 7). This white tuff, which is biotite-bearing, lies on thin lake beds of fine sand to silt. It lies a few meters below the base of the lowermost rock-avalanche breccia described by Leslie et al. (this

volume). U-Pb results for 19 Miocene zircon grains are bimodal (Fig. 4e), with peaks at 19.1 and 20.4 Ma. The youngest mode, 19.1 Ma, is taken as the best approximation for the age of the tuff bed. Mesozoic and Proterozoic zircons in the tuff indicate a source from the south, which may reflect either sediment that the tuff entrained or the path of the tuff flow itself (Leslie et al., this volume).

Yermo Hills (Emerald Basin). Emerald Basin in the western Yermo Hills is named informally for its green mudstone and altered tuff that appear in various shades of green. The strata in the basin form an asymmetric anticline that has a steep limb adjacent to a fault on the west side of the basin; most strata dip east. Near the base of the sequence are a few intervals of coarse sand and fine gravel, some bearing well-rounded pebbles (Reynolds, 2004; Reynolds et al., this volume) that were sourced near Goldstone, because they include distinctive red and green calcsilicate metasedimentary rocks. Upward, the section is mostly thin-bedded fine sandstone and mudstone, with over ten tuff beds. We sampled the highest relatively coarse-grained tuff bed in the section, a medium- to coarse-grained greenish-white biotite tuff about 80 cm thick. Glass is altered to a pale yellowish white or green material (probably a clay mineral), in many places, and replacement by silica is also widespread. The sample was of glass-shard tuff with little alteration. Zircons in the tuff are distinctive shape and size (Figure 3) and although

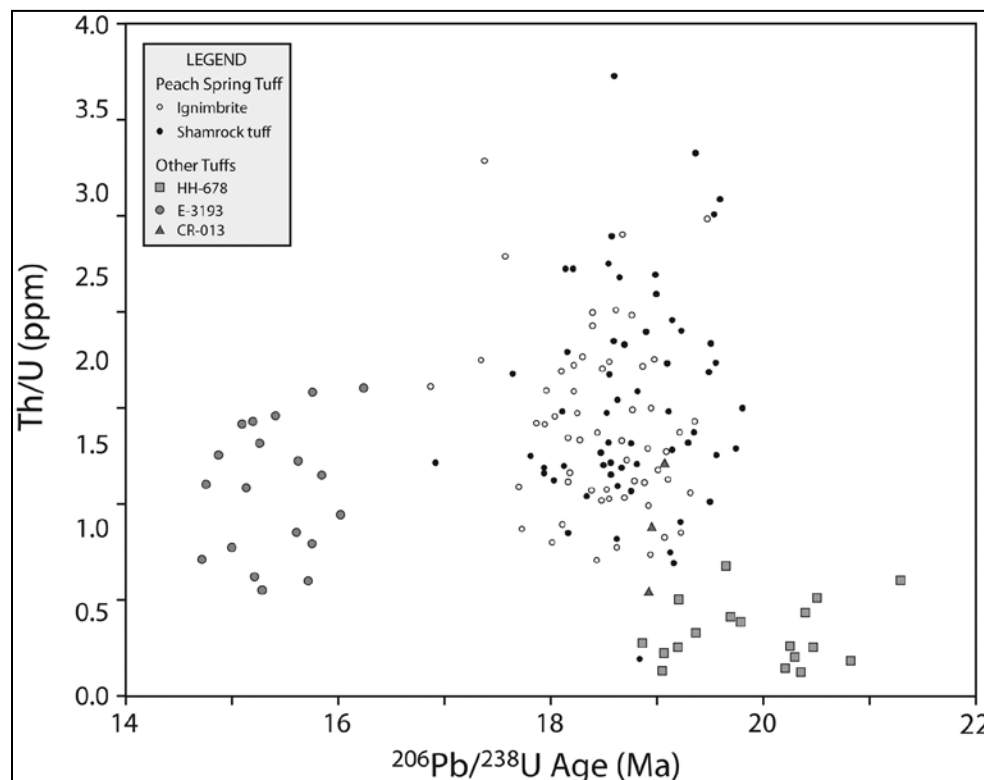


Figure 6. Plot of age vs. Th/U for zircon grains dated. The only overlap of values for Peach Spring tuff zircons and others is near Th/U=1.0.

probable detrital zircons of various sizes and shapes are present, we mainly probed the long, thin oscillatory zoned zircons. We dated 20 zircon grains that yielded a bimodal histogram but a single peak in the probability density curve (fig. 4f). The peak of the probability curve is 15.3 ± 0.2 Ma, whereas the younger histogram mode is about 15.1 Ma and four zircons are younger than 15.0 Ma. We consider the tuff bed to be 15.3 Ma or somewhat younger.

Southern Calico Mountains (Cemetery Ridge). Cemetery Ridge is a well-exposed section of the Barstow Formation that is bracketed by two strands of the Calico

fault (Singleton and Gans, 2008) and lies west of Calico ghost town. Barstow beds are tightly folded and also appear to undergo lateral facies changes from west to east of gravel- and sand-dominated alluvial fan conglomerate to lacustrine fine sandstone and limestone beds. Within the section are two white tuffaceous sandstone beds. We sampled the upper bed, which appears to be interbedded with arenaceous sandstone in places. The sample yielded few, and assorted, zircon grains. We explored 5 grains, two of which yielded old ages (23.5 and 168 Ma) and three of which yielded ages between 18.9 and 19.1 Ma. The coherent group of three yields a weighted mean age of 19.0 ± 0.3 Ma. This result is exploratory at this time, and more zircons are needed to confirm or deny the age of the sandstone bed.

Geochronology interpretation

Several factors can complicate the simplest scenario for interpreting zircon U-Pb ages to determine a tuff's eruption age. The simplest scenario is that zircons nucleate and grow for a short time in a magma body before eruption terminates crystal growth and propels zircon, ash, and other mineral grains to distant locations. In this simplest scenario, tuff will have a single, sharply defined peak of zircon grain ages with Gaussian dispersion about the peak representing analytical imprecision (e.g., Simon et al., 2009). However, because zircon is not easily resorbed and recrystallized in magmas, pre-eruption zircons are commonly entrained in magmas (e.g., Bacon and Lowenstern, 2005; Charlier et al., 2005; Miller et al., 2007) as a result of rejuvenation of young intrusions and assimilation of wall rocks. In addition, during an eruption the forceful explosion of magma through capping rocks may entrain older zircons. For these scenarios, the zircon ages will be skewed to older ages, and the youngest mode may approximate the age of eruption.

After the juvenile magma erupts, the processes of distribution and deposition can also affect the zircon population. Simple airfall tuff, reflecting atmospheric transport

and fallout, should produce a size-sorted but otherwise unmodified zircon population from the magma. Multiple air falls may have changing zircon populations bed to bed, however. Eruption of rhyolitic magma may yield contrasting zircon age populations if the parent magma chamber is compositionally zoned or is composed of juxtaposed magma bodies (e.g., Charlier et al., 2003). In addition, dynamic sorting during the highly energetic runout process, and entrainment of zircons from underlying older deposits may occur (e.g., Buesch, 1992). Lastly, a tuff bed may be reworked by fluvial, eolian, and lacustrine processes, which may introduce foreign materials.

All of these complicating factors may introduce older zircon grains to the juvenile population. As a result, an age distribution showing a prominent mode with a tail skewed to old ages is expected, and weighted mean ages of the distribution are unreliable. Zircons incorporated into the tuff from foreign materials are generally easily recognized because they are much older; they are not considered in our analysis. In addition, if the tuff is reworked significantly after deposition, younger zircons are also introduced.

In general, the age distributions we measured for Miocene tuffs conform with the scenarios outlined above: a single-peak probability density distribution that is skewed to older ages. Histograms reveal complex, generally bimodal age distributions, but the probability density distributions, which take into account the errors associated with each grain analysis (generally 1σ of 0.2 to 0.5 Ma), tend to be simpler and indicate that detail in the histogram is not statistically significant. Either the peak probability value or a slightly younger age is therefore taken as the age of the eruption. However, a significant proportion of the zircon grains we analyzed for Peach Spring Tuff is younger than the peak age, and some fairly large samples, such as the S-1084 sample ($n=30$), yielded peak values younger than the overall population. As described by Leslie et al. (this volume), the zircon grains that yield ages younger than ~ 18.5 Ma are not easily discounted.

Table 1. Summary of U-Pb data for zircon from 8 Miocene tuff samples. Weighted mean ages and errors are for the coherent groups of zircons that are chemically similar and excludes outlier zircons with Mesozoic and Proterozoic ages.

Sample Number	Name	Location	No. Grains Analyzed	Weighted Mean Age (Ma) & Error	MSWD	Peak Density Age (Ma) ¹	Interpreted Age (Ma) and Error ²
M09NS-3233	H-3233	Harvard Hill	39	18.74 ± 0.13	2.1	18.63	18.6 ± 0.1
SL09HH-871	HH-871	Harvard Hill	20	18.93 ± 0.23	2.3	19.11	19.1 ± 0.2
M09NS-3221	CG-3221	NE Daggett R.	23	18.66 ± 0.17	1.5	18.73	18.7 ± 0.2
SL09NS-1084	S-1084	Stoddard Valley	30	18.41 ± 0.16	2.3	18.37	18.4 ± 0.2
M09NS-2684	B-2684	N. Cady Mts.	4	18.68 ± 0.18	1.02		18.7 ± 0.2^3
		All Peach Spring Tuff samples	116	18.66 ± 0.08	2.3	18.67	18.7 ± 0.1
SL09HH-678	HH-678	Harvard Hill	19	19.86 ± 0.34	7.2	19.14	19.1 ± 0.3
M09NS-3193	E-3193	Yermo Hills	20	15.43 ± 0.21	2.0	15.27	15.2 ± 0.2
M10NS-013	CR-013	Calico Mts.	3	18.99 ± 0.27	0.26		19.0 ± 0.3^3

¹ Age value for youngest peak of probability density distribution if multimodal. No value if number of analyses is small.

² Age from probability density distribution. No Th-U disequilibrium correction applied. Error estimated from weighted mean age and error.

³ Weighted mean and standard error for zircon population

Although many of these younger grains have larger errors than typical, tests for Pb loss in their rims and other evidence for damage of some kind have failed to identify the source of younger ages. The possibility that the Peach Spring Tuff is as young as 18.0 Ma cannot be excluded, nor the possibility that the U-Pb and K-Ar systems yield inherently discordant results.

This interpretation of the age of the Peach Spring Tuff from U-Pb zircon data can be tested against ages determined by other methods. Ages have been determined using the K-Ar system (Nielsen et al., 1990; Miller et al., 1998) and magnetostratigraphy (Hillhouse et al., this volume). The Ar determined age for sanidine from pumice blocks in the Peach Spring Tuff is 18.5 to 18.8 Ma, depending on assumptions made for the age of monitor minerals (see Hillhouse et al., this volume, for discussion). This timing corresponds well with the ~18.7 Ma magnetostratigraphic position of the Peach Spring Tuff in the Hector Formation of the northeastern Cady Mountains (Hillhouse et al., this volume). The Ar-determined age for the Peach Spring Tuff corresponds well with the peak of the probability density distribution peak for the aggregated Peach Spring Tuff samples (116 zircon grains) of 18.7 ± 0.1 Ma (Fig. 4) and for the most completely studied Peach Spring Tuff sample (H-3233, Harvard Hill, 18.7 ± 0.1 Ma; $n=39$). Other samples vary in peak probability age but have smaller populations of zircon grains and presumably are less well constrained (Table 1).

Other factors must be considered in this Peach Spring Tuff inter-method comparison. Shärer (1984) found that Th-U disequilibrium in low-Th zircon grains effectively reduces the age recorded by zircons by nearly 0.1 Ma, and requires adjustment. In addition, a comparison of Ar dates on sanidine and U-Pb dates on zircon must consider uncertainties in both isotopic systems, the possibility of differential sorting of sanidine and zircon populations in the tuff, and entrainment of old sanidine that is not fully reset (Spell et al., 2001). In view of these complications, the match of the 18.7 Ma U-Pb zircon age in the Peach Spring Tuff and the 18.5 ± 0.1 Ma to 18.8 ± 0.1 Ma Ar age for sanidine in the Peach Spring Tuff is satisfactory. We find that our interpretation of probability density distributions to estimate age of a tuff bed, even without adjustment for Th-U disequilibrium effects, is reliable. Table 1 provides interpreted U-Pb ages for the tuffs we analyzed.

Synthesis of geochronology

Our new U-Pb dates indicate that lacustrine beds in previously undated sections of the Barstow Formation range from older than 19.1 Ma to younger than 15.3 Ma. This new data set somewhat expands the range of ages previ-

ously reported for the Mud Hills and Calico Mountains, but more importantly, it greatly expands the Barstow exposures where lacustrine strata are considered to be early Miocene, as old as the Peach Spring Tuff. Our study of Peach Spring Tuff has provided a dated marker demonstrating that several lacustrine sections of the Barstow Formation in the Daggett Ridge area predate the ~18.7 Ma Peach Spring Tuff. Using these data, Reynolds et al. (this volume) suggest that lacustrine deposition in two areas near Daggett Ridge commenced as early as 19.5 Ma.

The Peach Spring Tuff is now dated by multiple studies of $^{40}\text{Ar}/^{39}\text{Ar}$ from sanidine in several locations in California and Arizona (Nielsen et al., 1990; Miller et al., 1998; Hillhouse et al., this volume) yielding either about 18.5 ± 0.1 Ma to 18.8 ± 0.1 Ma depending on assumptions about ages for monitor minerals, and with our dating of 116 zircon grains from various localities. At this time, we consider the best determination for the age of the Peach Spring Tuff to be about 18.7 Ma, based on the magnetostratigraphic interpretation, but as more is learned about the details of the laboratory methods, assumptions, and the volcanic system itself, this age may change somewhat.

Depositional history and paleogeography

Precursor basin sedimentary sequences to the Barstow Formation have been attributed to extensional tectonic environments (e.g., Dokka, 1989; Woodburne, 1991; Fillmore, 1993; Cox, 1995; Fillmore and Walker, 1996). The Pickhandle Formation (~24 to 19 Ma) is typified by volcanic flows, tuff breccias, volcanoclastic sediments, rock avalanche breccias, and subordinate alluvial and lacustrine sediment (Fillmore and Walker, 1996). Basin studies from east of Barstow northwest to the Gravel Hills demonstrate rapid Pickhandle sedimentation from nearby steep basin margins on both the southwest and northeast (Dokka et al., 1991; Fillmore and Walker, 1996). In the Calico Mountains, the type Pickhandle is mainly volcanic flows and domes, and the basin was bounded on the southwest by a normal fault now locally occupied by the younger Calico fault (Singleton and Gans, 2008). Farther northeast, down dip of the detachment fault, Pickhandle type rocks are not present. The Clews and Spanish Canyon formations at Alvord Mountain probably represent similar timing. However, there, the alluvial and volcanic section includes avalanche breccia derived from the east, suggesting an extensional basin (Fillmore, 1993).

Farther south, contrasting interpretations of Daggett Ridge, Newberry Mountains, and areas to the east have created ambiguity. Glazner (1980), Dokka (1991), Cox (1995), and Miller (1994) have documented steeply tilted fault blocks of ~24-20 Ma volcanic rocks in sev-

eral places spanning the Newberry, Rodman, Cady, and Bristol Mountains and Lava Hills. Evidence for considerable extension above one or more shallow detachment faults seems to be inescapable. Similarly, volcanic breccia at Daggett Ridge of about the same age indicates steep basin margins and a probable extensional basin. However, Glazner et al (2002) interpreted this area as relatively unextended.

Several breaks in stratigraphy suggest basin margins and faults during Pickhandle time. The contrast between early Miocene at Daggett Ridge and the Newberry Mountains suggests separate fault blocks. Similarly the contrast between the Calico and Mud Hills stratigraphy suggest separate blocks.

Barstow Formation succeeded the Pickhandle with slight angular unconformity at the Mud Hills and Calico Mountains (Fillmore and Walker, 1996, Singleton and Gans, 2008) but with sharp angular discordance on "Pickhandle" breccia at Daggett Ridge. Equivalent sandstone that is interbedded with the Peach Spring Tuff at the Newberry Mountains is also angularly discordant over underlying tilted volcanic rocks (Cox, 1995; pers. commun., 2009). Lower Barstow lithofacies vary from alluvial in the Gravel Hills and Mud Hills, and northeast Daggett Ridge and Newberry Mountains, to lacustrine at southern Daggett Ridge, Calico Mountains, Yermo Hills area, and Harvard Hill. At Calico Mountains, breccias and coarse sediment from the south were deposited immediately before the Barstow (Singleton and Gans, 2008) and similar facies exist at Harvard Hill (Leslie et al., this volume). These observations point to basin dimensions for early Barstow sediments that were about the same as those for the upper Pickhandle.

Succeeding lacustrine rocks of the Barstow are harder to use to indicate basin margins but several observations suggest small basins for the lower lacustrine intervals: First, roundstone conglomerate, interpreted as fluvial in origin, is interbedded with lake beds at the Yermo and Toomey Hills, at Alvord Mountain (Byers, 1960) and at several other locations (Reynolds et al., this volume; Miller unpubl. mapping, 2009). The pebbles of these gravels are derived from distant northwesterly sources such as Goldstone and the Paradise Range area, suggesting longitudinal transport along lengthy basin axes. Second, lacustrine rock in the Mud Hills and Calico Mountains are mainly fine-grained clastics: sandstone and siltstone, punctuated by a few limestone beds that are of shallow water origin by virtue of stromatolites. In contrast, lacustrine rocks of the Yermo and Toomey Hills are mudstone, siltstone, and very fine sandstone lacking limestone. The different facies indicate a rapid facies change or separate basins. Third, Leslie et al. (this volume) suggested that

presence of Peach Spring Tuff as a thick lacustrine facies in the lower Barstow at Harvard Hill but not at other locations may indicate that the lake basins were separate. However, a notable feature of lacustrine rocks of the Barstow Formation is a dearth of coarse-grained nearshore facies such as beach gravel. This observation indicates a lack of steep shorelines, wave energy, or coarse material. Alluvial gravel is present in lower Barstow Formation in several places and fluvial gravels were introduced in several places, so the lack of nearshore facies indicates broad, shallow lakes with gently sloping margins. This conclusion requires that any segmentation of lake basins was by low, gently sloping ridges and hills.

Near the southern margin of the Barstow basin at Daggett Ridge, stratigraphic relations are relatively well exposed along tilted panels exposed nearly continuously for ~10 km east-west. On the north side of Daggett Ridge, sections containing the Peach Spring Tuff range from Columbus Gem mine on the east, to West Gem, to northwest Daggett Ridge on the west. They show east to west younging of lake beds. This change of depocenter is consistent with down-to-the-west tilting or with clastic influx from the east, forcing the depocenter to migrate westward. South of Daggett Ridge, Peach Spring Tuff lies high in the Barstow section: In the east, it lies on interbedded alluvial and lacustrine sandstone, whereas in the west it lies on a thick lacustrine section (Dibblee, 1970; Reynolds et al., this volume).

The marker bed sequence in lacustrine rocks of the Barstow Formation identified by Reynolds (2000, 2004) apparently indicates similar environmental conditions that are repeated in one or more lakes from place to place. The new dates on tuffs, however, make clear that the marker beds vary in age by as much as 3 million years (e.g., brown platy limestone below the Peach Spring Tuff at Columbus Gem and near the 16.3 Ma Rak tuff in the Mud Hills). The marker beds will continue to be an excellent tool for unraveling the Barstow Formation history as more chronological control is added.

Lacustrine beds in the upper Hector Formation and the Barstow Formation at Alvord Mountain are interbedded with volcanic rocks and alluvial fan facies, suggesting that these basins were separate from basins farther west typified by the fine-grained persistent lacustrine facies of the Barstow Formation.

Tectonic implications

A full discussion of the tectonic implications of our new dates and geologic mapping is beyond the scope of this paper because results are incomplete Miller et al. (2007) and a complete restoration of post-Miocene faulting is

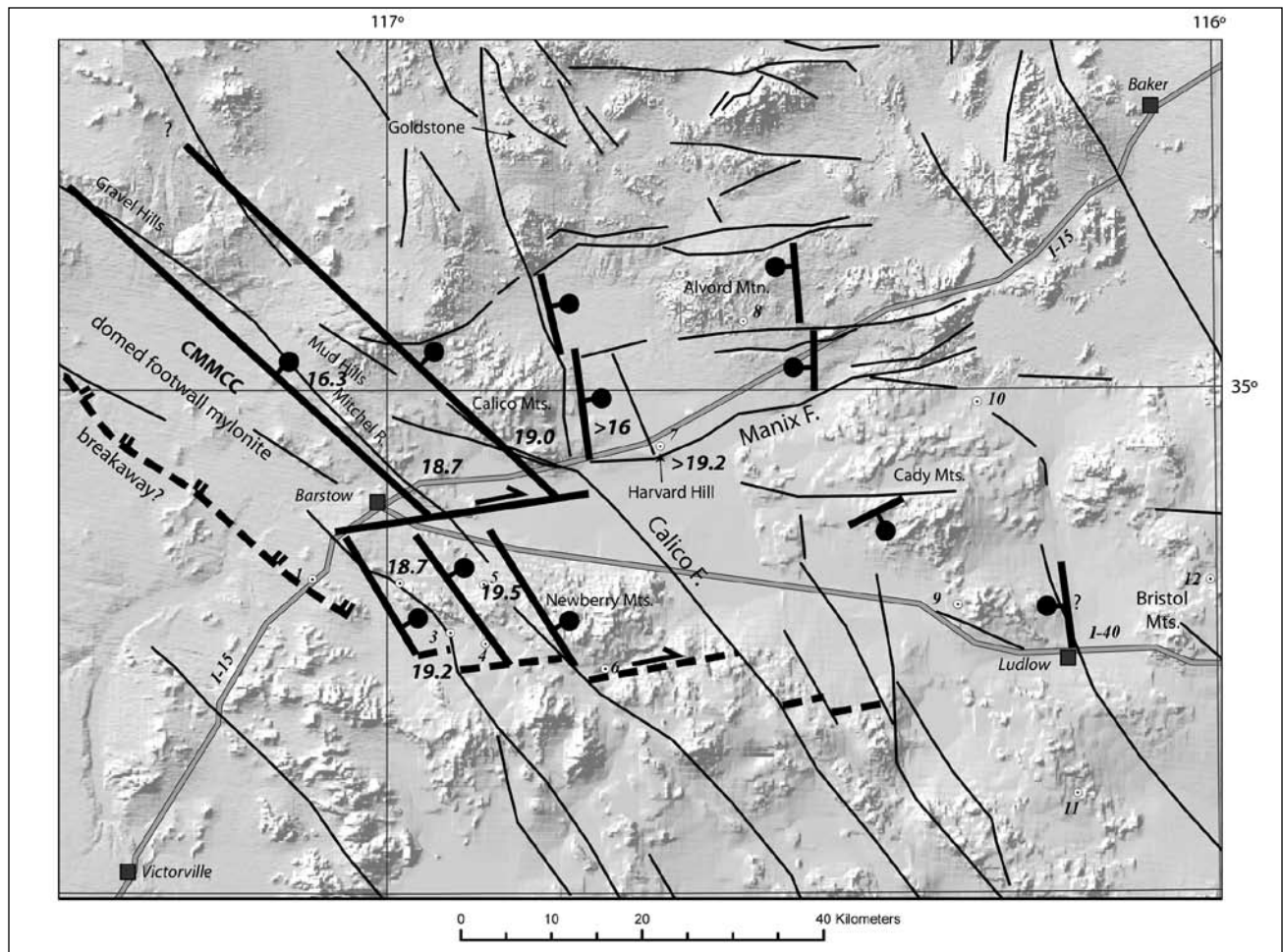


Figure 7. Interpretive map of central Mojave Desert extensional basin geometry. Each basin is labeled with the approximate age (in Ma) of the base of lacustrine strata. Data from sources cited in text. Heavy lines are normal faults; dashed lines approximate boundaries of significant extension

required to adequately evaluate early and middle Miocene tectonics. However, a few clarifications for Miocene tectonics are possible at this time.

Miocene basins inferred from basin-margin facies (Fillmore, 1993; Fillmore and Walker, 1996; Dokka et al., 1988) and rare identification of the bounding faults (Singleton and Gans, 2008) are 10-15 km wide, and northwest-elongate (Fig. 7). Sedimentary facies in both early and middle Miocene indicate longitudinal transport down long basin axes and transverse transport from nearby sides of basins. These dimensions offer clues for inferring early Miocene basins where covered in the greater Mojave River plain east of Barstow. Total thickness of early and middle Miocene strata where well studied are about 2 km, and should be distinguishable as gravity lows. Gravity inversion by Jachens et al. (2002) indicate that both exposed Miocene basins and potential basins beneath cover are not currently thicker than 1 km, with two exceptions: 1) a 2-3 km thick basin modeled for the area underlain by the city of Barstow, and 2) a ~2 km thick basin northeast of the Calico fault and south of the

Manix fault (fig. 7). Inferred basins at these locations may represent extensional or strike-slip tectonism. Except for these locations, thick Miocene sediments do not seem to occur in the region, which suggests that in most places Miocene basin sediments were thinner than 1 km or that they have undergone partial uplift and erosion.

Arguments that the ~24-19 Ma Pickhandle Formation and its correlatives represent extensional basin fill and overlying ~19-14 Ma Barstow Formation and its correlatives represent post-extension sedimentation (Fillmore and Walker, 1996; Glazner et al., 2002), while reasonable in a general way, need modification as more details on Miocene geochronology emerge. It is now clear that the Barstow lacustrine sediments are time-transgressive, with initial lacustrine sediments ranging from about 16.3 Ma to older than 19.1 Ma. Lacustrine depocenters shifted laterally with time as marked by the position of the Peach Spring Tuff in basins near Daggett Ridge, indicating that basins were responding to tectonism. Even basins far from the CMMCC and its breakaway zone (Alvard Mountain, Cady Mountains) were distinct, fairly narrow, and re-

ceived coarse clastic sediments in to the middle Miocene (Byers, 1960; Miller, 1980; Moseley, 1978). We suspect that these lines of evidence point to gradual decline of extensional tectonism from 19 to ~17 Ma, more in accord with proposals by Dokka (1989). This conclusion is also in accord with interpretation of thermochronology data for rocks of the core complex by Gans et al. (2005), who concluded that maximum unroofing and extension rates were from 21 to 17.5 Ma.

The magnitude of extension for core complexes can vary considerably perpendicular to the extension direction, and deserves to be re-examined for the CMMCC. Geologic maps for core complexes in the northern Great Basin clearly demonstrate that east-directed extension died out rapidly to the north (Albion Mountains) and south (Ruby Mountains). With the lateral reduction in extension, changes in thermal history and structural style also occur. A similar situation for the CMMCC may hold. Basins appear to become narrower and die out to the northwest because no basins 24 to 18 Ma in age are known across wide areas of Fort Irwin (Schermer et al, 1996). Southward from domal culminations of early Miocene mylonitic rocks, 24 to 20 Ma volcanic strata in the Newberry Mountains are steeply tilted by closely spaced normal faults that force the interpretation of shallow underlying, unexposed detachment faults (Cox, 1995; unpublished mapping). Breccia farther west at Daggett Ridge may indicate similar faulting. Farther east, in the southern Cady Mountains (Woodburne, et al., 1974) and Lava Hills and southern Bristol Mountains (Miller, 1994), similar relations are observed: 24 to 20 Ma volcanic rocks are tilted in narrow fault blocks, all overlain with angular discordance by the Peach Spring Tuff and associated sediments. Evidence for extension seems to be unequivocal, and we suggest that these relations at the Newberry Mountains and Daggett Ridge represent different tectonic styles than the CMMCC.

Conclusions

New U-Pb dates on zircon in tuff beds within lacustrine rocks of the Barstow Formation show that the beds range in age from older than 19.1 Ma to younger than 15.3 Ma. Along with U-Pb dating of the Peach Spring Tuff, the results point to much of the Barstow lacustrine facies being early Miocene, significantly older than counterpart lacustrine strata of the type section in the Mud Hills. The new data clarify patterns of lateral migration of the lacustrine depocenters from east to west, as well as clarifying correlations with the Clews Formation and Hector Formation farther to the east. Emerging lithofacies data for the Barstow Formation, when combined with chronology

information and insights from underlying Pickhandle deposits, indicate that the early Barstow Formation was probably deposited in northwest-elongate, separate narrow basins. It remains unclear if, and when, these basins coalesced.

Our data, along with data from the Calico Mountains (Singleton and Gans, 2008) indicate that the transition from coarse volcanoclastic deposits of the Pickhandle Formation to generally lacustrine deposits of the Barstow Formation is not a simple shift from extensional basins to post-extensional deposition. Barstow-type lacustrine deposits temporally overlap Pickhandle-type volcanic breccias, and the lacustrine depocenters migrate laterally, implying tilt-block basin settings for some early Miocene Barstow Formation lacustrine rocks. These observations indicate that extensional tectonism waned but continued after the 19.0 to 19.3 Ma termination of the Pickhandle deposition.

Acknowledgements

Discussions with Brett Cox, Pete Sadler, and Kevin Schmidt about their unpublished studies in the area have aided our work substantially. Kim Bishop kindly aided with our interpretation of rock avalanche breccia at Harvard Hill. Discussion about limestones in the Barstow Formation with Vicki Pedone was illuminating, as were discussions about the Peach Spring Tuff with Jonathan Miller and Calvin Miller. We thank Ariel Strickland and Dean Miller for their help with several phases of the mineral separations and preparation of mounts for the SHRIMP laboratory. We also thank Paul Stone and Sue Beard for helpful comments on an early draft of this paper.

References

- Bacon, C.R., and Lowenstern, J.B., 2005, Late Pleistocene granodiorite source for recycled zircon and phenocrysts in rhyodacite lava at Crater Lake, Oregon: *Earth and Planetary Science Letters*, v. 233, p. 277-293.
- Black, L.P., Kamo, S.L., Allen, C.M., Davis, D.W., Aleinikoff, J.N., Valley, J.W., Mundil, R., Campbell, I.H., Korsch, R.J., Williams, I.S., and Foudoulis, C., 2004, Improved 206Pb/238U microprobe geochronology by the monitoring of a trace-element-related matrix effect; SHRIMP, ID-TIMS, ELA-ICP-MS and oxygen isotope documentation for a series of zircon standards: *Chemical Geology* v. 205, p. 115-140.
- Buesch, D.C., 1992, Incorporation and redistribution of locally derived lithic fragments within a pyroclastic flow: *Geological Society of America Bulletin*, v. 104, p. 1193-1207.
- Byers, F.M., Jr., 1960, *Geology of the Alvord Mountain quadrangle, San Bernardino County, California*: U.S. Geological Survey Bulletin, 1089-A, 71 p.
- Charlier, B.L.A., Peate, D.W., Wilson, C.J.N., Lowenstern, J.B., Storey, M., Brown, S.J.A., 2003, Crystallization ages in coeval silicic magma bodies: 238U-230Th disequilibrium evidence from the Rototiti

- and Earthquake Flat eruption deposits, Taupo volcanic zone, New Zealand. *Earth and Planetary Science Letters*, 206: 441-457.
- Charlier, B.L.A., Wilson, C.J.N., Lowenstern, J.B., Blake, S., Van Calsteren, P.W., and Davidson, J.P., 2005, Magma generation at a large, hyperactive silicic volcano (Taupo, New Zealand) revealed by U-Th and U-Pb systematics in zircons: *Journal of Petrology*, v. 46, p. 3-32.
- Cox, B.F., 1995, Structure and tectonics at the south edge of an Early Miocene rift zone, central Mojave Desert, California: *San Bernardino County Museum Association Quarterly*, v. 42, p. 30-31.
- Dibblee, T.W., Jr., 1970, Geologic map of the Daggett quadrangle, San Bernardino County, California: U. S. Geological Survey Miscellaneous Geologic Investigations Map, I-592, scale 1:62,500.
- Dokka, R.K., 1989, The Mojave extensional belt of southern California: *Tectonics*, v. 8, p. 363-390.
- Dokka, R.K., McCurry, M.M., Woodburne, M.O., Frost, E.G., and Okaya, D.A., 1988, A field guide to the Cenozoic crustal structure of the Mojave Desert, in: Weide, D.L., and Faber, M.L., eds. *This extended land: Geological journeys in the southern Basin and Range: Geological Society of America, Cordilleran Meeting, field trip guidebook*, p. 21-44.
- Dokka, R.K., Henry, D.J., Ross, T.M., Baksi, A.K., Lambert, J., Travis, C.J., Jones, S.M., Jacobson, C., McCurry, M.M., Woodburne, M.O., and Ford, J.P., 1991, Aspects of the Mesozoic and Cenozoic geologic evolution of the Mojave Desert: *Geological Society of America, San Diego, Annual Meeting, field trip guidebook*.
- Fillmore, R.P., 1993, Sedimentation and extensional basin evolution in a Miocene metamorphic core complex setting, Alvord Mountain, central Mojave Desert, California, USA: *Sedimentology*, v. 40, p. 721-742.
- Fillmore, R.P., Walker, J. D., 1996, Evolution of a supradetachment extensional basin: The lower Miocene Pickhandle Basin, central Mojave Desert, California: *Geological Society of America Special Paper 303*, p. 107-126.
- Fletcher, J.M., Bartley, J.M., Martin, M.W., Glazner, A.F., Walker, J.D., 1995, Large-magnitude continental extension: An example from the central Mojave metamorphic core complex: *Geological Society of America Bulletin*, v. 107, p. 1468-1483.
- Gans, P.B., DeVecchio, D., Singleton, J., Van Pelt, J., Wong, M., and Reynolds, J., 2005, Cenozoic magmatic and structural evolution of the central Mojave Desert, California: New constraints from $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology and thermochronology: *Geological Society of America Abstract with Programs*, p. 103.
- Glazner, A.F., 1980, Geology of the Sleeping Beauty area, southeastern Cady Mountains, In Fife, D.L. and Brown, A.R., eds., *Geology and Mineral Wealth of the California Desert: South Coast Geological Society*, p. 249-255.
- Glazner, A.F., Nielson J.E., Howard K.A., and Miller D.M., 1986, Correlation of the Peach Springs Tuff, a large-volume Miocene ignimbrite sheet in California and Arizona: *Geology*, v. 14, p. 840-843.
- Glazner, A.F., Walker, J.D., Bartley, J.M., and Fletcher, J.M., 2002, Cenozoic evolution of the Mojave block of southern California: *Geological Society of America Memoir 195*, p. 19-41.
- Gusa, S., Nielson, J.E., Howard, K.A., 1987, Heavy-mineral suites confirm the wide extent of the Peach Springs Tuff in California and Arizona, U.S.A., *Journal of Volcanology and Geothermal Research*, v. 33, p. 343-347.
- Hillhouse, J.W., Miller, D.M., and Turrin, B.D., this volume, Correlation of the Miocene Peach Springs Tuff with the geomagnetic polarity time scale and new constraints on tectonic rotations in the Mojave Desert, California
- Jachens R.C., Langenheim, V.E., Matti, J.C., 2002, Relationship of the 1999 Hector Mine and 1992 Landers fault ruptures to offsets on Neogene faults and distribution of late Cenozoic basins in the eastern California shear zone: *Bulletin of the Seismological Society of America*, v. 92, p. 1592-1605.
- Leslie, S.R., Miller, D.M., Wooden, J.L., and Vazquez, J.A., this volume, Stratigraphy, age, and depositional setting of the Miocene Barstow Formation at Harvard Hill, central Mojave Desert, California
- Link, M.H., 1980, Sedimentary facies and mineral deposits of the Miocene Barstow Formation, California, In Fife, D.L. and Brown, A.R., eds., *Geology and Mineral Wealth of the California Desert: South Coast Geological Society*, p. 191-203.
- Ludwig, K., 2001, *Isoplot/EXversion 2.49: a geochronological toolkit for Microsoft Excel*. Berkeley Geochronology Center Special Publication, v. 1a.
- Ludwig, K., 2002, *SQUID 1.02, a Users Manual*: Berkeley Geochronology Center Special Publication, v. 2.
- MacFadden, B.J., Woodburne, M.O., and Opdyke, N.D., 1990a, Paleomagnetism and Neogene clockwise rotation of the northern Cady Mountains, Mojave Desert of southern California: *Journal of Geophysical Research*, v. 95, p. 4597-4608.
- MacFadden, B.J., Swisher, C.C. III, Opdyke, N.D., and Woodburne, M.O., 1990b, Paleomagnetism, geochronology, and possible tectonic rotation of the middle Miocene Barstow Formation, Mojave desert, southern California: *Geological Society of America Bulletin*, v. 102, p. 478-493.
- Miller, D.M., 1994, Cenozoic deposits in the Lava Hills and southern Bristol Mountains, southeastern California, in Sherrod, D.R., and Nielson, J.E., eds., *Tertiary Stratigraphy of highly extended terranes, California, Arizona, and Nevada: U.S. Geological Survey Bulletin 2053*, p. 99-107.
- Miller, D.M., and Yount, J.L., 2002, Late Cenozoic tectonic evolution of the north-central Mojave Desert inferred from fault history and physiographic evolution of the Fort Irwin area, California: *Geological Society of America Memoir 195*, p. 173-197.
- Miller, D.M., Dudash, S.L., Green, H.L., Lidke, D.J., Amoroso, L., Phelps, G.A., and Schmidt, K.M., 2007, A new Quaternary view of northern Mojave Desert tectonics suggests changing fault patterns during the late Pleistocene, in Miller, D.M. and Valin, Z.C., eds., *Geomorphology and tectonics at the intersection of Silurian and Death Valleys, southern California: U.S. Geological Survey Open-File Report 2007-1424*, p. 157-171.
- Miller, J.S., Heizler, M. T., and Miller, C. F., 1998, Timing of magmatism, basin formation, and tilting at the west edge of the Colorado River Extensional Corridor: Results from single-crystal $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology of Tertiary rocks in the Old Woman Mountains area, southeastern California: *Journal of Geology*, v. 106, p. 195-209.
- Miller, J.S., Matzel, J.E.P., Miller, C.F., Burgess, S.D., and Miller, R.B., 2007, Zircon growth and recycling during the assembly of large, composite arc plutons: *Journal of Volcanology and Geothermal Research*, v. 167, p. 282-299.
- Miller, J.S., and Walker, J.D., 2002, Mesozoic geologic evolution of Alvord Mountain, central Mojave Desert, California: *Geological Society of America Memoir 195*, p. 59-77.
- Miller, S.T., 1980, Geology and mammalian biostratigraphy of a part of the northern Cady Mountains, Mojave Desert, California: U.S. Geological Survey Open-File Report 80-878, 122 p.
- Moseley, C.G., 1978, Geology of a portion of the northern Cady Mountains, Mojave Desert, California: Univ. Calif. Riverside, Dept. of Earth Sciences (Master's Thesis), 131 p.
- Nielson, J.E., Lux, D.R., Dalrymple, G.B. and Glazner, A.F., 1990, Age of the Peach Springs Tuff, southeastern California and western Arizona: *Journal of Geophysical Research*, v. 95, p. 571-580.
- Pearthree, P.A., Ferguson, C.A., Johnson, B.J., and Guynn, J., 2009, Geologic map and report for the proposed State Route 95 realign-

- ment corridor, Mohave County, Arizona, version 1.0: Arizona Geological Survey, Digital Geologic Map Series DGM 65, 47 p., 5 sheets, 1:24,000 scale.
- Reynolds, R.E., 2000, Marker units suggest correlation between the Calico Mountains and the Mud Hills, central Mojave Desert, California: San Bernardino County Museum Quarterly, v. 47, no. 2, p. 1-10.
- Reynolds, R.E., 2004, Widespread early Miocene marker beds unite the Barstow Formation of the central Mojave Desert: Inland Geological Society, p. 1-10.
- Reynolds, R.E., Miller, D.M., and Woodburne, M.O., this volume, Extending the boundaries of the Barstow Formation in the central Mojave Desert
- Ross, T.M., Luyendyk, B.P., and Haston, R.B., 1989, Paleomagnetic evidence for Neogene clockwise tectonic rotations in the central Mojave Desert, California: *Geology*, v. 17, p. 470-473.
- Schermer, E.R., Luyendyk, B.P., Cisowski, S., 1996, Late Cenozoic structure and tectonics of the northern Mojave Desert: *Tectonics*, v. 15, p. 905-932.
- Shärer, U., 1984, The effect of initial ^{230}Th disequilibrium on young Pb ages: the Makalu case, Himalaya: *Earth and Planetary Science Letters*, v. 67, p. 191-204.
- Simon, J.L., Vazquez, J.A., Renne, P.R., Schmitt, A.K., Bacon, C.R., Reid, M.R., 2009, Accessory mineral U-Th-Pb ages and $^{40}\text{Ar}/^{39}\text{Ar}$ eruption chronology, and their bearing on rhyolitic magma evolution in the Pleistocene Coso volcanic field, California: *Contributions to Mineralogy and Petrology*, DOI 10.1007/s00410-009-0390-9.
- Singleton, J.S., and Gans, P.B., 2008, Structural and stratigraphic evolution of the Calico Mountains: Implications for early Miocene extensions and Neogene transpression in the central Mojave Desert, California: *Geosphere*, v. 4, p. 459-479.
- Spell, T.L., Smith, E.I., Sanford, A., Zanetti, K.A., 2001, Systematics of xenocrystic contamination: preservation of discrete feldspar populations at McCullough Pass Caldera revealed by $^{40}\text{Ar}/^{39}\text{Ar}$ dating: *Earth and Planetary Science Letters*: v. 190, p. 153-165.
- Valentine, G.A., Buesch, D.C., Fisher, R.V., 1989, Basal layered deposits of the Peach Springs Tuff, northwestern Arizona, USA: *Bulletin of Volcanology*, v. 51, 395-414.
- Walker, J.D. Martin, M.W., and Glazner, A.F., 2002, Late Paleozoic to Mesozoic evolution of the Mojave Desert and environs, California: *Geological Society of America Memoir* 195, p. 1-18.
- Wells, R.E., and Hillhouse, J.W., 1989, Paleomagnetism and tectonic rotation of the lower Miocene Peach Springs Tuff: Colorado Plateau, Arizona, to Barstow, California: *Geological Society of America Bulletin*, v. 101, p. 846-863.
- Woodburne, M.O., 1991, The Mojave Desert Province, in *Inland Southern California: the last 70 million years*: San Bernardino County Museum Association Quarterly, v. 38, p. 60-78.
- Woodburne, M.O., 1998, Arikareean and Hemingfordian faunas of the Cady Mountains, Mojave Desert province, California: *Geological Society of America Special Paper* 325, p. 197-210.
- Woodburne M.O., and Reynolds, R.E., this volume, The mammalian litho- and biochronology of the Mojave Desert Province
- Woodburne, M.O., Tedford, R.H., Stevens, M.S., and Taylor, B.E., 1974, Early Miocene mammalian faunas, Mojave Desert, California: *Journal of Paleontology*, v. 48, p. 65-76.
- Woodburne, M.O., Tedford, R.H., and Swisher, C.C., III., 1990, Lithostratigraphy, biostratigraphy and geochronology of the Barstow Formation, Mojave Desert, southern California. *Geological Society of America Bulletin*, v. 102, p.459-477.

Stratigraphy, age, and depositional setting of the Miocene Barstow Formation at Harvard Hill, central Mojave Desert, California

Shannon R. Leslie¹, David M. Miller¹, Joseph L. Wooden², and Jorge A. Vazquez¹

¹U.S. Geological Survey, 345 Middlefield Road, Menlo Park, CA 94025

²U.S.G.S.-Stanford Ion Microprobe Laboratory, Stanford University, Stanford, CA 94305

Abstract

New detailed geologic mapping and geochronology of the Barstow Formation at Harvard Hill, 30 km east of Barstow, CA, help to constrain Miocene paleogeography and tectonics of the central Mojave Desert. A northern strand of the Quaternary ENE-striking, sinistral Manix fault divides the Barstow Formation at Harvard Hill into two distinct lithologic assemblages. Strata north of the fault consist of: a green rhyolitic tuff, informally named the Shamrock tuff; lacustrine sandstone; partially silicified thin-bedded to massive limestone; and alluvial sandstone to pebble conglomerate. Strata south of the fault consist of: lacustrine siltstone and sandstone; a rhyolitic tuff dated at 19.1 Ma (U-Pb); rock-avalanche breccia deposits; partially silicified well-bedded to massive limestone; and alluvial sandstone and conglomerate.

Our U-Pb zircon dating of the Shamrock tuff by SHRIMP-RG yields a peak probability age of 18.7 ± 0.1 Ma. Distinctive outcrop characteristics, mineralogy, remanent magnetization, and zircon geochemistry (Th/U) suggest that the Shamrock tuff represents a lacustrine facies of the regionally extensive Peach Spring Tuff (PST). Here we compare zircon age and geochemical analyses from the Shamrock tuff with those of the PST at Stoddard Wash and provide new insight into the age of zircon crystallization in the PST rhyolite.

Results of our field studies show that Miocene strata at Harvard Hill mostly accumulated in a lacustrine environment, although depositional environments varied from a relatively deep lake to a very shallow lake or even onshore setting. Rock-avalanche breccias and alluvial deposits near the base of the exposed section indicate proximity to a steep basin margin and detrital studies suggest a southern source for coarse-grained deposits; therefore, we may infer a southern basin-margin setting at Harvard Hill during the early Miocene. Our geochronology demonstrates that deposition of the Barstow Formation at Harvard Hill extended from before ~19.1 Ma until well after ~18.7 Ma, similar to timing of Barstow Formation lake deposition in the Calico Mountains but at least 3 million years older than comparable lacustrine facies in the Mud Hills type section. These observations are consistent with either of two paleogeographic models: westward transgression of lacustrine environments within a single large basin, or sequential development of geographically distinct eastern and western sub-basins.

Introduction

In our efforts to better understand the evolution of the Eastern California Shear Zone (ECSZ) in the central Mojave Desert, particularly interactions between NW-striking dextral and east-striking sinistral faults, we have realized the importance of fully understanding the effects of early Neogene tectonics. Harvard Hill, ~30 km east of Barstow, California, lies at the intersection between the ENE-striking sinistral Manix fault and NNW-striking dextral Dolores Lake fault (Fig. 1). Outcrop exposures at Harvard Hill mostly consist of lower-Miocene lacustrine strata that we consider to post-date the volcanoclastic Pickhandle Formation and correlate with the Barstow Formation (further discussed later). Two major observations based on our mapping at Harvard Hill are particu-

larly relevant for Neogene tectonics and paleogeography: (1) offset of Barstow Formation strata by a strand of the Manix fault; and, (2) the presence of the regionally extensive Peach Spring Tuff. Several workers (e.g., Dokka, 1989; Gans et al., 2005; Miller et al., this volume) have indicated that lacustrine deposition of Barstow Formation strata partly coincided with the most rapid extension for the nearby central Mojave metamorphic core complex. However, at least the upper part of the Barstow Formation has been thought to post-date extension (e.g., Fillmore, 1993; Fillmore and Walker, 1996) and pre-date the ECSZ. Similarly, the Peach Spring Tuff was emplaced subsequent to major tectonic extension in this area (Wells and Hillhouse, 1989). Therefore, the upper strata of the

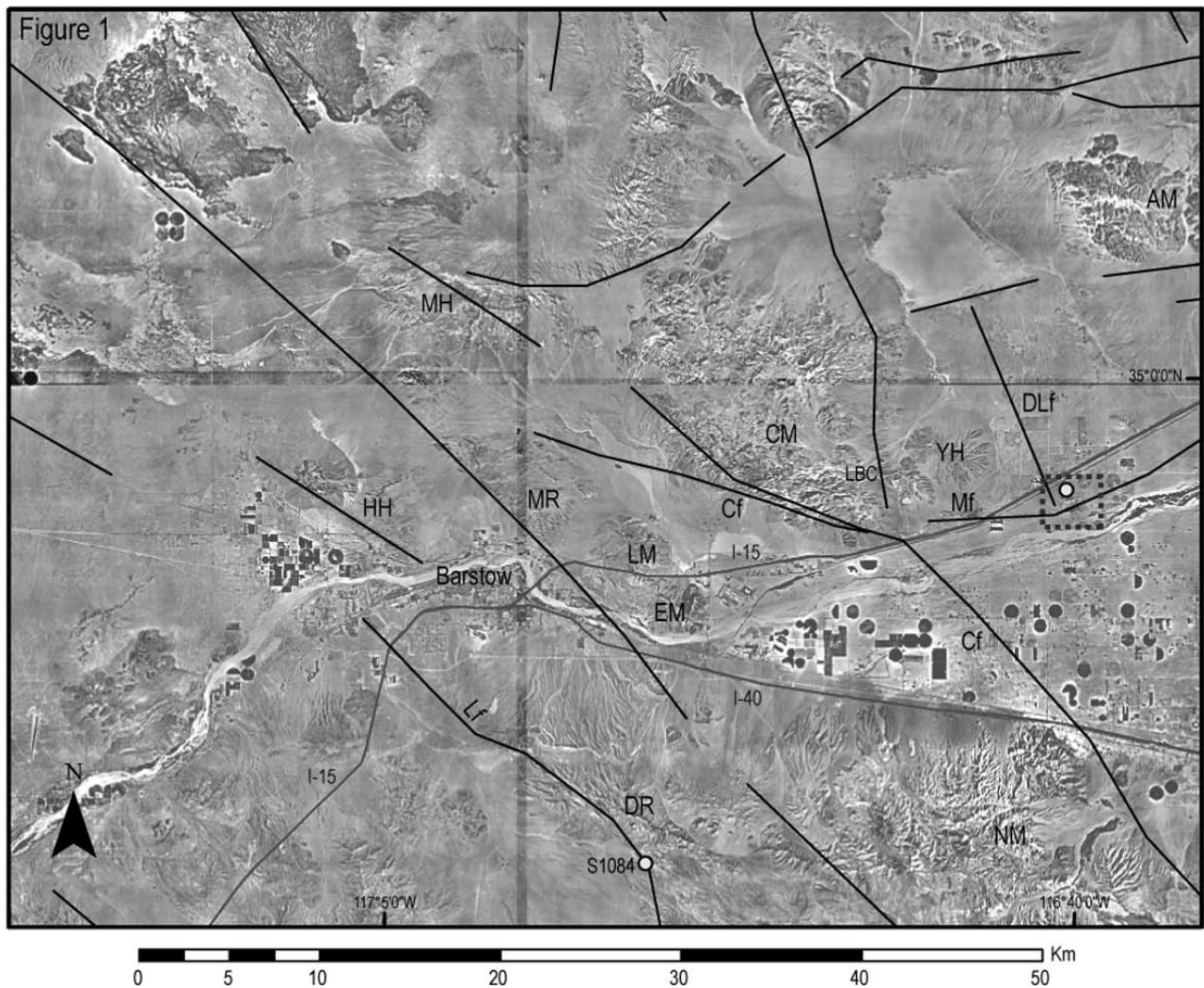


Figure 1. Location map of the central Mojave Desert region near Barstow, CA on DOQ base. Faults in black simplified from Miller et al., 2007. Harvard Hill study area delineated in dashed box. Circles indicate geochronology sample locations at Harvard Hill (in dashed box) and Stoddard Wash, S1084. Mf, Manix fault; DLf, Dolores Lake fault; Cf, Calico fault; Lf, Lenwood fault; NM, Newberry Mountains; DR, Daggett Ridge; EM, Elephant Mountain; LM, Lead Mountain; MR, Mitchel Range; HH, Hinkley Hills; MH, Mud Hills; CM, Calico Mountains; LBC, Little Borate Canyon; YH, Yermo Hills; AM, Alvord Mountain.

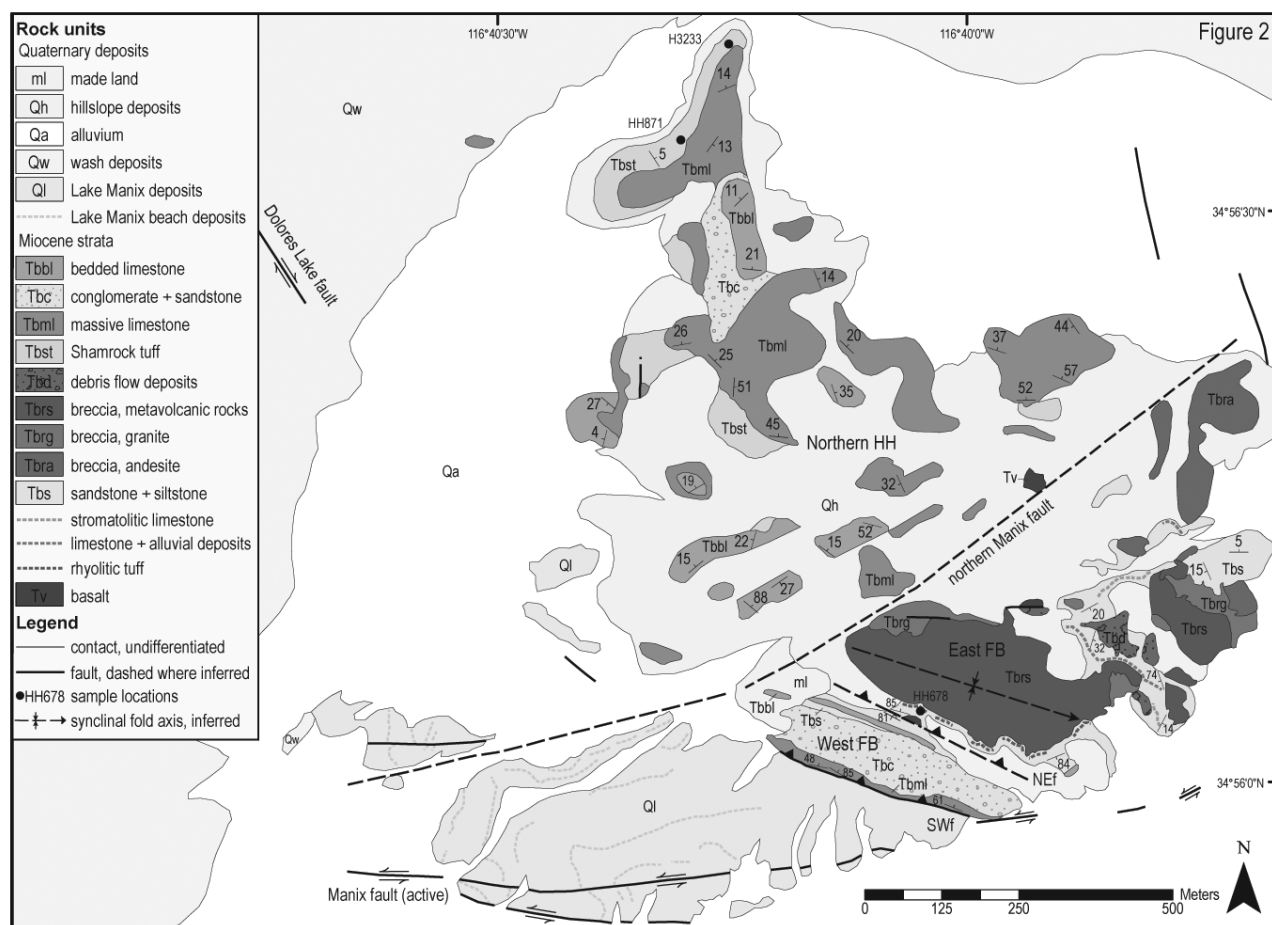


Figure 2. Geologic map of Harvard Hill, central Mojave Desert, CA. Major faults labeled: northern Manix fault; Manix fault (active); Dolores Lake fault; SWf, southwestern thrust fault; NEf, northeastern thrust fault. Northern HH (Harvard Hill), indicated, separated from southern HH by northern Manix fault; West FB (fault block) and East FB, indicated and separated by NEf. Note roughly NW-trending fold axis in Lake Manix beach deposits.

Barstow Formation and the Peach Spring Tuff serve as fundamental markers for unraveling ECSZ deformation in the central Mojave Desert.

Our study at Harvard Hill is part of a larger project of reconnaissance geologic mapping and chronologic studies of the Barstow Formation in several locations across the central Mojave Desert (see Miller et al., this volume). We have undertaken this work to better constrain the age of the Barstow Formation, characterize marker beds where they exist, and document local and regional facies changes within the strata. We have also documented the Peach Spring Tuff within Barstow Formation strata, establishing a well-characterized chronologic marker. This report provides early results of our geochronology and detailed geologic mapping of the Barstow Formation at Harvard Hill and provides preliminary interpretations of chronology, depositional environments, and paleogeography of the Barstow Formation and the Peach Spring Tuff.

Geologic setting

Harvard Hill is located about 30 km east of Barstow, California just south of Interstate 15 (Fig. 1). The ~1.5 x 1.5 km-wide hill is one of several tectonically uplifted blocks within a broad plain of Mojave River deposits. Outcrops at Harvard Hill consist of lower Miocene sedimentary strata and Pleistocene Lake Manix deposits (Fig. 2). This location lies near or within the transition from a domain of NW-striking dextral faults to the west from a domain of east-striking sinistral faults to the east as outlined by Miller and Yount (2002; Fig. 1). Particularly, Harvard Hill lies just northeast of the intersection between two presumably active faults: the ENE-striking sinistral Manix fault (Keaton and Keaton, 1977) and the NNW-striking dextral Dolores Lake fault (Meek, 1994; Fig. 1).

The greater Barstow area is underlain by several different types of pre-Miocene rocks that were uplifted and beveled by erosion during the early Cenozoic (Glazner et al., 2002), before Neogene tectonism. Miocene extension-

al tectonics characterized by NE-directed extension and accompanying NW-striking normal faults, basin development, and magmatism began about 24 Ma and continued to at least 19 Ma (e.g., Glazner et al., 2002); the cessation of extensional tectonics is a point of disagreement, with results ranging from 19 Ma to 17 Ma (Walker et al., 1990, 1995; Fillmore and Walker, 1996; Gans et al., 2005). Significant extension in the Waterman Hills, Mitchel Range, and Hinkley Hills region north and west of Barstow (Fig. 1) resulted in an exhumed core complex and synkinematic footwall plutons (Glazner et al., 1989, 2002). Northeast of the core complex, northwest-elongate basins formed in the hanging wall of an east-dipping detachment fault. Basinal sedimentary sequences are typically grouped as the lower volcanoclastic Pickhandle Formation (e.g., Fillmore and Walker, 1996) and the upper lacustrine and alluvial Barstow Formation (e.g., Woodburne, 1991). Lower Miocene strata have undergone vertical axis rotations (Ross et al., 1989; Hillhouse et al., this volume), either as small blocks rotating in the hanging wall of the detachment fault (Glazner et al., 2002) or by the rotation of regional crustal blocks (Dokka, 1989).

Beginning no earlier than 11 to 12 Ma (Schermer et al., 1996), strike-slip faulting of the ECSZ began across the region of former extension. Driven by approximately north-south maximum compressive stress, NW-striking dextral faults and conjugate NE-striking sinistral faults developed across the Barstow area (Fig. 1). In the west part of the area in Fig. 1, dextral faults partly occupied former normal faults (Singleton and Gans, 2008); whereas, to the east, sinistral faults formed. Studies of vertical-axis rotations of the ECSZ tectonic blocks show that blocks associated with sinistral faults have rotated as much as $\sim 45^\circ$ (Ross et al., 1989; Schermer et al., 1996; Hillhouse et al., this volume). A few tens of kilometers of dextral fault slip across the Mojave Desert has accommodated some of the strain associated with the ECSZ; both dextral and sinistral faults (i.e., the Manix fault) in the Barstow area are typically assigned total offsets of approximately 3 to 9 km. As a result, Miocene extensional basins are shuffled and rotated but only moderately displaced (see Hillhouse et al., this volume, for a summary of offsets).

Based on lithology and our chronology, we correlate the mostly lacustrine Miocene strata at Harvard Hill to the Barstow Formation, following Reynolds (2000, 2002). The Mud Hills type section of the Barstow Formation (Fig. 1) is well-dated from 19.3 to 13.4 Ma by ^{40}K - ^{40}Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating of six tuffs (MacFadden et al., 1990), and lithostratigraphy and biostratigraphy are also well-characterized (Woodburne et al., 1990). This section therefore provides the foundation for the Barstovian land mammal age. The Calico Mountains section

of the Barstow Formation (Fig. 1) is also well constrained chronologically from 19 to 16.9 Ma by $^{40}\text{Ar}/^{39}\text{Ar}$ techniques (Singleton and Gans, 2008), although fossil evidence suggests that beds may be as young as 14.8 Ma (Reynolds et al., this volume). Exposures of the Barstow Formation have also been observed at the Gravel Hills, Daggett Ridge, the Yermo Hills, and Alvord Mountain (Fig. 1), although lithofacies tend to vary (Woodburne, 1991). Reynolds (2004, pers. commun., 2010) suggested that regional correlations of the Barstow Formation may be possible using a sequence of marker beds including (in ascending order): massive stromatolitic limestone (MSL), brown platy limestones (BPL), and bed(s) rich in borates and strontium (SrB). He observed these marker strata in the Mud Hills, the Calico Mountains, Lead Mountain, Stoddard Valley, Daggett Ridge, the southern Cady Mountains, and Harvard Hill (Reynolds, 2004; Fig. 1). Uncertainties in these correlations have surfaced regarding whether the marker units are time-transgressive across the basin or if they are present in multiple separate basins but record very similar environmental conditions. The lack of chronologic constraints in several marker sequences has added to the ambiguity associated with basin timing and depositional environments.

In order to constrain the chronology of the Barstow Formation at Harvard Hill, we sampled two tuffs for age, one of which we interpret as a lacustrine facies of the regionally extensive Peach Spring Tuff (PST). The PST was first described by Young and Brennan (1974) near the western edge of the Colorado Plateau and several workers have since recognized the PST as far west as the Barstow area (e.g., Glazner et al., 1986; Wells and Hillhouse, 1989). The typical PST deposit is gray, pink, or dark purple and weakly to densely welded. Conspicuous sanidine with a distinctive blue schiller makes up 70–90% of the phenocrysts (Young and Brennan, 1974; Glazner et al., 1986); abundant sphene is also characteristic (Gusa et al., 1987). The PST at Stoddard Wash (S1084; Fig. 1), which was also sampled for geochronology, is a slightly welded pink tuff about 5 m thick that carries conspicuous sanidine and trace sphene and hornblende. The age of PST emplacement has been estimated at 18.5 Ma and 18.4 Ma (Nielson et al., 1990 and Miller et al., 1998, respectively) based on $^{40}\text{Ar}/^{39}\text{Ar}$ techniques (PST geochronology discussed later).

Methods

Detailed geologic mapping of Harvard Hill was conducted at a scale of 1:4000 to provide a foundation for understanding Miocene stratigraphy. In addition, four samples of three rhyolitic tuffs were dated by SIMS U-Pb geo-

chronology. Zircon grains were separated and analyzed by ion microprobe-reverse geometry (SHRIMP-RG) at the joint Stanford-USGS facility. Zircons were separated using standard procedures, mounted in epoxy, polished, and imaged by cathodoluminescence (CL). Probe spots for analyses were selected based on CL zoning; individual analyses used a 5-6 nA primary beam of O_2^- to sputter an approximately 15 to 30 micrometer pit, allowing for separate zones in zircon crystals to be analyzed. In addition to radiogenic daughter products of the U-Pb system, we collected data on several trace and rare-earth elements. Concentrations for U, Th and all of the measured trace elements are standardized against well-characterized, homogeneous zircon standard MAD, and U-Pb ages are referenced to zircon standard R33 (419 Ma; Black et al., 2004). Data were reduced using programs SQUID 1 and Isoplot (Ludwig, 2001, 2002).

Structural geology

In order to depict Barstow Formation stratigraphy at Harvard Hill, the structural context must be established. For convenience, we refer to three areas at Harvard Hill, each corresponding to a fault block discussed below: north, southwest, and southeast (Fig. 2). The primary structure at Harvard Hill is the Quaternary ENE-striking, sinistral Manix fault (Fig. 1, 2). We have mapped several strands of the Manix, including the presumably most active local strand along the south margin of Harvard Hill (Leslie et al., 2009; Fig. 2). This strand offsets earliest Holocene

alluvial fan deposits and Pleistocene Lake Manix deposits (Leslie et al., 2009; Reheis et al., 2007). We also document significant variations in Barstow Formation stratigraphy and chronology on either side of a poorly exposed northern Manix fault strand or strands (Fig. 2). Therefore, we divide our descriptions of Barstow Formation stratigraphy into northern and southern sections, separated by this northern Manix fault structure (Fig. 3). Although we have not been able to identify a surficial trace of this structure, mostly due to eolian sand cover, our inferred location for the northern Manix fault is north of the ENE-striking fault mapped by Dibblee and Bassett (1966). Another important structure that lies west of Harvard Hill and likely also affects Miocene stratigraphy is the NNW-striking, dextral Dolores Lake fault of Meek (1994; Fig. 1, 2). We have identified two strands of the Dolores Lake fault that are marked by scarps in latest Pleistocene to Holocene Mojave River fluvial plain deposits (Leslie et al., 2009).

Harvard Hill is located at the intersection between the Manix and Dolores Lake faults and outcrops here reveal complex deformation. At southwest Harvard Hill, adjacent to the intersection, compressional structures are observed; whereas, just northeast of Harvard Hill less than ~0.5 km from the apparent compressional regime, extensional graben are mapped (Leslie et al., 2009). The graben associated with extension just northeast of Harvard Hill are not known to affect exposed Miocene strata, therefore they are not described further in this study (for details see Leslie et al., 2009 and Miller et al., 2009). At

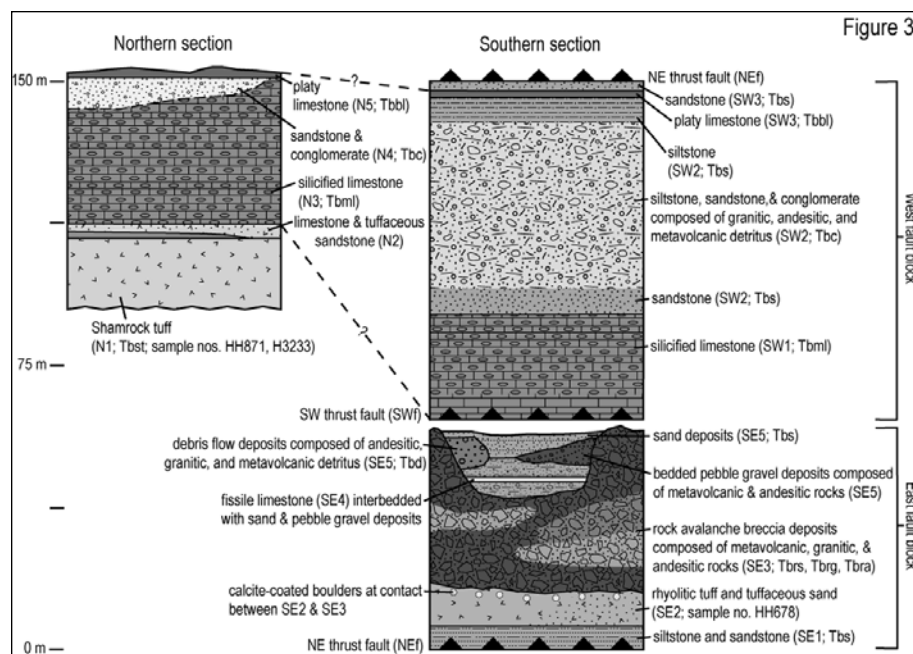


Figure 3. Stratigraphic columns for northern and southern Harvard Hill. Stratigraphic units from text, i.e., SE3, indicated. Geologic map (Fig. 2) units, i.e., Tbs, also indicated where applicable.

southwest Harvard Hill, Dibblee and Bassett (1966) mapped only conformable contacts and stratigraphy, but our mapping has identified two NW-striking and NE-dipping thrust faults (Fig. 2). The southwestern thrust fault (SWf) places subvertically tilted Miocene Barstow Formation strata over subhorizontal beds of Pleistocene Lake Manix beach deposits and NW-trending open folds occur in the footwall Lake Manix deposits (Leslie et al., 2009; Fig. 2). About 125 m to the northeast, we infer a second thrust fault (NEf) that places older Barstow Formation strata over younger strata of the same unit and SE-trending folds occur in the hanging wall strata (Fig. 2).

Bedding patterns are compatible with a possible tight NW-trending fold in the Miocene strata between the two thrust faults (Fig. 2), but tentative stratigraphic correlations (R.E. Reynolds, pers. commun., 2009) and top-to-the-NE stratigraphic indicators in the deposits there instead suggest a conformable sequence younging to the northeast (stratigraphy described in next section). Alternatively, we suggest that the bedded limestone (unit T_{bbl}; Fig. 2) has been overturned, possibly as a result of more complex faulting than indicated at map scale. In addition, complex bedding patterns at southern Harvard Hill, such as highly variable bedding orientations in lacustrine deposits (unit T_{bs}; Fig. 2) east of the northeastern thrust fault (NE_f) and south of the rock avalanche breccia deposits, suggest complicated faulting or beds disrupted by emplacement of the rock avalanche breccias. Moreover, variation in bedding attitudes at northern Harvard Hill (Fig. 2) may indicate more complex faulting and/or folding that we have not identified in our mapping.

The presence of the northeastern thrust fault (NE_f) is based on correlating the massive limestones (unit T_{bml}; Fig. 2) at northern Harvard Hill and at southwest Harvard Hill (R.E. Reynolds, pers. commun., 2009), and our geochronology (discussed later) showing that northeast-younging strata at southwest Harvard Hill are younger than strata at southeast Harvard Hill. Thus, a thrust fault (NE_f) is postulated to separate younger strata to the southwest from older strata to the northeast, and it parallels the southwestern thrust fault (SW_f; Fig. 2). Alternatively, if the massive limestone at southwest Harvard Hill does not correlate with that at northern Harvard Hill, but is older, the entire southern section would young eastward and the northeastern thrust fault (NE_f) would not exist. We consider the evidence to support the first interpretation that a thrust fault (NE_f) separates a southwest fault block from a southeast fault block (Fig. 3).

Miocene stratigraphy

Miocene strata at Harvard Hill consist primarily of limestone, sand and gravel deposits, and rock-avalanche breccias (Fig. 2, 3); due to the preponderance of lacustrine deposits at Harvard Hill, we consider these Miocene strata to be Barstow Formation rather than the volcanoclastic Pickhandle Formation (discussed later). Dibblee and Bassett (1966) recognized the following Tertiary units at Harvard Hill: limestone and claystone (“T_l”), fanglomerate (“T_f” and “T_{sf}”), sandstone and claystone (“T_s”), and andesite breccia (“T_{ab}”). We have further subdivided the strata (Fig. 2, 3).

North of the ENE-striking fault mapped by Dibblee and Bassett (1966), they indicate that “T_l” is conform-

able on “T_{ab}”; south of the fault, they interpret the section to be conformable, younging to the southwest. The apparently oldest unit of southern Harvard Hill is a fanglomerate of Mesozoic detritus (“T_f”) consisting of mostly metavolcanic rocks, as well as fine-grained schist, pink granite, and Tertiary andesite (Dibblee and Bassett, 1966). We recognize a similar assemblage of rocks at southeast Harvard Hill, but we further separate this unit into several units based on distinct lithology, different modes of deposition, and recognition that a thrust fault (NE_f) disrupts the section. Moreover, we suggest that the andesite breccia (“T_{ab}”) north of the fault mapped by Dibblee and Bassett (1966) is part of the southeast fault block; this interpretation has led to our mapping of a northern Manix fault structure farther north than that of Dibblee and Bassett (1966; Fig. 2).

Reynolds (2002) noted early Miocene marker beds in the Barstow Formation at northern Harvard Hill, including (in ascending order): “granitic sediments”, “green tuffs”, “massive stromatolite limestone”, and “brown platy limestone”. We have mapped similar units, although we do not observe granitic sediments and multiple green tuffs; we have mapped one thick tuff stratigraphically below limestone at northern Harvard Hill (Fig. 2). Although Reynolds (2002) correlates marker units at Harvard Hill with those at Daggett Ridge, the Calico Mountains, and the Mud Hills type section (Fig. 1), the ages of the marker beds are poorly constrained. Our detailed stratigraphic and geochronologic analyses may help to further constrain Barstow Formation correlations across the Miocene basin.

Northern section

Exposed strata of the Barstow Formation at northern Harvard Hill are ~40 to 50 m thick. The section consists of the following conformable units (in ascending order; Fig. 3): (N1) a thick rhyolitic tuff; (N2) thin- to upwardly thick-bedded limestone with variable silicification, locally interbedded with tuffaceous lacustrine sandstone; (N3) massive silicified limestone with stromatolites; (N4) well-sorted alluvial sandstone to pebble conglomerate interbedded with lacustrine sandstone and siltstone; and, (N5) fossiliferous platy limestone.

N1: A rhyolitic tuff, here informally named the Shamrock tuff, is the lowest exposed stratigraphic unit of the Barstow Formation at northern Harvard Hill. This unit may be related to one of the green tuffs recognized by Reynolds (2002). It is at least 9 m thick and may be up to 17 m thick, although this larger thickness is tentative as it was measured in a partly covered, structurally complex area. The tuff is a celadonite-altered, light lime green color and nonwelded. It is mostly massive except near its up-

per contact, where it is locally thin-bedded, silicified, and a paler green. The Shamrock tuff contains pumice and conspicuous megascopic phenocrysts of sanidine, plagioclase, biotite, and sphene in an altered, fine-grained matrix. Lithics are small and inconspicuous, but are more common near the lowest exposures and upper contact, within the bedded section.

N2: Conformably overlying the Shamrock tuff is variably bedded and silicified limestone, locally interbedded with poorly indurated lacustrine sandstone. A bed of white to light gray limestone with dark gray to black chert nodules marks the lower contact with the tuff; it ranges in thickness from ~1 m to 0 where it pinches out. Locally above this limestone bed and above the Shamrock tuff where the limestone pinches out is green tuffaceous sandstone, interpreted here as a reworked lacustrine deposit of the Shamrock tuff. At the northernmost part of Harvard Hill, these tuffaceous sand deposits occur in multiple beds within the limestone, presumably intertonguing with the limestone and representing multiple local transgression and regression events. Where present, the tuffaceous sandstone laterally varies in thickness from 15 cm to ~3 m. In some locations, green well-sorted lacustrine sandstone, lacking tuffaceous minerals, is interbedded with the limestone. Stratigraphically higher is white, gray, and tan thin- to thick-bedded limestone with dark gray, black, and brown silicified nodules, lenses, and beds (Fig. 4A). This limestone notably varies laterally both in silicification and thickness, indicating facies variation within just a few hundred meters.

N3: Bedding thickens upsection, grading into massive, light- to dark-gray stromatolitic limestone up to 30 m thick. Stromatolites are flat to domal lamellar structures and typically up to ~75 cm in diameter (Fig. 4B). This unit is largely silicified and forms rugged ridges at northern Harvard Hill. It may correlate with the massive stromatolite limestone described by Reynolds (2002).

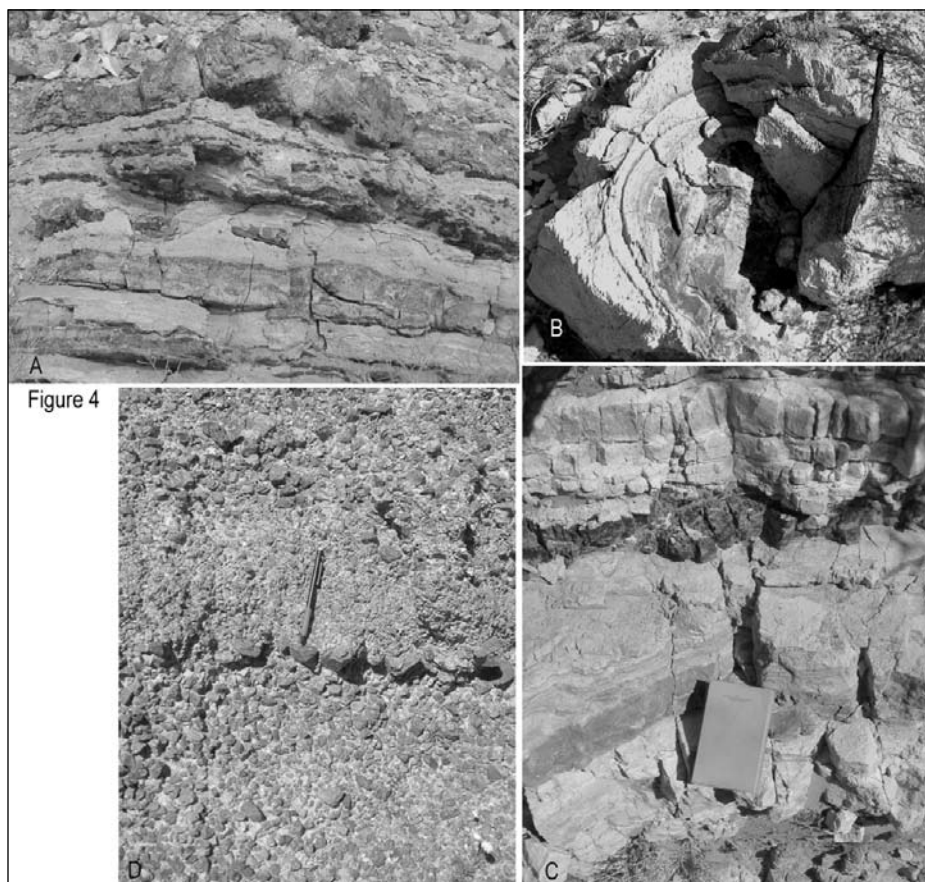


Figure 4. Field photographs illustrating Barstow Formation lithofacies at Harvard Hill. (A) Unit N2; white, gray, and tan thin- to thick-bedded limestone with dark gray, black, and brown silicified nodules, lenses, and beds. May correlate with unit SW1 in photo (C). (B) Large flat to domal lamellar stromatolite common in unit N3. (C) Unit SW1; tan, brown, and light gray thin- to thick-bedded limestone with dark brown silicified beds. May correlate with unit N2 in photo (A). (D) Unit SE5; bedded alluvial gravel deposits.

N4: Although exposures are poor above the limestone, well-sorted, coarse-grained alluvial sandstone to pebble conglomerate is locally present. It mostly consists of sub-rounded to sub-angular feldspar and quartz grains and basalt and metavolcanic lithic fragments in a white fine-grained tuffaceous matrix and it is interbedded with well-sorted lacustrine sandstone and siltstone. This N4 unit is up to ~6 m thick and appears to pinch out southward; in particular, it is absent at a key outcrop in the west-central part of northern Harvard Hill where massive limestone (unit N3) is directly overlain by platy limestone (unit N5; Fig. 2).

N5: The stratigraphically highest unit of the Barstow Formation at northern Harvard Hill is locally exposed fetid, fossil-rich, platy limestone that is less than 0.5 m thick. It is very well-bedded with bedding thickness ranging from a few millimeters to ~5 cm and there are wavy (mm-scale) undulations in the bedding. This limestone is tan, gray, white, light orange-yellow, and dark brown; the dark brown beds tend to be highly silicified and especially

rich in needle-shaped water reed fossils (R.E. Reynolds, pers. commun., 2009). This thin-bedded limestone may correlate with the brown platy limestone described by Reynolds (2002).

Southern section

The southern section of the Barstow Formation is divided into a west fault block of mostly lacustrine strata, possibly correlative with strata at northern Harvard Hill, and the east fault block of older deposits, primarily rock-avalanche breccia (Fig. 2).

West fault block

In the west fault block exposed at southern Harvard Hill, strata are exposed in a steeply dipping section ~100 m thick. This section consists of the following units (in ascending order; Fig. 3): (SW1) variably silicified well-bedded to massive limestone; (SW2) lacustrine siltstone and sandstone interbedded with alluvial siltstone, sandstone, and conglomerate; and, (SW3) platy limestone.

SW1: The lowest exposed stratigraphic unit above the southwestern thrust fault (SWf) is tan, brown, and light gray thin- to thick-bedded limestone with dark brown silicified beds (Fig. 4C). This well-bedded limestone is locally entirely truncated by the fault and it is 4 m thick where most completely exposed. Upward, the bedded limestone grades into a ~14-m-thick massive limestone. This massive limestone is white to dark gray, locally stromatolitic, largely silicified, and shapes a NW-trending ridge that culminates southwestern Harvard Hill (Fig. 2). This unit is similar to and may correlate with unit N3 at northern Harvard Hill (Fig. 2, 3) and the massive stromatolite limestone of Reynolds (2002). The stromatolites here are also analogous to those of unit N3, i.e., large, flat to domal lamellar structures; they indicate a top-to-the-NE stratigraphic relationship (R.E. Reynolds, personal commun., 2009). Because the base of this section is truncated by a fault (SWf), we have not confirmed the Shamrock tuff beneath this limestone.

SW2: Conformably overlying the massive limestone is green lacustrine siltstone and sandstone interbedded with tan, brown, and pink alluvial siltstone, sandstone, and conglomerate. The lacustrine deposits contain fissile siltstone beds and beds of well-sorted fine-grained sandstone with rounded grains. A mine pit at southwest Harvard Hill is in a thick deposit (up to ~12 m) of well-bedded lacustrine siltstone that pinches out to the southeast (Fig. 2). The interbedded alluvial deposits, up to 80 m thick, contain a wide grain-size range, from silt to cobble, of mostly granitic lithic fragments in addition to metavolcanic and andesitic volcanic clasts in a red fine-grained arkosic sand matrix. Clasts are angular to subrounded, matrix-supported, and poorly sorted. Beds are lenticular

and contain fining upward sequences toward the north-east.

SW3: Stratigraphically higher is white, gray, tan, and brown platy limestone with dark brown to black silicified beds. Locally, bedding is lenticular and stromatolites are present. This distinctive limestone is less than 75 cm thick and poorly exposed. It is overlain by a ~5-m-thick sequence of green well-sorted lacustrine sandstone. The limestone may correlate with unit N5 of northern Harvard Hill and the brown platy limestone of Reynolds (2002), but poor exposures of both units create uncertainty in stratigraphic correlations. The section is truncated to the northeast by the poorly exposed northeastern thrust fault (NEf; Fig. 2).

East fault block

In the east fault block of southern Harvard Hill, Barstow Formation strata are at least 47 m thick. Here, the strata consist of the following units (in ascending order; Fig. 3): (SE1) lacustrine siltstone and sandstone; (SE2) a biotite-rich rhyolitic tuff; (SE3) rock-avalanche breccia deposits; (SE4) stromatolitic limestone; and, (SES) alluvial sand and gravel deposits.

SE1: The lowest exposed stratigraphic unit of the Barstow Formation at Harvard Hill consists of thin-bedded lacustrine siltstone and sandstone. It consists of pale green fissile siltstone beds and green well-sorted fine- to medium-grained sandstone with rounded grains. This unit is truncated to the southwest by a thrust fault (NEf; Fig. 2).

SE2: The lacustrine deposits of unit SE1 grade upward into a white medium-grained rhyolitic tuff. The tuff contains quartz, plagioclase, potassic feldspar, and biotite phenocrysts and abundant lithic grains. This unit varies in thickness, from ~13 m to 0, and grades laterally into tuffaceous lacustrine sand deposits.

SE3: Rock-avalanche breccia deposits overlie the tuff and lacustrine sand deposits of unit SE2. Our observations at or adjacent to the contact between unit SE2 and the overlying rock-avalanche breccia deposits indicate a depositional relationship. Locally, the upper ~10-20 cm of the tuff (unit SE2) is silicified adjacent to the contact. Where the tuff pinches out, angular pebble-sized clasts consisting of the breccia material are locally present adjacent to the uppermost exposed tuffaceous lacustrine sand deposits (SE2). Calcite-supported clasts and calcite-coated boulders consisting of breccia material are also locally observed at this contact.

The breccia deposits are composed of a suite of monolithologic blocks of metavolcanic, granitic, and andesitic rocks. We interpret these breccias as rock-avalanche deposits because of the lack of matrix material and the

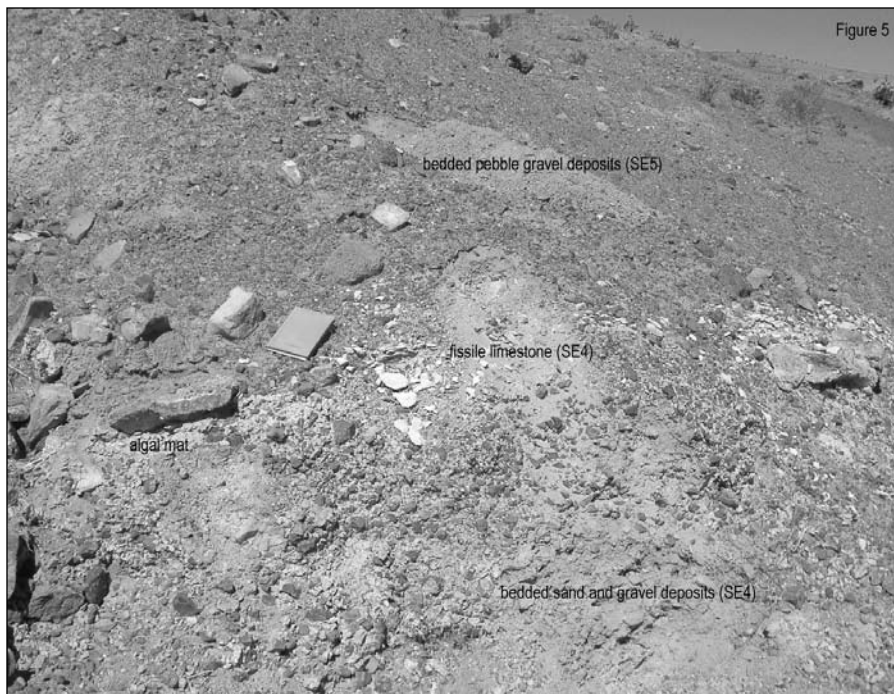


Figure 5. Field photograph illustrating thin fissile cream-colored limestone interbedded in sand and gravel deposits (SE4), and overlying bedded pebble gravel deposits (SE5). The limestone locally contains flat algal mats capped by gravel-rich limestone providing stratigraphic "up" indicators.

preservation of internal, relict stratigraphy in each of the monolithologic blocks over hundreds of meters (K. Bishop, pers. commun., 2009; Yarnold and Lombard, 1989); although pervasively fractured, the fracture-bounded fragments are large, up to tens of meters, and show no discernable rotation relative to one another (Yarnold and Lombard, 1989).

The largest monolithologic rock-avalanche breccia block is composed of mafic to intermediate metavolcanic rocks. This ~40-m-thick block consists of aphanitic metabasalt and porphyritic metaandesite locally with phyllitic to schistose texture. The aphanitic metabasalt is dark gray to black, fine-grained to microcrystalline with rare phenocrysts of plagioclase, and locally shows relict flow banding. The porphyritic metaandesite is dark gray, commonly contains abundant phenocrysts of white euhedral to subhedral plagioclase, and rarely contains biotite clusters. Locally the plagioclase phenocrysts of

the metaandesite are larger, less abundant, pink, and anhedral. We tentatively correlate these metavolcanic rocks with those of the Jurassic Sidewinder Volcanics (e.g., Schermer and Busby, 1994). The brecciated granitic block is composed of pink-gray medium-grained equigranular leucocratic granite. It contains brown smoky quartz, pink-white potassic feldspar, white plagioclase, and biotite. Locally, this granitic breccia is matrix-supported with poorly sorted, rounded granitic pebbles to boulders in a red arkosic sand matrix. The andesitic rock-avalanche breccia block is characterized by a red, brown, and purple-gray vesicular porphyritic andesite, locally with flow banding. It contains biotite, quartz, and white equant plagioclase phenocrysts in a glassy, fine-grained to microcrystalline matrix.

SE4: Cream-colored limestone interbedded in green lacustrine sand and gravel deposits occurs stratigraphically above the rock-avalanche breccia deposits (Fig. 5). The poorly sorted sandy to gravelly matrix-supported limestone is commonly bedded and fissile, although locally massive; it is ~3 to 30 cm thick and contains stroma-

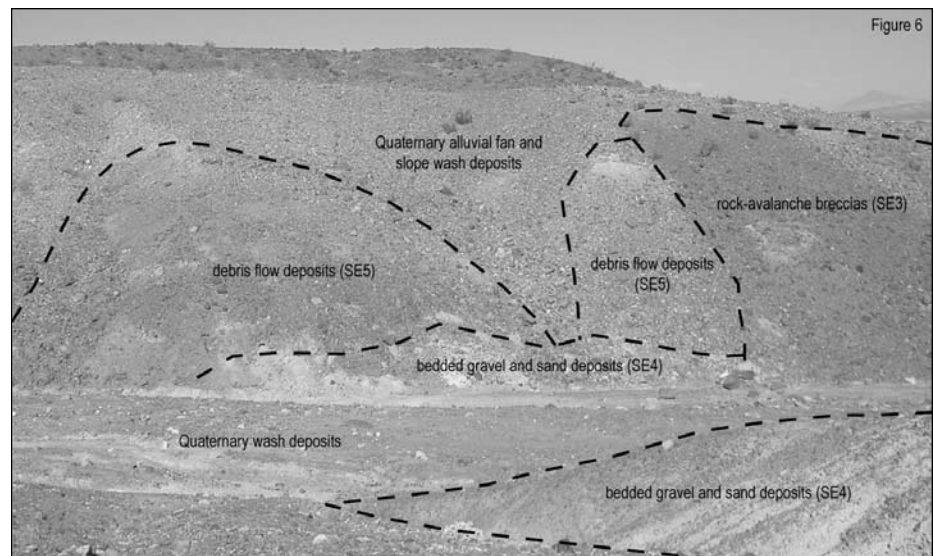


Figure 6. Field photograph showing units SE4 and SE5 lying in erosional unconformity against subjacent rock avalanche breccia, illustrating steep local paleotopography. Bedding in the SE4 gravel and sand deposits is sub-horizontal and fans out from the subvertical contact against the breccia. Debris flow deposits (SE5) lie on the SE4 bedded deposits and abut against the steep contact on breccia.

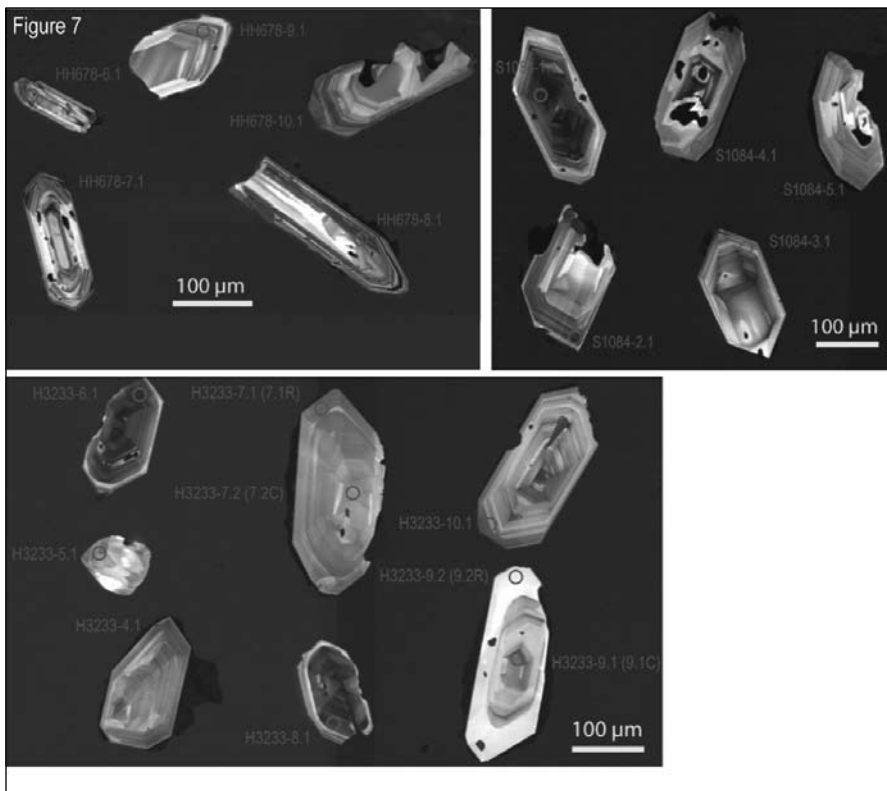


Figure 7. Cathodoluminescence images of HH678, H3233, and S1084 zircons. Circles show approximate spot size and location.

tolites and fossil *Physa* sp. aquatic snails (R.E. Reynolds, pers. commun., 2009). Stromatolites here differ from the stromatolites elsewhere at Harvard Hill (units N3 and SW1) in that they are smaller (up to 10 cm in diameter), tubular to mound shaped, and not silicified. The gravel found locally in the limestone and in the interbedded lacustrine gravel deposits is composed of angular to subangular metavolcanic fragments, most likely derived from the rock-avalanche breccia deposits stratigraphically below it. The units stratigraphically above the breccia deposits are sub-horizontal and fan out from the sub-vertical buttress contact with the breccias (Fig. 6). This relationship indicates relatively steep local paleotopography following rock-avalanche breccia deposition. The stromatolites, locally observed as distinct flat algal mats and rarely silicified, are occasionally capped by gravel-rich limestone providing a stratigraphic “up” indicator (Fig. 5). The interbedded lacustrine deposits, up to ~15 m thick, consist of green, well-sorted sand with rounded grains and well-sorted, matrix-supported fine gravel in a green silty sand matrix. A coarsening upward sequence is locally observed in these lacustrine deposits just below the stromatolitic limestone.

SE5: Stratigraphically higher than the fissile stromatolitic limestone is a sequence of alluvial deposits interbedded and intertonguing with lacustrine sand deposits.

These alluvial facies consist of bedded pebble gravel deposits and a poorly sorted debris flow pebble to cobble gravel deposit. The bedded gravel deposits are poorly to well sorted, matrix to clast supported, and locally massive or lenticular bedded (Fig. 4D). They are composed of subangular to angular metavolcanic (>90%) and andesitic (<5-10%) clasts in a weakly indurated, fine-grained sand (and locally calcite-cemented) matrix. We interpret these gravel deposits as near shore alluvial facies intertonguing with and grading into lacustrine sand facies, apparently abutted against the steep paleotopographic slope of the rock-avalanche breccia deposits (Fig. 6). We also observe a general fining upward sequence within these gravel deposits.

The largely matrix supported debris flow deposit consists of mostly cobble-sized andesite

(90-95%), granite, and metavolcanic angular to rounded rock fragments in a red silty fine-grained sand matrix. This deposit is commonly massive, although locally crudely bedded, perhaps displaying inverse grading, and moderately sorted where finer grained (sand to pebble gravel). The clast lithologies of the deposit correspond to the assemblage of rock types in the rock-avalanche breccia suite. Near the base of the debris flow deposit, where it is depositional on lacustrine sand deposits of unit SE4, the matrix is composed of green fine- to medium-grained sand with rounded grains; we interpret the matrix here to be of lacustrine origin, suggesting that the debris flow was deposited in a lacustrine environment.

The highest Barstow Formation strata in the east fault block are green well-sorted lacustrine sand deposits, just above the alluvial facies and possibly gradational with the bedded pebble gravel deposits. However, Pleistocene Lake Manix deposits, also green and well-sorted sand with rounded grains, overlie Barstow Formation lacustrine sand deposits, creating some ambiguity in interpreting the stratigraphic position of many outcrops of green lacustrine sand.

In three separate areas at Harvard Hill, we observe calcite amygdaloidal basalt (Fig. 2). In all localities, the basalt is highly weathered and plagioclase phenocrysts are locally preserved. The basalt’s common presence in or

adjacent to fault zones suggests to us that it may occur as fault slivers of some unit unrelated to the Barstow Formation.

Geochronology and geochemistry

We present data from ^{207}Pb -corrected $^{206}\text{Pb}/^{238}\text{U}$ analyses of multiple zircon grains from four samples in probability distribution plots; the histograms show the frequency of each age within a given bin value while the relative probability curves illustrate primary age populations as peaks (Fig. 8). The probability curve takes uncertainty in each analysis into account and therefore represents the zircon age data in a more statistically reliable fashion than the histogram. We also present $^{206}\text{Pb}/^{238}\text{U}$ age vs. Th/U graphs in order to demonstrate geochemical variance per sample (Fig. 9).

Zircons from HH678, a reworked rhyolitic tuff at southeast Harvard Hill (Fig. 2, 3), are euhedral to subhe-

dral, prismatic to needle-like (3-6:1), and display oscillatory zoning, with several grains also displaying complex zonation (Fig. 7). Many grains display resorbed anhedral interiors bounded by euhedral rims (Fig. 7). Grains range in size from approximately 100 to 325 μm measured across the longest axis of the grain. Out of 25 zircon grain analyses for HH678, 19 are represented in a probability distribution graph (Fig. 8). The 19 analyses range in age from 18.9 to 21.3 Ma and yield a weighted mean age of 19.9 ± 0.3 Ma (MSWD = 7.2). The histogram shows two moderate age modes at approximately 19.2 and 19.7 Ma and a more prominent mode at ~ 20.4 Ma. The probability distribution curve demonstrates at least two peak probability ages at ~ 19.1 and 20.3 Ma, with a minor hump indicated at ~ 19.6 Ma. Six older detrital grains are excluded from this interpretation, but they present ages that may have relevance for interpreting detrital systems (discussed later): 48, 145, 147, 148, 151, and 217 Ma.

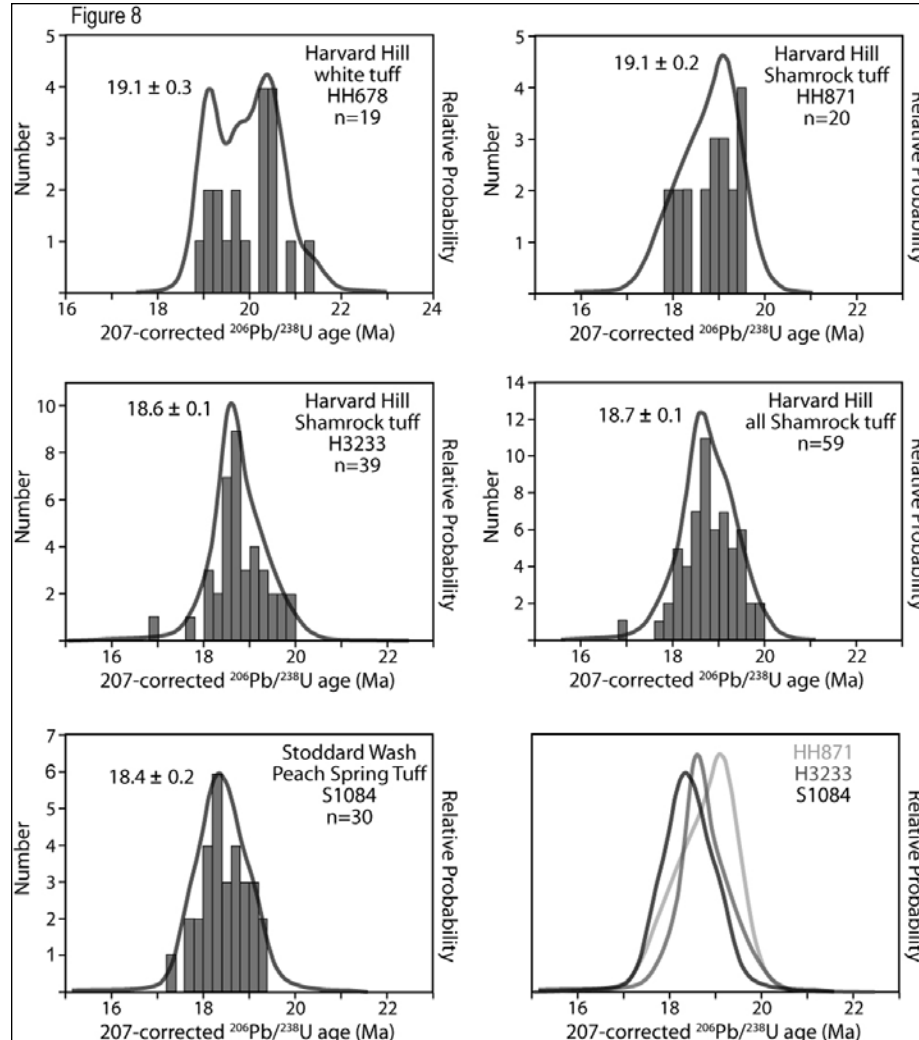


Figure 8. Probability distribution plots of $^{206}\text{Pb}/^{238}\text{U}$ zircon ages for HH678, HH871, H3233, and S1084, with youngest peak probability value in Ma indicated. Combined plot showing probability density curves for HH871, H3233, and S1084 also shown. Older detrital grains not shown.

Zircons from two samples of the Shamrock tuff, HH871 and H3233, were analyzed in this study. HH871 was collected near the top of the Shamrock tuff section in a reworked, distinctly bedded part, whereas H3233 was collected near the base of the exposed section, about 7 m below its upper contact, in massive altered tuff (Fig. 2, 3). Zircons from the Shamrock tuff (HH871, H3233) are euhedral to subhedral, prismatic 2-3:1, and primarily display oscillatory zoning, with some grains also displaying sector and/or complex zoning (Fig. 7). Melt inclusions are common in the zircons. A few grains display resorbed anhedral cathodoluminescent (CL)-dark cores, rich in U and Th, surrounded by euhedral oscillatory-zoned rims (Fig. 7). Grains range in size from approximately 100 to 300 μm measured across the longest axis of the grain. Twenty out of the 23 zircon grain analyses for HH871 are represented in a probability distribution plot (Fig. 8). The 20 grain analyses range in age from 17.8 to 19.5 Ma and yield a weighted mean

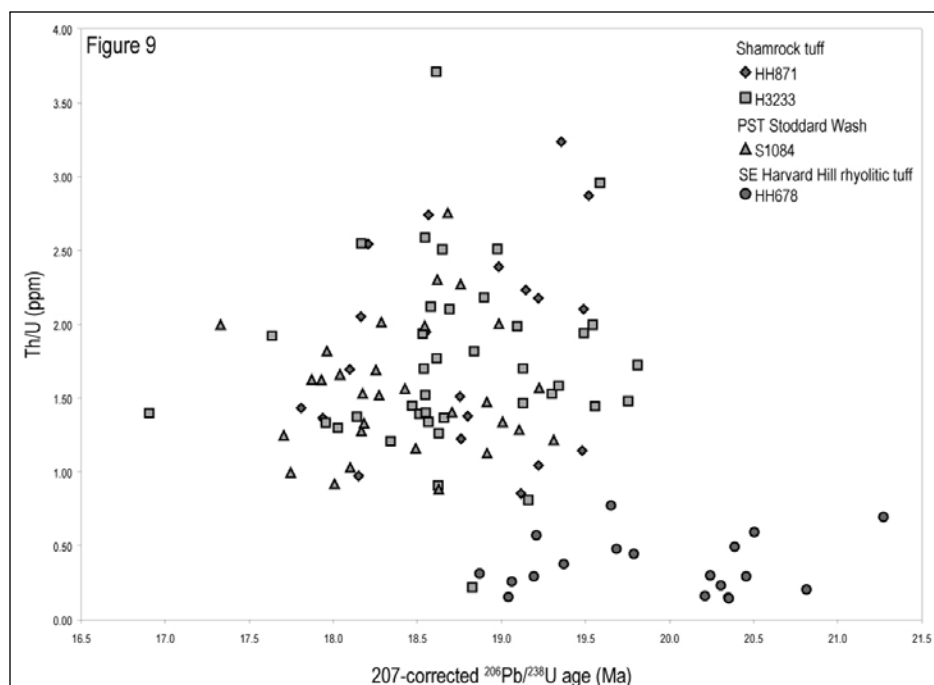


Figure 9. Plot of Th/U vs. $^{206}\text{Pb}/^{238}\text{U}$ age for zircon grains dated. Note the relatively low values for southeast Harvard Hill rhyolitic tuff HH678 and higher values for Shamrock tuff HH871 and H3233, and Stoddard Wash Peach Spring Tuff S1084. Older detrital grains not shown.

age of 18.9 ± 0.2 Ma (MSWD = 2.3). The histogram illustrates three age modes at about 18.1, 19.0, and 19.5 Ma; the modes decrease in frequency, yet taper in width with increasing age. The probability distribution curve indicates a peak probability age value of approximately 19.1 Ma, with a tail skewed to younger ages. The 3 grains excluded from this interpretation include 2 older detrital grains (248 and 1662 Ma) and 1 grain contaminated by common Pb.

All of the 39 zircon grain analyses for H3233 are displayed in a probability distribution plot (Fig. 8) and yield a single peak probability age of ~ 18.6 Ma. Older detrital zircon grains were not present in our analyzed group of grains. Zircon grain analyses range in age from 16.9 to 19.8 Ma and define a weighted mean age of 18.7 ± 0.1 Ma (MSWD = 2.1). The histogram illustrates two modest modes at about 18.1 and 19.1 Ma; a third mode at ~ 18.6 Ma is much more prominent. Nine interior-rim pairs were analyzed for single zircons from H3233. Interiors and rims most commonly yielded the same age, within (1σ) error. Three grains reveal conspicuously older interiors, by ~ 0.5 to 1.5 Ma. Interiors from these grains are both CL-dark (U- and Th-rich), resorbed cores and oscillatory-zoned interiors (Fig. 7). When zircon grain analyses from the two Shamrock tuff samples (HH871, H3233) are combined ($n = 59$), the interpreted peak probability age value is approximately 18.7 Ma, with the curve slightly skewed to both younger and older ages (Fig.

8). The combined analyses define a weighted mean age of 18.8 ± 0.1 Ma (MSWD = 2.2). The histogram demonstrates a major age mode at ~ 18.7 Ma, with a substantial quantity of both older and younger grains.

Because of the distinctive age, mineralogy, outcrop thickness, and remanent magnetization (Hillhouse et al., this volume) of the Shamrock tuff, we permissively correlated it with the Peach Spring Tuff. In order to test this possible correlation, we sampled and conducted zircon U-Pb age analyses on the well-characterized Peach Spring Tuff at Stoddard Wash (Wells and Hillhouse, 1989; S1084; Fig. 1). Zircons from S1084 are similar to those of the Shamrock tuff (HH871, H3233) in CL: typically euhedral,

prismatic 2-3:1, commonly contain melt inclusions, and display oscillatory zoning, with few grains also displaying sector and/or complex zoning (Fig. 7). Grains range in size from approximately 200 to 300 μm measured across the longest axis of the grain. All of the 29 zircon grain analyses for S1084 are represented in a probability distribution plot (Fig. 8). Similar to H3233, older detrital zircon grains were not present in our analyzed group of grains. The 29 analyses range in age from 17.3 to 19.3 Ma and yield a weighted mean age of 18.4 ± 0.2 Ma (MSWD = 2.3). Although only one primary age mode is represented in the histogram at ~ 18.3 Ma, several grains are both younger and older than this value. The peak probability age is similar, at about 18.4 Ma.

Geochemical analyses were also conducted on zircons from the three Harvard Hill tuff samples (HH678, HH871, H3233) and the Peach Spring Tuff sample at Stoddard Wash (S1084). Zircons from the Shamrock tuff (HH871, H3233) are characterized by Th/U values from 0.8 to greater than 3.0, whereas zircons from the older tuff from southern Harvard Hill (HH678) yield values from 0.8 to 0.1 (Fig. 9). In addition, zircons from the Peach Spring Tuff at Stoddard Wash (S1084) yield the conspicuously high Th/U (values 0.8 to 2.8) observed in zircons from the Shamrock tuff (Fig. 9), strengthening the correlation of the Shamrock tuff and the Peach Spring Tuff.

Discussion: geochronology and geochemistry

Interpretation of probability distribution plots

Our ^{207}Pb -corrected $^{206}\text{Pb}/^{238}\text{U}$ analyses of zircons from each sample are presented in probability distribution plots, where the curves statistically best represent the zircon age data (Fig. 8). Typically, our probability distribution curves display at least one prominent age mode with a tail skewed to older ages (Fig. 8). Because older zircon grains can be introduced to the juvenile population of grains by several magmatic and post-eruption depositional processes (e.g., Miller et al., 2007; Miller et al., this volume), a skewed older tail or even older mode(s) is expected. Consequently, though the histogram and probability distribution curve for HH678 shows two to three age modes (Fig. 8), we interpret the youngest histogram mode and probability curve peak, 19.1 ± 0.3 Ma, as the best approximation for the age of zircon crystallization that is most closely associated with the eruption of the tuff, with the older mode(s) representing older magmatic and/or detrital zircons. Though parts of grains, i.e. interiors, also may contribute to an older age distribution, only three out of nine interior-rim pairs that were analyzed from H3233 reveal distinctly older interiors than their rim counterparts; interiors and rims commonly yielded the same age. Although the eruption age of a tuff may be slightly younger than the (youngest) zircon peak probability age value, estimating an eruption age somewhere along the younger slope of the probability curve is arbitrary (e.g., Simon et al., 2008). At this time we interpret the youngest zircon peak probability value as the best approximation of the eruption age of a tuff (Miller et al., this volume).

Although we expect to observe a skewed older tail or mode of ages displayed in the probability distribution plots, indicating detrital and/or older magmatic zircons as discussed above, a younger distribution of grains is more puzzling. Our most extensive zircon grain analysis ($n = 39$) was conducted on H3233 to test how significant the younger population of ages is for the final interpretation. Although the histogram shows at least one modest mode younger than the peak probability age of ~ 18.6 Ma, zircon grains with age values ≤ 18.4 Ma ($n = 7$) have an average 1σ error of ± 0.5 Ma; this is much greater than the average error of ± 0.2 Ma associated with 18.6 Ma values ($n = 7$). As a result of the larger uncertainty associated with these younger grains, the probability distribution curve does not indicate a significant probability that the zircon population is that young. These younger grains may represent a post-eruptive process such as minor lead loss that we cannot evaluate with existing data, or more importantly, they may indicate a younger eruption age than that estimated

by the probability distribution curve peak(s). Nevertheless, because of the larger errors associated with the younger population of grains, we infer that the peak probability value, 18.6 ± 0.1 Ma, best represents the dominant age of zircon crystallization in sample H3233 and the age most closely associated with eruption.

The probability distribution curve for HH871 is unique in that it yields an older peak zircon age (19.1 ± 0.2 Ma) relative to the other Shamrock tuff sample (H3233, 18.6 ± 0.1 Ma) and the combined PST samples (18.7 ± 0.1 Ma); in addition, both the histogram and probability curve display a scattered, though significant population of younger grains (Fig. 8). In fact, the histogram reports that 12 of 20 analyzed grains are younger than the ~ 19.1 Ma peak probability value, suggesting that the peak value may not best represent the age of eruption, i.e., the youngest significant population of zircon grains. Age values between 19.0 and 19.2 Ma ($n = 5$) have an average 1σ error of ± 0.3 Ma, while age values ≤ 18.8 Ma ($n = 11$) have a slightly greater average error of ± 0.4 Ma. As discussed above, these younger grains may represent a younger eruption age or post-eruptive lead loss. Alternatively, because we have a large data set for HH3233 ($n = 39$) that indicates a peak zircon age of 18.7 ± 0.1 Ma, we interpret the probability age distribution of HH871 ($n = 20$) to be less constrained due to a smaller data set, and, therefore a non-representative subset of the total population of analyzed zircon grains. Furthermore, the similarity between H3233 and the combined Shamrock (H3233 and HH871) probability distribution plots provides confidence that the combined probability distribution peak (Fig. 8), 18.7 ± 0.1 Ma, better represents the eruption age of the Shamrock tuff.

The zircon peak probability age for the PST at Stoddard Wash (S1084) is 18.4 ± 0.2 Ma, slightly younger than that for the Shamrock tuff. However, the probability distribution curves for each sample are similar (Fig. 8) and the peak probability values for the Shamrock tuff and PST at Stoddard Wash are within 1σ error (18.7 ± 0.1 Ma and 18.4 ± 0.2 Ma, respectively). The young age distribution of S1084 ($n = 30$) also may be less well-constrained due to a smaller data set and may not statistically represent the total population of analyzed zircon grains. Thus, although there is variability in smaller contributing zircon age data sets, we assume that the combined Shamrock tuff peak probability age (Fig. 8), 18.7 ± 0.1 Ma, is: (1) best represented by the largest single sample (H3233, $n = 39$) and, (2) the most statistically reliable interpretation of the age of zircon crystallization most closely associated with the PST eruption.

Implications of results

Our new zircon U-Pb dates from two tuffs at Harvard Hill help to constrain the age of the Barstow Formation in this region. In addition, the Shamrock tuff is characterized by distinctive zircon geochemistry that allows us to compare it to other tuffs in the area, like the Peach Spring Tuff at Stoddard Wash, and provides implications for its volcanic source.

The 19.1 ± 0.3 Ma rhyolitic tuff (HH678) at southeast Harvard Hill (Fig. 2, 3) is underlain by and interbedded with lacustrine deposits (unit SE1, SE2), indicating that lacustrine sedimentation initiated at or before ~ 19.1 Ma. Because the rock-avalanche breccias (unit SE3) overlie the tuff, the rock avalanche events must have occurred at or after ~ 19.1 Ma. Additionally, since these breccias are not present in the northern Harvard Hill Miocene section nor do we observe the Shamrock tuff in the east fault block of the southern section, we assume that the breccias were deposited before the Shamrock tuff (i.e., before ~ 18.7 Ma). Based on this same assumption, we conclude that all of the Barstow Formation strata in the southeast fault block are older than the strata at northern Harvard Hill and the possibly correlative strata in the southwest fault block. Although strata below the Shamrock tuff at northern Harvard Hill are not exposed, we suggest that the tuff was deposited in a lacustrine environment (see discussion below) ~ 18.7 Ma.

Our large zircon U-Pb age and geochemistry data set for the Shamrock tuff also permits crucial comparisons with other tuffs, particularly the Peach Spring Tuff (PST). The PST at Kingman, Arizona was previously dated at 18.5 ± 0.2 Ma based on $^{40}\text{Ar}/^{39}\text{Ar}$ ages on bulk sanidine separates from pumice blocks (Nielson et al., 1990). In addition, Miller et al. (1998) reported a single-crystal $^{40}\text{Ar}/^{39}\text{Ar}$ sanidine age of 18.4 ± 0.1 Ma for the PST at the southern Little Piute Mountains in southeastern California. As noted by Hillhouse et al. (this volume), Ar ages for sanidine are somewhat uncertain due to inconclusive ages assumed for monitor minerals; therefore, Ar ages possibly range from 18.5 to 18.7 Ma. We have determined a similar age of 18.7 ± 0.1 Ma for the PST based on zircon U-Pb dating of five samples (HH871, H3233, and S1084 in this study; see also, Miller et al., this volume). High Th/U in zircons, specifically observed in the Shamrock tuff and Stoddard Wash PST samples (HH871, H3233, S1084; Fig. 9), is typical of zircons that we have analyzed in volcanic and crustal rocks of the Colorado River Extensional Corridor. This high Th/U signature is also indicated in zircons from two other PST samples (see Miller et al., this volume). Therefore, based on distinctive attributes, including the following: (1) mineralogy, i.e., megascopic sanidine and sphene; (2) outcrop thickness;

(3) remanent magnetization (Wells and Hillhouse, 1989; Hillhouse et al., this volume); (4) lack of detrital zircons; (5) cathodoluminescent zoning and textures; (6) zircon $^{206}\text{Pb}/^{238}\text{U}$ age determined in this study; (7) high Th/U in zircons; and, (8) zircon U-Pb age compatibility with sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ ages, we conclude that the Shamrock tuff is correlative with the Peach Spring Tuff at Stoddard Wash, and was therefore derived from the same ignimbrite eruption.

Discussion: depositional environments

The Shamrock tuff has unusual outcrop characteristics compared to other occurrences of the Peach Spring Tuff, which we attribute to its unique depositional environment. The typical Peach Spring Tuff deposit, is gray, pink, or dark purple and weakly to densely welded, whereas the Shamrock tuff is a light lime green color and nonwelded. Buesch (1991) investigated several environments in which the Peach Spring Tuff was deposited and noted that: (1) laterally continuous thin beds, (2) a lack of erosional or scoured surfaces, and, (3) the presence of limestone and stromatolitic limestone, together indicate local lacustrine deposition. In addition, non-clastic deposits, e.g. limestone, overlying the Peach Spring Tuff may indicate accumulation in an internally drained basin (Buesch, 1991). Lacustrine deposition, particularly in a shallow lake, also helps to preserve a nonwelded tuff, as the area affected by wind erosion is reduced and continuous sedimentation following tuff deposition is provided before erosion can occur (Buesch, 1991). Stratigraphically above the Shamrock tuff at Harvard Hill, beds of tuffaceous sandstone and limestone (unit N2), which become stromatolitic up-section (unit N3), are mostly laterally continuous and are conformable on the tuff. Therefore, we conclude that ~ 18.7 Ma the Shamrock tuff was deposited in a lacustrine setting, most likely internally drained, helping to preserve the nonwelded tuff. Moreover, the sparsity of lithic fragments in and considerable thickness (9 - 17 m, unexposed base) of the mostly massive Shamrock tuff suggests that it may have been deposited in a relatively deep lake. Deposition of this thick unit presumably caused an instantaneous decrease in lake depth, thus changing the depositional environment for subsequent sediments.

Although the intertonguing of limestone and tuffaceous sandstone (unit N2) stratigraphically above the Shamrock tuff at northern Harvard Hill demonstrates local lake transgression and regression cycles, it also more broadly indicates a nearshore environment adjacent to topography blanketed by tuff available to be reworked. Furthermore, the abundance of stromatolites in the massive limestone at northern Harvard Hill (unit N3) and at

southwest Harvard Hill (unit SW1) suggests deposition in a shallow lacustrine environment, where algal mats typically thrive. However, Reynolds (2004) suggested that large, non-tubular stromatolites may form in deeper water, so we cannot explicitly establish lake depth based on stromatolite structure.

Alluvial sedimentation followed basal lacustrine sedimentation as evidenced by the conglomerate units overlying the stromatolitic limestones at northern Harvard Hill and in the west fault block of southern Harvard Hill (i.e., N4 and SW2, respectively). The alluvial sedimentation may represent alluvial fan progradation and/or lake regression. Although the underlying stromatolitic limestones may be correlative, the markedly different lithologies, textures, and grain sizes of these conglomerates at northern and southwest Harvard Hill indicate distinct source areas. The clast assemblage in the SW2 conglomerate in the west fault block is identical to that of the rock-avalanche breccias and debris flow deposits in the east fault block, indicating similar source areas for deposits from both fault blocks at southern Harvard Hill. However, the northern conglomerate (N4) clast assemblage seems to have originated from a distinct source, possibly still influenced by Shamrock tuff material as the matrix of the conglomerate is tuffaceous. The stratigraphically highest units of the Barstow Formation at Harvard Hill, well-bedded limestone with water reed fossils (unit N5) or stromatolites (unit SW3) interbedded with lacustrine sandstone, suggest a return to a shallow lake environment and therefore, the onset of a transgressive cycle.

Silicification of the limestones at Harvard Hill may have developed during a secondary mineralization process similar to that described by Schuiling (1999) in the southern Calico Mountains. He indicated that chert exhalites developed as a result of sublacustrine geothermal hot spring activity associated with the Calico vein system and silver mineralization in the Calico District (Schuiling, 1999). Since we know that volcanism in the central Mojave Desert was at least in part coeval with Barstow basin lacustrine deposition, we assume that geothermal processes, and associated secondary mineralization, were likely to occur at Harvard Hill.

The oldest locally exposed strata of the Barstow Formation, located in the southeast fault block, are lacustrine siltstone and sandstone (unit SE1). This fine-grained, thinly bedded facies, stratigraphically below the white rhyolitic tuff (unit SE2), suggests deposition in a relatively deep lake at or before ~19.1 Ma. The rock avalanche breccia deposits (unit SE3) overlying the tuff and tuffaceous sand deposits of unit SE2 indicate steep source topography adjacent to the lacustrine setting. The ≥ 40 -m-thick breccia deposit almost certainly drastically changed the

depositional environment for subsequent sedimentation. Presumably the rock-avalanche event(s) would cause rapid local lake regression, decreasing water depth almost instantaneously. The shallow nearshore to offshore facies stratigraphically above the breccias verify this hypothesis. The clastic-rich limestone overlying the breccias (unit SE4) contains fossil *Physa* sp. aquatic snails; this species can subsist onshore, close to the lake margin, and in moist plant litter (R.E. Reynolds, written commun., 2010). Also, the tubular to mound-shaped stromatolites, interbedded lacustrine sand and gravel deposits, and reed fossils of unit SE4 all indicate a shallow nearshore depositional environment. Stratigraphically higher alluvial gravel deposits intertonguing with lacustrine sand deposits (unit SE5) again indicate a shallow lacustrine setting. Thus, since the rock-avalanche event(s) and coupled instantaneous local lake regression occurred sometime after ~19.1 Ma, overlying strata suggest that lake transgression ensued. We have no record of events between deposition of unit SE5 and the ~18.7 Ma Shamrock tuff (unit N1).

Detritus in the alluvial and debris flow deposits and rock-avalanche breccias of southern Harvard Hill (units SW2, SE5, and SE3, respectively) provide information on source areas. The prevalence of clasts that resemble the Jurassic Sidewinder Volcanics suggests a derivation from the south, as that formation is exposed in the Newberry Mountains and several other mountain ranges to the south and west (Schermer and Busby, 1994). To the southeast, metamorphosed Sidewinder Volcanics are exposed in the northwestern Cady Mountains (Miller and Leslie, unpublished map, 2009). In contrast, Sidewinder Volcanics exposures to the west and northwest are limited to small tracts less than a few hectares in size in the Lead Mountain and southeast Calico Mountains area (Cox and Wilshire, 1993; Singleton and Gans, 2008). Farther north, no exposures of the Sidewinder are known over a broad region; the only Sidewinder Volcanics lie far to the northeast in the Cronese Hills area (Walker et al., 1990).

Supporting the conclusion that rock avalanches and coarse-grained sedimentary rocks were derived from the south are the detrital zircon analyses we obtained from tuff samples HH871 and HH678 (units N1 and SE2, respectively). Ages obtained include: 1661, 248, 217, 150, 148, 147, 146, and 50 Ma. Paleoproterozoic rocks are known to the south in the Joshua Tree National Park area (e.g., Barth et al., 2000), but only a small exposure of Proterozoic rock, and limited in age to about 1400 Ma, is known to the north (Schermer et al., 1996). Triassic plutons are scattered both south and north of Harvard Hill, and in both places some are about 248 Ma (Miller et al., 1995; Barth et al., 1997); however younger Triassic plutons tend to be present only to the south (A. Barth,

written commun., 2009). Likewise, middle to late Jurassic intrusive and extrusive rocks are common to the south but most plutons north of Harvard Hill are older than the 150-146 Ma ages we obtained (Schermer and Busby, 1994; Miller et al., 1995; Barth et al., 2008). The detrital zircons are interpreted to represent local debris incorporated in the tuffs, and their data, therefore, support the interpretation that coarse clastic materials in the section were derived from a generally southerly direction.

Overall, a variable depositional environment, from a relatively deep lake to a very shallow lake or even onshore setting is evident in facies changes observed in just 150 m of section at Harvard Hill. We suggest that multiple transgression and regression cycles occurred locally during the early Miocene and perhaps into mid-Miocene time. Major depositional events, such as the emplacement of the thick Shamrock tuff and the massive rock avalanche breccias block(s), presumably helped to facilitate local and possibly regional adjustments in lake level. Moreover, the nearshore to onshore facies at Harvard Hill differ from basinal lacustrine deposits in the nearby Yermo Hills; coarse-grained sedimentation is more prominent in sections farther east (e.g., Alvord Mountain), most likely representing a basin margin setting (Woodburne, 1991). Coarse-grained facies near the base of the section at Harvard Hill are presumably sourced from the south and indicate proximity to steep topography; based on these interpretations, we may also speculate a southern basin margin setting at Harvard Hill during the early Miocene.

Discussion: stratigraphic correlations and paleogeographic implications

In this study, we have indicated that all Miocene strata at Harvard Hill are part of the Barstow Formation. However, the oldest strata, ~19 Ma, are comparable in age and largely consist of coarse-grained facies similar to Pickhandle Formation rocks. The Pickhandle Formation was defined in the northern Calico Mountains by McCulloh (1952) as a volcanoclastic sequence that contains multiple volcanic flows, but only minor amounts of lacustrine strata. In contrast, the overlying Barstow Formation tends to have sparse volcanoclastic beds, with most sediment interpreted as distal alluvial fan and lacustrine deposits. The age of the uppermost Pickhandle volcanoclastic strata is about 19.3 Ma in the Mud Hills (MacFadden et al., 1990; Woodburne et al., 1990) and about 19.0 Ma in the Calico Mountains (Singleton and Gans, 2008). Although lacustrine facies are more common in the Pickhandle Formation at its southern exposures, these facies are well below the top of the unit at Elephant Mountain (Fig. 1), perhaps ~21.0 to 21.5 Ma (Fillmore and Walker, 1996). However, rock avalanche deposits are common in the

upper part of the Pickhandle Formation. Still, we consider the lacustrine deposits and lack of volcanoclastic facies at Harvard Hill to indicate that the strata exposed are not the Pickhandle Formation; rock avalanche breccias are similar to upper Pickhandle facies, but they are mainly composed of Sidewinder Formation metavolcanic rocks, an uncommon constituent of the Pickhandle breccias. Therefore, we infer that the strata at Harvard Hill were deposited post-Pickhandle and are characterized as part of the Barstow Formation.

In an effort to elucidate Miocene paleogeography in the central Mojave Desert, we compare the timing and types of sedimentation and inferred depositional environments of the Barstow Formation at Harvard Hill with that of two well-described and well-dated sections: the Mud Hills type section and the Calico Mountains section. The Mud Hills are located ~30 km northwest of Harvard Hill and the Calico Mountains are ~12 km west of Harvard Hill (Fig. 1). We are particularly interested in understanding the relative timing of initial lacustrine sedimentation in each section and whether these facies were deposited in a single large basin or in multiple separate basins.

Fine-grained lacustrine deposition at Harvard Hill initiated at or before ~19.1 Ma. In contrast, fine-grained lacustrine sedimentation at the Mud Hills (Fig. 1) did not occur until at least ~16.3 Ma, as constrained by a ^{40}K - ^{40}Ar age of the Rak Tuff at the base of the Middle Member of the Barstow Formation (MacFadden et al., 1990; Woodburne et al., 1990). The Owl Conglomerate Member of the Barstow Formation, stratigraphically below the Middle Member, ranges in age from about ≤ 19.3 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$, Red Tuff, MacFadden et al., 1990) to 16.3 Ma. This unit is a cobble to boulder conglomerate and interfingers with granitic fanglomerates of the Pickhandle Formation (Dibblee, 1968; Woodburne et al., 1990). In the upper part of the Owl Conglomerate Member, Woodburne et al (1990) noted local fine-grained deposits and Reynolds (2004) indicated stromatolites and brown platy limestones dated at ~16.8 and 16.4 Ma, respectively, based on the magnetic polarity timescale (MacFadden et al., 1990); however, perennial fine-grained lacustrine deposition in the Mud Hills did not occur until at least ~16.3 Ma.

Different lithologies and modes of deposition of the basal part of the Barstow Formation are represented at Harvard Hill and the Mud Hills, indicating distinct depositional environments. Fine-grained sand and silt were deposited in a lacustrine environment at Harvard Hill at or before ~19.1 Ma, whereas coarse-grained gravels, presumably representing onshore alluvial facies, were rapidly accumulating in the Mud Hills (Woodburne et al., 1990). Moreover, coarse-grained facies that are interbedded

with the fin-grained lacustrine deposits near the base of the section at Harvard Hill differ in lithology and texture from the age-correlative coarse-grained facies of the Mud Hills. The rock avalanche breccias, alluvial gravel deposits, and debris flow deposits at southern Harvard Hill consist of mostly angular fragments of metavolcanic, granitic, and andesitic detritus, presumably sourced from the south. In contrast, Owl Conglomerate detritus consists of subrounded clasts of quartz monzonite, aplite, pegmatite, quartz, and andesitic porphyry sourced from the north (Dibblee, 1968; Woodburne et al., 1990).

In addition to basinal facies variation in the Barstow Formation, the marker units of Reynolds (2002, 2004) are also both chronologically and lithologically distinct. The marker beds in the coarse-grained Owl Conglomerate Member in the Mud Hills are ~16.8 to 16.4 Ma (Reynolds, 2004), but similar markers occur in a finer-grained and older section (~18.7 Ma) at Harvard Hill (Reynolds, 2002). In the Mud Hills, isolated stromatolite structures occur in gravel deposits (Reynolds, 2004), whereas the massive stromatolite limestone at Harvard Hill (units N4 and SW1; Reynolds, 2002) consists of domal-shaped stromatolites in a ≥14-m-thick, largely silicified limestone. A stromatolitic limestone that occurs in a coarse-grained interval associated with the rock avalanche breccias at Harvard Hill (unit SE4) is even older, ~19 Ma. Moreover, the brown platy limestones in the Mud Hills (Reynolds, 2004) also occur in gravel deposits, while the brown platy limestone at Harvard Hill (units N5 and SW3; Reynolds, 2002) is fossiliferous, locally silicified, and interbedded with finer-grained deposits. Furthermore, the distinct, thick Shamrock tuff of Harvard Hill (i.e., green tuff of Reynolds, 2002) is not known in the Mud Hills section, which may indicate poor preservation in alluvial sediments or it may be an artifact of the geographic extent of the rhyolite ash flow.

The Barstow Formation section at Harvard Hill better correlates, both chronologically and lithologically, with the Barstow Formation in the closer Calico Mountains (Fig. 1). The lacustrine section in the Calico Mountains is between ~19.0 and 16.9 Ma, based on $^{40}\text{Ar}/^{39}\text{Ar}$ dates for: (1) an upper Pickhandle Formation dacite whose detritus is present in beds at the base of the Barstow section and, (2) a dacite dome that intrudes the upper part of the section, respectively (Singleton and Gans, 2008). Thus, the lacustrine part of the Barstow Formation in the Calico Mountains is ~2.7 Ma older than the Mud Hills lacustrine section but is similar in age to lacustrine deposits at Harvard Hill, ≥19.1 to ≤18.7 Ma.

North of the Calico fault in the Calico Mountains, the Barstow Formation lacustrine interval overlies the Pick-

handle Formation and consists of limestone, sandstone, siltstone, and claystone; south of the fault, metavolcanic basement rocks are directly overlain by several breccia and conglomerate deposits that are stratigraphically below lacustrine limestone, siltstone, and sandstone (Singleton and Gans, 2008). The coarse- and fine-grained facies in the Barstow Formation south of the Calico fault may correlate with analogous strata at Harvard Hill. Chronology of both sections is similar; coarse-grained facies were deposited approximately 19.0 Ma in the Calico Mountains (Singleton and Gans, 2008) and ≤19.1 Ma at Harvard Hill. However, although older and younger parts of the Barstow Formation at Harvard Hill are missing, precluding detailed correlation, we can at least demonstrate that fine-grained lacustrine sedimentation predated rock avalanche breccia and alluvial deposition; this differs from observations in the Calico Mountains that indicate that coarse-grained facies deposition preceded fine-grained lacustrine deposition in the section south of the Calico fault (Singleton and Gans, 2008).

Coarse-grained non-lacustrine facies at the base of the Barstow Formation section south of the Calico fault in the Calico Mountains contain similar clast assemblages and possibly indicate similar timing and paleogeography as those of the rock avalanche and alluvial deposits near the base of the section at Harvard Hill. Clasts in the coarse-grained facies in the Calico Mountains include: metavolcanic rocks (possibly Jurassic Sidewinder Volcanics correlative), granite, basalt, reworked ash-fall tuff, and dacite (Singleton and Gans, 2008). Singleton and Gans (2008) indicate that the breccia may have been derived from the south. Similar rock types in coarse-grained facies at Harvard Hill, also apparently derived from the south, strengthen the possible correlation between the sections. Additionally, the monolithologic granite breccia in the Calico Mountains may have been deposited in a rock avalanche event, similar to the granite breccia at Harvard Hill. Overall, coarse-grained facies from both regions suggest proximity to a topographically steep basin margin at about the same time, ~19.0 Ma. This paleogeographic basin margin is discussed further in Miller et al. (this volume).

Marker units of Reynolds (2004) in Little Borate Canyon in the eastern Calico Mountains may correlate lithologically with strata at Harvard Hill (Reynolds, 2002), though chronology is less constrained. The stromatolites at Little Borate are large domal structures that occur in a massive silicified limestone, apparently close to the ~19.0 Ma base of the Barstow Formation (Reynolds, 2004; Singleton and Gans, 2008). This unit is very similar lithologically and perhaps in age to the ~18.7 Ma stromatolitic limestone at Harvard Hill. Although the brown

platy limestones in the Calico Mountains (Reynolds, 2004) occur in a fine-grained section similar to those at Harvard Hill, their age is more poorly constrained, making correlations difficult. Furthermore, the ~18.7 Ma Shamrock tuff, a distinctive marker unit of the Barstow Formation at Harvard Hill, was not observed by Singleton and Gans (2008) in the Calico Mountains section. If the Calico Mountains and Harvard Hill sections belonged to a single basin, which might be inferred based on corresponding lithology and chronology, we might expect to observe such a regionally extensive, thick rhyolite flow in both sections. The lack of a thick tuff near the base of the Calico Mountains section may indicate that the basins were separate at ~18.7 Ma.

Based on our geochronology and interpretations of depositional environments of Barstow Formation facies in distinct sections, we suggest that: (1) although basin infilling of the Barstow Formation may have started at about the same time across the region, lacustrine sedimentation initiated at Harvard Hill at least three million years before lake sediments were deposited in the Mud Hills; (2) lacustrine deposition initiated at Harvard Hill contemporaneous with or somewhat before lacustrine deposition south of the Calico fault in the Calico Mountains; and, (3) fairly similar coarse-grained facies were deposited at Harvard Hill and south of the Calico fault in the Calico Mountains at about the same time, ~19 Ma, and both potentially sourced from the south, suggesting similar depositional environments and possibly related basin margin settings at that time.

The aforementioned conclusions help to constrain models for basin paleogeography in the central Mojave Desert during the early to middle Miocene. Because the Barstow Formation section at Harvard Hill is markedly older than the lacustrine section at the Mud Hills, a model for a single, time-instantaneous lacustrine basin can be discarded. And, although presumably correlative coarse-grained facies were deposited at Harvard Hill and in the Calico Mountains ~19 Ma, suggesting a shared basin at that time, the lack of the regionally extensive Peach Spring Tuff (i.e., Shamrock tuff of Harvard Hill) in the Calico Mountains section suggests that the basins were separate ~18.7 Ma. Therefore, either of two paleogeographic models are allowed by our recent findings: (1) a single, roughly westward time-transgressive basin or, (2) multiple separate westward-migrating basins.

Conclusions

Our studies of Miocene strata at Harvard Hill have documented two rhyolitic tuffs, dated by U-Pb at about 19.1 and 18.7 Ma. The younger tuff is at least 9 m thick

and massive; its age, zircon geochemistry, mineralogy, and outcrop characteristics strongly support the interpretation that it represents a lacustrine facies of the Peach Spring Tuff. The tuffs define an age range of greater than 19.1 Ma to less than 18.7 Ma for the lacustrine sandstone, siltstone, limestone, and alluvial and breccia facies of the Barstow Formation at Harvard Hill. The alluvial and rock-avalanche breccia facies indicate a southerly source from a steep basin margin ~19.1 Ma.

Breccia at Harvard Hill is similar in age and lithology to the breccias at the Calico Mountains, but older and lithologically distinct from coarse-grained deposits at the more distant Mud Hills. In addition, the ~18.7 Ma age of limestone marker units at Harvard Hill is much older than analogous limestones at the Mud Hills. These observations help constrain models for basin paleogeography by requiring that distinctive facies in the Barstow Formation vary in age across the basin. Because similar coarse-grained facies were deposited at Harvard Hill and in the Calico Mountains at ~19 Ma, a shared basin at that time is possible, but the lack of a thick bed of the regionally extensive Peach Spring Tuff in the Calico Mountains section suggests that the basins were separate at ~18.7 Ma. More work is needed to distinguish between the two basic paleogeographic models for the Barstow Formation: (1) a single, roughly westward time-transgressive basin or, (2) multiple separate basins.

Acknowledgements

We are greatly thankful and indebted to Dean Miller, Ariel Strickland, and Trevor Dumitru for their help during several phases of the mineral separations of the SHRIMP samples. This research was in large part inspired by a field-trip led by Bob Reynolds...we appreciate Bob's enthusiasm to share his extensive knowledge on the Barstow Formation. We are grateful to Kim Bishop for enlightening us with his invaluable expertise on rock avalanche breccias. Discussions about the Peach Spring Tuff with Jonathan Miller, Calvin Miller, Bob Fleck, and Jack Hillhouse were illuminating. We also thank Jonathan Miller and Keith Howard for helpful comments and an insightful review on this paper.

References

- Barth, A.P., Tosdal, R.M., Wooden, J.L., and Howard, K.A., 1997, Triassic plutonism in Southern California: southward younging of arc initiation along a truncated continental margin: *Tectonics*, v. 16, no. 2, p. 290-304.
- Barth, A.P., Wooden, J.L., Coleman, D.S., and Fanning, C.M., 2000, Geochronology of the Proterozoic basement of southwesternmost North America, and the origin and evolution of the Mojave crustal province: *Tectonics* 19, 616-629.
- Barth, A.P., Wooden, J.L., Howard, K.A., and Richards J.L., 2008, Late Jurassic plutonism in the southwest U.S. Cordillera, *in* Ophiolites,

- arcs, and batholiths: a tribute to Cliff Hopson, Eds. J.E. Wright and J.W. Shervais: Geological Society of America Special Paper 438, p. 379-396.
- Black, L.P., Kamo, S.L., Allen, C.M., Davis, D.W., Aleinikoff, J.N., Valley, J.W., Mundil, R., Campbell, I.H., Korsch, R.J., Williams, I.S., and Foudoulis, C., 2004, Improved $^{206}\text{Pb}/^{238}\text{U}$ microprobe geochronology by the monitoring of a trace-element-related matrix effect; SHRIMP, ID-TIMS, ELA-ICP-MS and oxygen isotope documentation for a series of zircon standards: *Chemical Geology* v. 205, p. 115-140.
- Buesch, D.C., 1991, Changes in depositional environments resulting from emplacement of a large-volume ignimbrite, *in* Fisher, R.V., and Smith, G.A., eds., *Sedimentation in Volcanic Settings*, Society for Sedimentary Geology Special Publication No. 45, p. 139-153.
- Cox, B.F., and Wilshire, H.G., 1993, Geologic map of the area around the Nebo Annex, Marine Corps Logistics Base, Barstow, California: U.S. Geological Survey Open-File Report 93-568, 36 p., scale 1:12,000.
- Dibblee, T.W., Jr., 1968, Geology of the Opal Mountain and Fremont Peak quadrangles, California, California Division of Mines and Geology Bulletin, v. 188, 64 p.
- Dibblee, T.W., Jr., and Bassett, A.M., 1966, Geologic map of the Newberry Quadrangle, San Bernardino County, California, U.S. Geological Survey Miscellaneous geologic investigations, M(200), v. I, no. 461.
- Dokka, R.K., 1989, The Mojave extensional belt of southern California: *Tectonics*, v. 8, p. 363-390.
- Fillmore, R.P., 1993, Sedimentation and extensional basin evolution in a Miocene metamorphic core complex setting, Alvord Mountain, central Mojave Desert, California, USA: *Sedimentology*, v. 40, p. 721-742.
- Fillmore, R.P., Walker, J. D., 1996, Evolution of a supradetachment extensional basin: The lower Miocene Pickhandle Basin, central Mojave Desert, California: Geological Society of America Special Paper 303, p. 107-126.
- Gans, P.B., DeVecchio, D., Singleton, J., Van Pelt, J., Wong, M., and Reynolds, J., 2005, Cenozoic magmatic and structural evolution of the central Mojave Desert, California: New constraints from $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology and thermochronology: Geological Society of America Abstract with Programs, p. 103.
- Glazner, A.F., Bartley, J.M., and Walker, J.D., 1989, Magnitude and significance of Miocene crustal extension in the central Mojave Desert, California: *Geology*, v. 17, p. 50-53.
- Glazner, A.F., Nielson J.E., Howard K.A., and Miller D.M., 1986, Correlation of the Peach Springs Tuff, a large-volume Miocene ignimbrite sheet in California and Arizona: *Geology*, v. 14, p. 840-843.
- Glazner, A.F., Walker, J.D. Bartley, J.M., and Fletcher, J.M., 2002, Cenozoic evolution of the Mojave block of southern California: Geological Society of America Memoir 195, p. 19-41.
- Gusa, S., Nielson, J.E., Howard, K.A., 1987, heavy-mineral suites confirm the wide extent of the Peach Springs Tuff in California and Arizona, U.S.A., *Journal of Volcanology and Geothermal Research*, v. 33, p. 343-347.
- Hillhouse, J.W., Miller, D.M., and Turrin, B.D., this volume, Correlation of the Miocene Peach Springs Tuff with the Geomagnetic Polarity Time Scale and New Constraints on Tectonic Rotations in the Mojave Desert, California.
- Keaton, J.R., and Keaton, R.T., 1977, Manix fault zone, San Bernardino County, California, California Geology, p. 177-186.
- Leslie, S.R., Miller, D.M., Wooden, J.L., and Mahan, S.A., 2009, Neotectonics at the intersection between the Manix and Dolores Lake faults, Harvard Hill, central Mojave Desert, CA: Geological Society of America Abstracts with Programs, v. 41, no. 7, p. 456.
- Ludwig, K., 2001, Isoplot/EXversion 2.49: a geochronological toolkit for Microsoft Excel. Berkeley Geochronology Center Special Publication, v. 1a.
- Ludwig, K., 2002, SQUID 1.02, a Users Manual: Berkeley Geochronology Center Special Publication, v. 2.
- MacFadden, B.J., Swisher, C.C. III, Opdyke, N.D., and Woodburne, M.O., 1990, Paleomagnetism, geochronology, and possible tectonic rotation of the middle Miocene Barstow Formation, Mojave desert, southern California: Geological Society of America Bulletin, v. 102, p. 478-493.
- McCulloh, T.H., 1952, Geology of the southern half of the Lane Mountain quadrangle, California [Ph.D. thesis]: Los Angeles, University of California, 182 p.
- Meek, N., 1994, The stratigraphy and geomorphology of Coyote Basin, central Mojave Desert, California: San Bernardino County Museum Association Quarterly, vol. 47, no. 2, p. 32-34.
- Miller, D.M., and Yount, J.L., 2002, Late Cenozoic tectonic evolution of the north-central Mojave Desert inferred from fault history and physiographic evolution of the Fort Irwin area, California: Geological Society of America Memoir 195, p. 173-197.
- Miller, D.M., Leslie, S.R., Hillhouse, J.W., Wooden, J.L., Vazquez, J.A., and Reynolds, R.E., this volume, Reconnaissance geochronology of tuffs in the Miocene Barstow Formation; implications for basin evolution and tectonics in the central Mojave Desert.
- Miller, D.M., Dudash, S.L., Green, H.L., Lidke, D.J., Amoroso, L., Phelps, G.A., and Schmidt, K.M., 2007, A new Quaternary view of northern Mojave Desert tectonics suggests changing fault patterns during the late Pleistocene, *in* Miller, D.M. and Valin, Z.C., eds., *Geomorphology and tectonics at the intersection of Silurian and Death Valleys*, southern California: U.S. Geological Survey Open-File Report 2007-1424, p. 157-171.
- Miller, D.M., Leslie, S.R., Reheis, M., Schmidt, K.M., Amoroso, L., Dudash, S.L., and Mahan, S.A., 2009, Late Quaternary faulting and distributed deformation in the central Mojave Desert, California: Geological Society of America Abstracts with Programs, v. 41, no. 7, p. 171.
- Miller, J.S., Heizler, M. T., and Miller, C. F., 1998, Timing of magmatism, basin formation, and tilting at the west edge of the Colorado River Extensional Corridor: Results from single-crystal Ar/Ar geochronology of Tertiary rocks in the Old Woman Mountains area, southeastern California: *Journal of Geology*, v. 106, p. 195-209.
- Miller, J.S., Glazner, A.F., Walker, J.D., and Martin, M.W., 1995, Geochronologic and isotopic evidence for Triassic-Jurassic emplacement of the eugeoclinal allochthon in the Mojave Desert region, California: Geological Society of America Bulletin, v. 107, p. 1441-1457.
- Miller, J.S., Matzel, J.E.P., Miller, C.F., Burgess, S.D., and Miller, R.B., 2007, Zircon growth and recycling during the assembly of large, composite arc plutons: *Journal of volcanology and geothermal research*, v. 167, p. 282-299.
- Nielson, J.E., Lux, D.R. Dalrymple, G.B. and Glazner, A.F., 1990, Age of the Peach Springs Tuff, southeastern California and western Arizona: *Journal of Geophysical Research*, v. 95, p. 571-580.
- Reheis, M.C., Miller, D.M., Redwine, J.L., 2007, Quaternary stratigraphy, drainage-basin development, and geomorphology of the Lake Manix basin, Mojave Desert, Guidebook for fall field trip, Friends of the Pleistocene, Pacific Cell, Oct. 4-7.
- Reynolds, R.E., 2000, Marker units suggest correlation between the Calico Mountains and the Mud Hills, central Mojave Desert, California: San Bernardino County Museum Quarterly, v. 47, no. 2, p. 1-10.

- Reynolds, R.E., 2002, Between the basins: Field guide *in* Exploring the western Mojave and southern Basin and Range province, April 2002 Desert Symposium.
- Reynolds, R.E., 2004, Widespread early Miocene marker beds unite the Barstow Formation of the central Mojave Desert: Inland Geological Society, p. 1-10.
- Reynolds, R.E., Miller, D.M., and Woodburne, M.O., this volume, Extending the boundaries of the Barstow Formation in the central Mojave Desert.
- Ross, T.M., Luyendyk, B.P., and Haston, R.B., 1989, Paleomagnetic evidence for Neogene clockwise tectonic rotations in the central Mojave Desert, California: *Geology*, v. 17, p. 470-473.
- Schermer, E.R., and Busby, C.J., 1994, Jurassic magmatism in the central Mojave Desert: Implications for arc paleogeography and preservation of continental volcanic sequences, *Geological Society of America Bulletin*, v. 106, p. 767-790.
- Schermer, E.R., Luyendyk, B.P., Cisowski, S., 1996, Late Cenozoic structure and tectonics of the northern Mojave Desert: *Tectonics*, v. 15, p. 905-932.
- Simon, J.I., Renne, P.R., and Mundil, R., 2008, Implications of pre-eruptive magmatic histories of zircons for U-Pb geochronology of silicic extrusions: *Earth and Planetary Science Letters*, v. 266, p. 182-194.
- Schuling, T., 1999, A Miocene hot spring exhalite in the southern Calico Mountains, *in* Reynolds, R.E., and Reynolds, J., eds., *Tracks Along the Mojave: A Field Guide from Cajon Pass to the Calico Mountains and Coyote Lake: Redlands, California*, San Bernardino County Museum Association Quarterly, v. 46 (3), p. 89-94.
- Singleton, J.S., and Gans, P.B., 2008, Structural and stratigraphic evolution of the calico Mountains: Implications for early Miocene extensions and Neogene transpression in the central Mojave Desert, California: *Geosphere*, v. 4, p. 459-479.
- Walker, J.D., Bartley, J.M., Martin, M.W., and Coleman, D.S., 1990, Timing and kinematics of deformation in the Cronese Hills, California, and implications for Mesozoic structure of the southwestern Cordillera: *Geology*, v. 18, p. 554-557.
- Walker, J.D., Fletcher, J.M., Fillmore, R.P., Martin, M.W., Taylor, W.J., Glazner, A.F., and Bartley, J.M., 1995, Connection between igneous activity and extension in the central Mojave metamorphic core complex: *Journal of Geophysical Research*, v. 100, no. B6, p. 10477-10494.
- Wells, R.E., and Hillhouse, J.W., 1989, Paleomagnetism and tectonic rotation of the lower Miocene Peach Springs Tuff: Colorado Plateau, Arizona, to Barstow, California: *Geological Society of America Bulletin*, v. 101, p. 846-863.
- Woodburne, M.O., 1991, The Mojave Desert Province, *in* *Inland Southern California: the last 70 million years*: San Bernardino County Museum Association Quarterly, v. 38, p. 60-78.
- Woodburne, M.O., Tedford, R.H., and Swisher, C.C., III., 1990, Lithostratigraphy, biostratigraphy and geochronology of the Barstow Formation, Mojave Desert, southern California. *Geological Society of America Bulletin*, v. 102, p.459-477.
- Yarnold, J.C. and Lombard, J.P., 1989, A facies model for large rock-avalanche deposits formed in dry climates, *in* Colburn, I.P., Abbott, P.L., and Minch, J., eds., *Conglomerates in Basin Analysis: A Symposium Dedicated to A.O. Woodford*: Pacific Section S.E.P.M., v. 62, p. 9-31.
- Young, R.A. and Brennan, W.J., 1974, Peach Springs Tuff: Its bearing on structural evolution of the Colorado Plateau and development of Cenozoic drainage in Mojave County, Arizona, *Geological Society of America Bulletin*, v. 85, p. 83-90.

Correlation of the Miocene Peach Spring Tuff with the geomagnetic polarity time scale and new constraints on tectonic rotations in the Mojave Desert, California

John W. Hillhouse¹, David M. Miller¹, and Brent D. Turrin²

¹*U. S. Geological Survey, MS 975, 345 Middlefield Rd., Menlo Park, CA 94025*

²*Geological Sciences, Rutgers University, Piscataway, NJ 08854-8066*

Abstract

We report new paleomagnetic results and $^{40}\text{Ar}/^{39}\text{Ar}$ ages from the Peach Spring Tuff (PST), a key marker bed that occurs in the desert region between Barstow, California, and Peach Springs, Arizona. The $^{40}\text{Ar}/^{39}\text{Ar}$ ages were determined using individual hand-picked sanidine crystals from ash-flow specimens used in previous paleomagnetic studies at eight sites correlated by mineralogy, stratigraphic position, and magnetic inclination. Site-mean ages, which range from 18.43 Ma to 18.78 Ma with analytical precision (1 s.d.) typically 0.04 Ma, were obtained from areas near Fort Rock, AZ; McCullough Mts, NV; Cima Dome, Parker Dam, Danby, Ludlow, Kane Wash, and Stoddard Wash, CA. The regional mean age determination is 18.71 ± 0.13 Ma, after the data were selected for sanidine crystals that yielded greater than 90% radiogenic argon ($N = 40$). This age determination is compatible with previous $^{40}\text{Ar}/^{39}\text{Ar}$ dating of the PST after taking various neutron-flux monitor calibrations into account. We report paleomagnetic results from eight new sites that bear on reconstructions of the Miocene basins associated with the Hector Formation, Barstow Formation, and similar fine-grained sedimentary deposits in the Barstow region. Key findings of the new paleomagnetic study pertain to age control of the Hector Formation and clockwise rotation of the Northeast Mojave Domain. Our study of a rhyolitic ash flow at Baxter Wash, northern Cady Mountains, confirms the correlation of the PST within the Hector Formation and prompts reinterpretation of the previously determined magnetostratigraphy. Our model correlates the PST to the normal-polarity zone just below the C6–C5E boundary (18.748 Ma) of the astronomically tuned Geomagnetic Polarity Time Scale. After emplacement of the Peach Spring Tuff at Alvord Mountain and the Cady Mountains, the southern part of the Northeast Mojave Domain (between Cady and Coyote Lake faults) underwent clockwise rotation of 30° – 55° . Clockwise rotations increase with distance northward from the Cady fault and may reflect Late Miocene and younger accommodation of right-lateral motion across the Eastern California Shear Zone. The new results also expand the area known to be affected by the Peach Springs eruption, and confirm that a pink ash-flow tuff surrounding Daggett Ridge near Barstow is part of the PST.

Introduction

Previous investigations of the Miocene Peach Springs Tuff of Young and Brennan (1974) have demonstrated its usefulness for stratigraphic correlation across a 350 km-wide zone that comprises the Colorado River extensional corridor, the Basin-and-Range province, and the Eastern California Shear Zone (Nielson and Glazner, 1986; Glazner et al., 1986; Buesch and Valentine, 1986; Gusa et al., 1987; Nielson et al., 1990). The tuff is unique by being the only Miocene marker bed that links tectonically disturbed domains in the central Mojave Desert to the stable Colorado Plateau (Figure 1). Paleomagnetic studies of the Peach Spring Tuff indicated that the ash flow quickly acquired near-uniform directions of remanent magnetization after the voluminous eruption settled and cooled (Wells and Hillhouse, 1989; Hillhouse and Wells, 1991). At most outcrops, the tilt-corrected direction of remanent

magnetization is useful for inferring vertical axis rotations relative to the Colorado Plateau. As noted by Wells and Hillhouse (1989) and Varga et al. (2006), however, some accumulations of the tuff show variations in the direction of remanent magnetization that may be due to rapid changes in the geomagnetic field during emplacement and/or cooling of the ash flow. Distal deposits of the tuff generally preserve a distinctive shallow magnetic inclination (36.4°) that lies near the outer fringe of geomagnetic secular variation relative to the expected mean dipole inclination for Miocene time. Some sites preserve a steeper inclination (54.8°) near the mean dipole-field value, particularly in the proximal distributions of the tuff between Kingman, Arizona, and the Piute Mountains, California (Wells and Hillhouse, 1989).

Isotopic dating of the Peach Spring Tuff has been problematic. In a review of results from conventional

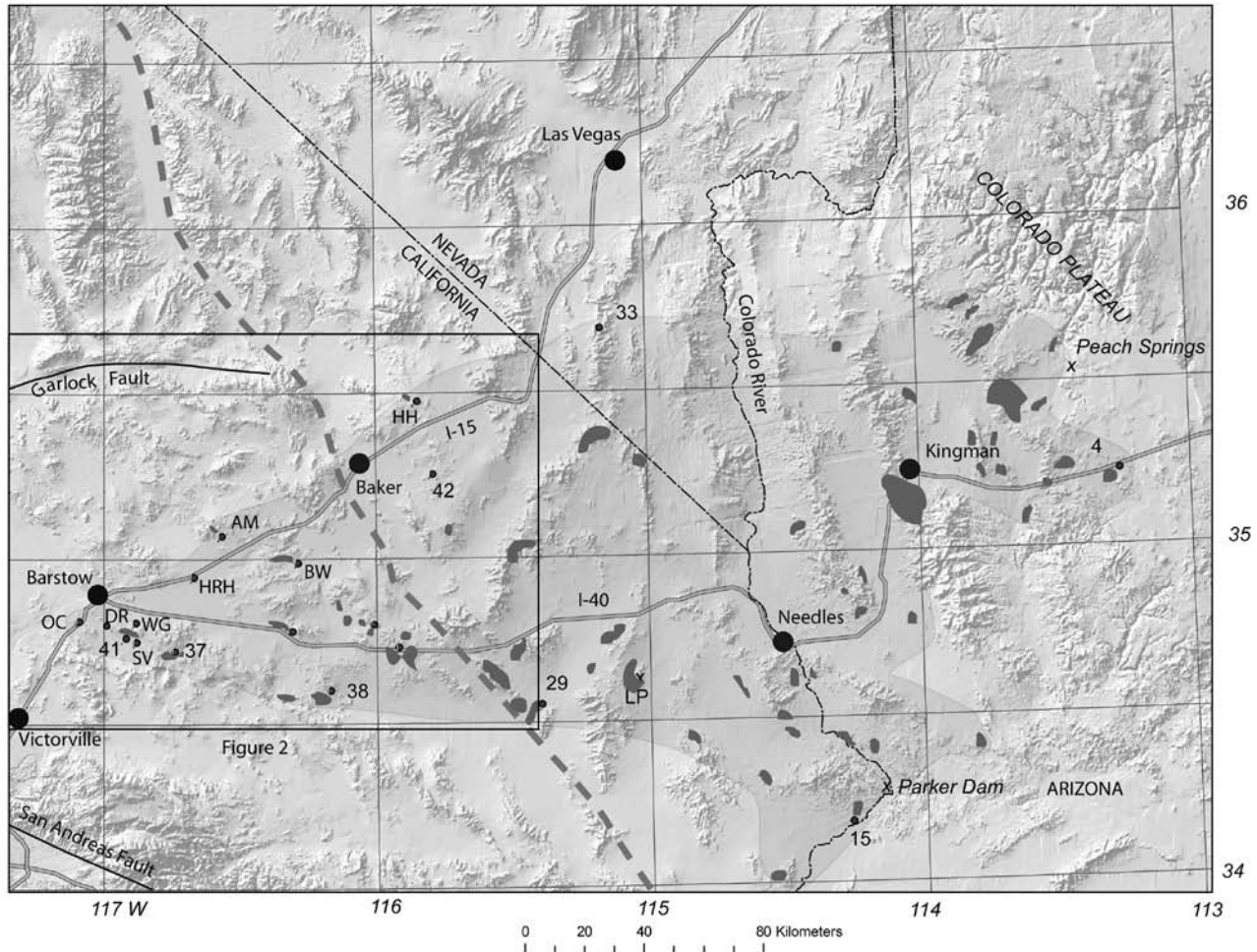


Figure 1. Maximum known extent (light shaded area) and generalized outcrops (dark shaded areas) of the Peach Spring Tuff in California, Arizona, and Nevada, modified from Glazner et al. (1986). Heavy dashed line marks eastern boundary of the Eastern California Shear Zone. Sites studied for paleomagnetism: AM, Alvord Mountain; BW, Baxter Wash, DR, Daggett Ridge; HH, Halloran Hills; HRH, Harvard Hill; OC, Outlet Center; SV, Stoddard Valley; WG, West Gem (This study). Sites dated by $^{40}\text{Ar}/^{39}\text{Ar}$ (Table 1): 4, Fort Rock; 15, Parker Dam; 29, Danby; 33, McCullough Mountains; 37, Kane Wash; 38, Pacific Mesa; 41, Stoddard Wash; 42, Cima; LP, Little Piute (Miller et al., 1998).

potassium-argon and various $^{40}\text{Ar}/^{39}\text{Ar}$ dating methods, Nielson et al. (1990) found that biotite and sanidine separates from bulk samples of the Peach Spring Tuff produced inconsistent ages ranging from 16.2 Ma to 20.5 Ma. Their best estimate of the emplacement age (18.5 ± 0.2 Ma) was obtained by $^{40}\text{Ar}/^{39}\text{Ar}$ methods (laser fusion and incremental release techniques) applied to sanidine from juvenile pumice. Incomplete extraction of radiogenic argon, excess argon, and, in some cases, xenocrystic contamination of mineral separates were cited as possible explanations for the wide range of ages determined previously from the Peach Spring Tuff. Ash-flow tuffs are especially susceptible to entrainment of xenocrysts as the flow advances over the landscape or by the addition of vent wall material during explosive eruptions. Single-crystal $^{40}\text{Ar}/^{39}\text{Ar}$ dating offers a means for assessing xenocrystic contamination, and the method has been applied successfully to the Peach Spring Tuff. In reporting an isochron

age of 18.42 ± 0.07 Ma for the Peach Spring Tuff in the Little Piute Mountains, Miller et al. (1998) identified an obvious outlying age from a xenocryst among eight clear sanidine crystals. Emplacement temperatures of welded ash-flow tuffs are sometimes insufficient to fully reset the K/Ar system in sanidine, especially in thin units (Spell et al., 2001).

A major objective of our study is to extend $^{40}\text{Ar}/^{39}\text{Ar}$ dating of the Peach Spring Tuff throughout its extent, and in particular to confirm correlations of the ash flow in fine-grained sedimentary sections in the vicinity of Barstow, California. Questions have been raised about the correlation to the Peach Spring Tuff to "pink tuffs" surrounding Daggett Ridge and Stoddard Valley and the relationship between the marker beds and vertebrate fossil localities. We present single-crystal $^{40}\text{Ar}/^{39}\text{Ar}$ ages from eight sites that were sampled for paleomagnetism by Wells and Hillhouse (1989) and Hillhouse and Wells

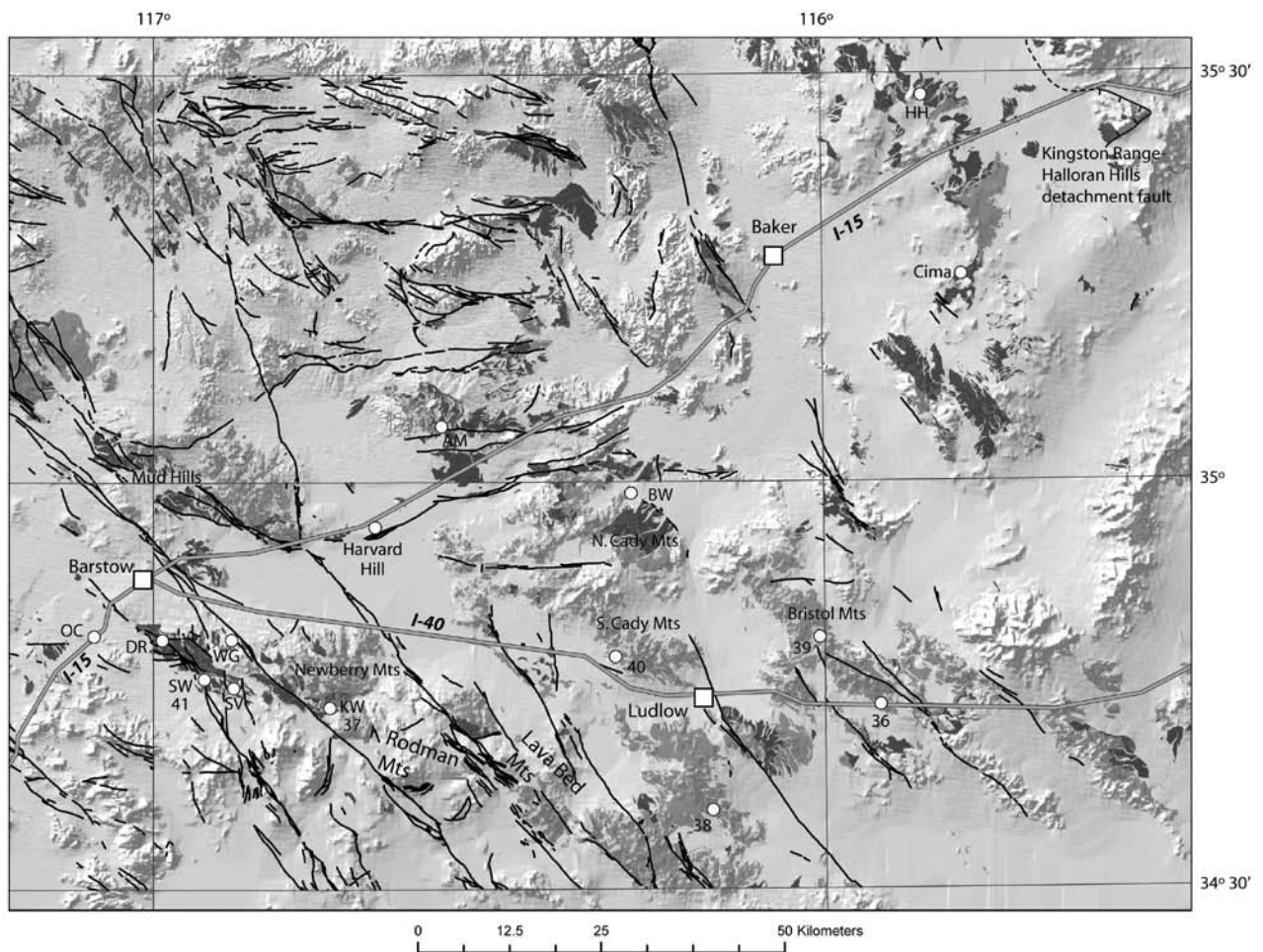


Figure 2. Map showing Quaternary faults (Miller et al., 2007), excluding the Kingston Range-Halloran Hills detachment fault of Miocene age, Tertiary volcanic deposits (light shading), Tertiary sedimentary deposits (dark shading). Sites (dots) in the Peach Spring Tuff studied for paleomagnetism: AM, Alvord Mountain; BW, Baxter Wash, Cima (Hillhouse and Wells, 1991); DR, Daggett Ridge; HH, Halloran Hills; Harvard Hill (HRH); OC, Outlet Center; SV, Stoddard Valley; SW, Stoddard Wash; WG, West Gem. Numbered paleomagnetism sites from Wells and Hillhouse (1989): 36, Bristol Mts.; 37, Kane Wash; 38, Pacific Mesa; 39, Broadwell Lake; 40, Sleeping Beauty; 41, Stoddard Wash.

(1991). We compare the new ages to previous determinations after converting all results to a common monitor mineral age.

We have extended paleomagnetic sampling of prospective Peach Spring Tuff localities in Stoddard Valley, Daggett Ridge, Harvard Hill, northern Cady Mountains, Alvord Mountain, and Halloran Hills for a total of eight new sites (Figure 2). The first four localities, which surround Barstow, are critical for correlating limestones and fine-grained lake deposits in the now-dismembered Barstow Basin. Reconstruction of the basin deposits is critical for understanding tectonic events and vertebrate evolution in the region. Our site in the northern Cady Mountains is especially pertinent to calibrating the age of the Hector Formation, which contains a rhyolite ash-flow tuff long considered to be correlative with the Peach Spring Tuff (Miller, 1980, and personal communication, 1985; Woodburne, 1991; Buesch, 1994; Woodburne, 1998). Confirming the presence of Peach Spring Tuff in the Hector Formation via paleomagnetism would provide a critical tie point for the magnetostratigraphy (MacFadden et al., 1990a) and vertebrate biostratigraphy (Woodburne, 1998) of the lower Miocene deposits. Confirmation of the correlation would also allow refinement of the clockwise rotation measurement obtained from the Hector Formation by MacFadden et al. (1990a). Our paleomagnetic site at Alvord Mountain tests correlation of the Peach Spring Tuff to a volcanic ash within the Spanish Canyon Formation, which would extend the known distribution of the marker 20 km farther north into the Northeast Mojave Domain and test clockwise rotation estimates from earlier paleomagnetic studies (Ross et al., 1989; Schermer et al., 1996). Sampling at Halloran Hills extends correlation of the tuff 50 km northward into the Shadow Valley basin, with the objective of testing for upper-plate rotations in an area known for detachment faulting and large-scale gravity sliding (Davis and Friedmann, 2005).

Geologic setting of sample sites

The Peach Springs Tuff of Young and Brennan (1974) is a distinctive sphene-bearing ignimbrite that was originally identified in lower Miocene sedimentary and volcanic sequences in western Arizona. Another common feature of the tuff is the presence of large sanidine crystals that show a blue flash (schiller) under a hand lens. The widespread extent of the Peach Spring Tuff was recognized by Glazner et al. (1986), who proposed that isolated outcrops could be traced from Peach Springs, Arizona, as far south as Parker Dam, California, and west to Barstow, California. Subsequent work involving geologic mapping, stratigraphy, geochronology, mineralogy, and paleomagnetism

firmly established the regional correlation (Buesch and Valentine, 1986; Gusa et al., 1987; Wells and Hillhouse, 1989; Nielson et al., 1990; Nielson and Beratan, 1995). In general, the tuff exceeds 10 m in thickness and is densely to incipiently welded. Our $^{40}\text{Ar}/^{39}\text{Ar}$ age determinations are from representative sites throughout the distribution of the Peach Spring Tuff; geologic and structural settings of the dated sites are given in Wells and Hillhouse (1989) and Hillhouse and Wells (1991). The following subsections describe geologic details of sites used in the current paleomagnetism study (Figure 2).

Daggett Ridge and Stoddard Valley

A sequence of Tertiary sedimentary rocks that comprises sandstone, lacustrine limestone, and tuff is in unconformable contact with older conglomerates and volcanic breccia of Daggett Ridge. The volcanic ash unit (Unit Tst of Dibblee 1964, 1970) within the finer-grained sequence is correlated with the Peach Spring Tuff. Wells and Hillhouse (1989) reported paleomagnetic data (Site 41) from an outcrop of Tst near Stoddard Wash on the south side of Daggett Ridge. Some of the paleomagnetic samples from this site were used in the $^{40}\text{Ar}/^{39}\text{Ar}$ study. We recently extended the paleomagnetic sampling to a second site ("Stoddard Valley": 9J835) in subhorizontal beds of Tst approximately 2 km southeast of Site 41. Northeast of Daggett Ridge, the Tertiary sedimentary sequence is tilted steeply and displaced downward by the Camp Rock fault. In the part of the section just above granitic conglomerate, we sampled a sphene- and sanidine-bearing lithic pink tuff ("West Gem": 9J824) for paleomagnetism. Ten km west of the West Gem site, pink tuff is exposed near the northwest end of Daggett Ridge and is cut by the Lenwood fault. Dibblee (1970) mapped the tuff (Unit Tst) at the base of the Tertiary sedimentary section, essentially in the same stratigraphic position as the West Gem tuff; paleomagnetism site 6J101 ("Daggett Ridge") is from this tuff exposure.

Near the intersection of Outlet Center Drive and I-15, we collected samples from a quarried, isolated body of pink welded tuff (9J803). Although this outcrop was mapped as an intrusive dacite (Unit Tid) by Dibblee (1960), more recent quarrying has exposed the base which appears to be depositional on claystone. The mineralogy of this tuff is consistent with that of the sphene- and sanidine-bearing pink tuffs of Daggett Ridge.

Harvard Hill

Approximately 50 m of fine-grained sedimentary rocks, including waterlain tuff, siltstone, and siliceous limestone, are exposed on the northern side of Harvard Hill. At the base of this section is a greenish, sphene-bearing tuff, which is ~10 m thick. Uranium-lead dating on zircon

crystals from the tuff gave ages of 18 Ma–19 Ma, suggesting a possible correlation with the Peach Spring Tuff (Leslie et al., this volume; Miller et al., this volume). Site 9J855 was collected in the greenish tuff approximately 3 m from the base of the lacustrine sedimentary section.

Northern Cady Mountains

We sampled a welded rhyolite tuff within a 400-m-thick sequence of fine-grained volcanoclastic rocks, basalt, and tuff, which comprises the Hector Formation (Woodburne et al., 1974) in the northern Cady Mountains. Our paleomagnetism site (9J844, Baxter Wash) is in the same vicinity as stratigraphic sections that were measured by Miller (1980) and MacFadden et al. (1990a). Age control of the section is provided by conventional K–Ar (recalculated to modern decay constants by Miller, 1980), paleontology, and magnetostratigraphy. As shown in the measured section of MacFadden et al. (1990a), a tuff 23 m from the base of the section gave an age of 22.9 ± 0.4 Ma, basalt 110 m above the base gave 18.6 ± 0.2 Ma, and the welded rhyolite tuff 220 m from the base gave 17.9 Ma. Vertebrate fossils of Hemingfordian age from the Hector Formation constrain the upper part of the section to 16 Ma–19 Ma, on the basis of correlations with the lower part of the Barstow Formation (MacFadden et al., 1990b). MacFadden et al. (1990a) determined magnetostratigraphy of the Hector Formation, showing that the upper rhyolite tuff, which is the target of our paleomagnetism study, has normal polarity and is near the lower boundary of a thin reversed-polarity zone. The tuff was mapped as Unit Trf (rhyolitic felsite) by Dibblee and Bassett (1966) within finer sedimentary beds that overlie a suite of Tertiary volcanic rocks. Woodburne et al. (1991) and Woodburne (1998) noted that a similar welded tuff in the southern and northern Cady Mountains was possibly correlative with the Peach Spring Tuff.

Alvord Mountain

In the western part of Alvord Mountain, Mesozoic plutonic rocks are overlain by a Miocene section approximately 700 m thick and composed of the Clews fanglomerate, Alvord Peak basalt, Spanish Canyon Formation, and Barstow Formation (Byers, 1960). The Spanish Canyon Formation consists of 100 m of arkosic sandstone, granitic conglomerate, tuff, and olivine basalt flows. A pink, sanidine- and sphene-bearing tuff occurs near the middle of the Spanish Canyon Formation. This partly welded tuff contains flattened pumice clasts in a fine-grained matrix of devitrified glass shards, lithic fragments, and phenocrysts.

We collected samples for paleomagnetic analysis from the pink welded tuff approximately 2 km southwest of Clews Ridge. The 2-m-thick tuff bed fills a channel in coarse tuffaceous sandstone, which has been deformed

into a gently plunging anticline. Our sample locality is stratigraphically below basalts studied by Ross et al. (1989), who inferred a clockwise rotation of 53° for the Alvord Mountain region from paleomagnetic declinations of 9 flows.

Halloran Hills

In the southern Halloran Hills, Tertiary and Quaternary rocks form a thin veneer above a pediment cut into granitic rocks of the Teutonia batholith and minor Paleozoic limestone (Hewitt, 1956; Beckerman et al., 1982; Reynolds and Nance, 1988). Just east of Squaw Mountain, white volcanic ash and purple welded tuff make up the base of the Tertiary section, which in turn is overlain by lacustrine limestone and pyroxene andesite (Reynolds and McMackin, 1988). The tuff, which we sampled for paleomagnetism, dips moderately northward. Sedimentary beds near the level of the pyroxene andesite contain an assemblage of rodent fossils from the Clarendonian land mammal age (Reynolds and McMackin, 1988). Coarse fanglomerates and megabreccias of Paleozoic limestone overlie the tilted and faulted sedimentary section. The conglomerates, which suggest a period of rapid uplift and gravity sliding (Davis et al., 1991, 1993), are tilted and capped by less deformed basalt as old as 5.1 Ma (Turrin et al., 1985).

According to Davis and Friedmann (2005), several large autochthonous sheets and glide blocks lie above the Kingston Range–Halloran Hills detachment fault in the Shadow Valley basin. The tilted blocks of Halloran Hills have moved several tens of km westward. Our site is within welded tuff from the transported upper plate of the extensional complex.

$^{40}\text{Ar}/^{39}\text{Ar}$ methods

Specimens from the paleomagnetic studies of the Peach Spring Tuff were crushed with a mortar and pestle, and grains of sanidine were handpicked from the crushed bulk material. The grains were then treated with a 5% hydrofluoric acid solution in an ultrasonic bath to remove adhering matrix material, rinsed in distilled water, and dried. Seven grains of sanidine were selected from each site after examination under a binocular microscope. Some samples were irradiated at the TRIGA reactor, University of California, Berkeley, others at the Los Alamos Omega West reactor. The following potassium and calcium corrections were determined using optical-grade CaF₂ and a laboratory potassium glass: Berkeley TRIGA reactor: $^{40}\text{Ar}_K/^{39}\text{Ar}_K = 0.0086$; $^{39}\text{Ar}_{Ca}/^{37}\text{Ar}_{Ca} = 9.0 \times 10^{-4}$; $^{36}\text{Ar}_{Ca}/^{37}\text{Ar}_{Ca} = 2.59 \times 10^{-4}$; Los Alamos Omega West: $^{40}\text{Ar}_K/^{39}\text{Ar}_K = 0.0024$; $^{39}\text{Ar}_{Ca}/^{37}\text{Ar}_{Ca} = 7.2 \times 10^{-4}$; $^{36}\text{Ar}_{Ca}/^{37}\text{Ar}_{Ca} = 2.7 \times 10^{-4}$. Argon extractions and isotopic

Table 1. Single-crystal $^{40}\text{Ar}/^{39}\text{Ar}$ ages from sanidine, Peach Spring Tuff in California, Arizona, and Nevada.

Locality	Pmag Site	Latitude	Longitude	BGC ID	%rad Ar	Age (Ma)	s.d. (Ma)	s.e.mean (Ma)
Fort Rock, Arizona	4	35.227	246.740	1965-01	78.9	18.484	0.644	
				1965-02	97.0	18.658	0.094	
				1965-03	98.8	18.757	0.072	
				1965-04	97.6	18.608	0.090	
				1965-05	97.0	18.552	0.078	
				1965-06	96.1	18.708	0.094	
				1965-07	98.3	18.649	0.086	
				Arithmetic mean				
Weighted mean					18.657	0.034	0.035	
Parker Dam, CA	15	34.182	245.729	1963-01	89.3	18.794	0.084	
				1963-02	97.8	18.872	0.111	
				1963-03	89.0	18.684	0.137	
				1963-04	53.8	18.190	0.500	
				1963-05	97.2	18.743	0.108	
				1963-06	99.2	18.742	0.084	
				1963-07	86.4	18.725	0.174	
				Arithmetic Mean				
Weighted mean					18.764	0.043	0.084	
Danby, CA	29	34.553	244.601	1962-01	97.4	18.717	0.079	
				1962-02	98.4	18.631	0.076	
				1962-03	99.0	18.895	0.075	
				1962-04	97.6	18.769	0.071	
				1962-05	92.7	18.585	0.099	
				1962-06	94.3	18.713	0.082	
				1962-07	97.5	18.865	0.079	
				Arithmetic mean				
Weighted mean					18.750	0.030	0.043	
McCullough Mts, NV	33	35.690	244.833	1959-01	82.9	18.310	0.169	
				1959-02	96.9	18.766	0.116	
				1959-03	84.2	18.481	0.403	
				1959-04	99.4	18.852	0.079	
				1959-05	78.6	18.428	0.565	
				1959-06	96.9	18.757	0.096	
				1959-07	63.2	19.810	1.173	
				Arithmetic mean				
Weighted mean					18.753	0.051	0.189	
Kane Wash, CA	37	34.722	243.258	1964-01	90.9	18.732	0.205	
				1964-02	81.5	18.595	0.246	
				1964-03	94.1	18.743	0.098	
				1964-04	93.6	18.852	0.091	
				1964-05	94.9	18.759	0.093	
				1964-06	62.3	18.808	0.803	
				1964-07	98.5	18.631	0.080	
				Arithmetic mean				
Geometric mean					18.734	0.043	0.035	

Table 1 continues

Table 1 continued

Pacific Mesa, CA	38	34.598	243.831	1960-01	94.6	18.376	0.095	
				1960-02	80.4	18.493	0.189	
				1960-03	94.2	18.244	0.095	
				1960-04	93.6	18.691	0.226	
				1960-05	92.4	18.496	0.262	
				1960-06	95.7	18.691	0.106	
				1960-07	82.7	18.283	0.212	
				Arithmetic mean				
Geometric mean					18.434	0.050	0.068	
Stoddard Wash, CA	41	34.757	243.081	1961-01	77.2	18.339	0.532	
				1961-02	45.8	18.895	0.665	
				1961-03	94.4	18.772	0.080	
				1961-04	95.5	18.736	0.088	
				1961-05	97.7	18.811	0.072	
				1961-06	99.4	18.876	0.074	
				1961-07	98.2	18.708	0.069	
				Arithmetic mean				
Weighted mean					18.780	0.034	0.071	
Cima, CA	42	35.253	244.209	1966-01	99.0	18.707	0.069	
				1966-02	98.6	18.648	0.068	
				1966-03	98.8	18.727	0.080	
				1966-04	98.4	18.777	0.079	
				1966-05	82.5	18.785	0.098	
				1966-06	96.7	18.604	0.078	
				1966-07	98.3	18.881	0.078	
				Arithmetic mean				
Geometric mean					18.881	0.078	0.035	

analyses of single crystals were conducted at Berkeley Geochronology Center using a fully automated laser-fusion extraction system. The system is in-line with a Mass Analyzer Products (MAP) 215 mass spectrometer with a modified Neir-type source and an EMI 18 stage electron multiplier. Argon backgrounds for the system and mass spectrometer were typically: $^{40}\text{Ar} = 4.0 \times 10^{-12}$ cc (STP); $^{39}\text{Ar} = 1.0 \times 10^{-13}$ cc; $^{37}\text{Ar} = 7.6 \times 10^{-14}$ cc; and $^{36}\text{Ar} = 3.9 \times 10^{-14}$ cc. During and immediately after fusion of the sample, the gases were allowed to react with a Zr-V-Fe alloy getter at approximately 400° C and then with a liquid-nitrogen cold finger. After the initial clean-up period, the sample gases were purified further with an alloy getter operated at 125° C. Argon isotopic measurements were typically measured to $\pm 0.1\%$ at 10^{-14} to 10^{-16} moles (10^{-12} to 10^{-14} cc STP). The mineral standard used to monitor reactor neutron flux for the age calibration was Fish Canyon Tuff sanidine dated at 27.84 Ma; decay constants were those recommended by Steiger and Jaeger (1977).

$^{40}\text{Ar}/^{39}\text{Ar}$ results

Individual crystal ages and the radiogenic argon percentages are presented in Table 1. For each site, the ages are given as the arithmetic mean and as the weighted mean; the analytical precision is quoted to one standard deviation of the mean. Individual sanidine crystal ages range from 18.190 Ma to 19.810 Ma, with standard deviations ranging from 0.1 to 0.01 Ma. In contrast, weighted site-mean ages fall in a narrower range from 18.434 Ma to 18.780 Ma, with standard errors of the mean about 0.04 Ma. A histogram of the single-crystal ages shows a very strong peak in the single-crystal age distribution between 18.7–18.8 Ma for all sites combined (Figure 3). Also, the younger secondary peak (18.45 Ma) in the distribution is dominated by crystals with radiogenic argon ratios of less than 90%, most notably at Site 33 (McCullough Mts.) and Site 38 (Pacific Mesa). The younger analyses generally have large errors, and probability density analysis shows that the younger mode on the histogram is not significant

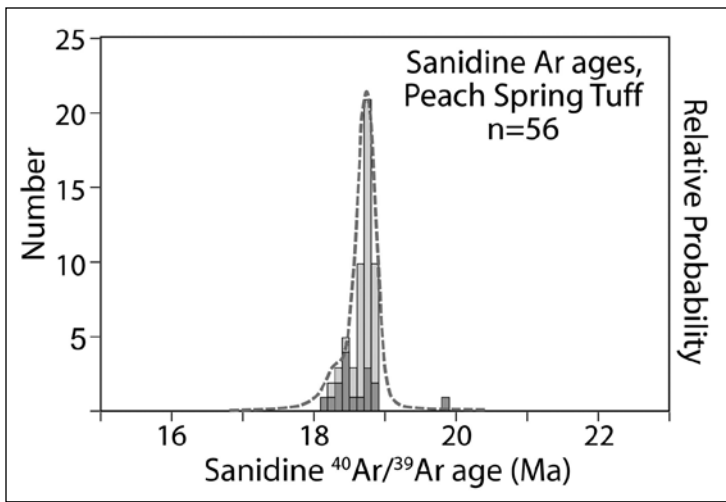


Figure 3. Histogram showing distribution of all $^{40}\text{Ar}/^{39}\text{Ar}$ single-crystal sanidine ages and relative probability curve (dashed line) from eight sites in the Peach Spring Tuff (Figure 1; Table 1). Dark shading indicates crystals yielding greater than or equal to 90% radiogenic argon; light (orange) shading indicates less than 90%.

(Figure 3). Selecting the remaining sites on the basis of higher radiogenic argon ratios gives site-mean ages ranging from 18.657 ± 0.034 Ma to 18.780 ± 0.034 Ma. The narrow spread in site-mean ages is consistent with the paleomagnetic and geologic observations that supported regional correlation of the Peach Spring Tuff, as presented in Wells and Hillhouse (1989) and Hillhouse and Wells (1991). Combining ages from all single crystals having radiogenic argon that exceeds 90% ($N = 40$) gives a mean age of 18.71 ± 0.13 Ma for our regional sampling of the Peach Spring Tuff (uncertainty is one standard deviation of the mean). A similar result is obtained by taking the weighted mean of all 56 crystal analyses: 18.72 ± 0.03 (s.e.m) and $\text{MSWD} = 1.8$.

Geochronology discussion

Ages of mineral standards that are used in $^{40}\text{Ar}/^{39}\text{Ar}$ age calculations vary amongst laboratories, and this difference in practice leads to accuracy uncertainties that

often exceed the quoted analytical precision. For example, the adopted age of a commonly used monitor, the Fish Canyon Tuff sanidine, may vary as much as 1.5 % between laboratories (27.6–28.0 Ma). Depending on the monitor calibration, ages determined in various laboratories differ by as much as 270,000 years for splits from the same 18 million-year-old tuff. Therefore, a valid comparison of $^{40}\text{Ar}/^{39}\text{Ar}$ age determinations requires that all be referred to a common monitor age. The problem is compounded in attempts to correlate $^{40}\text{Ar}/^{39}\text{Ar}$ ages of volcanic rocks with the Geomagnetic Polarity Time Scale (GPTS), the latest of which is calibrated by “astronomical tuning” (Lourens et al., 2004; Ogg and Smith, 2004). An earlier version of the GPTS (Cande and Kent, 1995) employed a small number of isotopically dated units as tie points to date polarity-reversal boundaries from sea floor magnetic anomalies.

Based on a different dating approach, the astronomical time scale (ATS) is calibrated via magnetostratigraphy of sedimentary sequences in which Earth’s orbital cycles (eccentricity, obliquity, and precession) are recognized. Climatic changes that track the orbital cycles cause biological, lithologic, and isotopic variations in the sediment.

The reversal boundaries of the ATS are calibrated primarily by matching the inferred cycles with an age model that is calculated from solar-system mechanics. For the early Miocene part of the time scale, gaps in the magnetostratigraphy were filled by combining sediment data with sea floor anomaly records. The age of the Oligocene-Miocene boundary is reduced to 23.03 Ma in the ATS of Lourens et al. (2004), compared to the boundary age of 23.8 Ma that was used in previous time scales.

For comparison of the $^{40}\text{Ar}/^{39}\text{Ar}$ age of the Peach Spring Tuff to the ATS, we follow Renne et al. (1994; 1998) who recommended using 28.02 Ma for the Fish Canyon Tuff sanidine to level the various monitors. This

Table 2. Model ages of the Peach Spring Tuff before and after adjustment to a common $^{40}\text{Ar}/^{39}\text{Ar}$ monitor age (Fish Canyon Tuff sanidine =28.02 Ma).

Location	Reported Ar/Ar Age Ma, 1 sd	Monitor - Age Ma	Recalculated: FCT=28.02 Ma, 1 sd	Reference
Kingman, AZ	18.5 ± 0.1 san	TCR - 27.88	18.8 ± 0.1	Nielson et al. (1990)
Little Piute Mtns, CA	18.42 ± 0.07 san	FCT - 27.84	18.5 ± 0.1	Miller et al. (1998)
Regional mean, AZ-CA Radiogenic Ar > 90%	18.71 ± 0.13 san	FCT - 27.84	18.8 ± 0.1	This study

Notes: FCT, Fish Canyon Tuff; TCR, Taylor Creek Rhyolite; san, sanidine

monitor age provides a better fit of $^{40}\text{Ar}/^{39}\text{Ar}$ ages with the ATS of Lourens et al. (2004) and Ogg and Smith (2004). Table 2 lists recalculated ages referred to this standard for $^{40}\text{Ar}/^{39}\text{Ar}$ studies of the Peach Spring Tuff. Recently, astronomically dated sediment and argon-dated tephra in Moroccan sedimentary sections support a modeled age of 28.201 ± 0.046 Ma for the Fish Canyon Tuff sanidine (Kuiper et al., 2008), and further revision the FCT monitor to an older age is under consideration by geochronologists.

Paleomagnetism methods

At each site, we used a portable diamond drill to collect 8 to 11 oriented cores distributed across several hundred square meters, mitigating possible errors from slumping or lightning. Azimuthal control for core orientations was determined by observations of sun angles. Structural attitudes of the welded tuff were measured on underlying sedimentary beds or, less frequently, on the compaction foliation defined by flattened pumice clasts.

Natural remanent magnetizations (NRM) were determined with a shielded, three-axis superconducting magnetometer. To remove secondary components of magnetization, two types of alternating-field devices were used. Three sites (0J001, 0J011, and 6J101) were given alternating-field demagnetization treatments in a 400 Hz, shielded coil with a reciprocating tumbler. Two pilot specimens were treated in progressively higher peak magnetic fields in a sequence of twelve steps to 100 mT (milliTesla). From orthogonal projections of the resultant magnetic vectors, we selected 4 steps from the coercivity range (30–80 mT) in which the magnetization direction remained stable. The final magnetization direction for each specimen was determined from the selected demagnetization steps by the least-squares, line-fitting method of Kirschvink (1980). For the other sites (9J803, 9J824, 9J835, 9J844, 9J855), all specimens were demagnetized through a progression of eight or nine steps up to 825 mT. This demagnetization device consisted of coils mounted in-line with the magnetometer, which operates in automatic mode. Line fitting was carried out as before after selecting demagnetization steps by visual inspection.

Paleomagnetism Results

In general, the alternating-field treatments produced well-defined magnetization directions at the specimen level and relatively little scatter at the site level. Site-mean directions and statistics are given in Table 3. One atypical result (Stoddard Valley) required the use of great-circle analysis (Kirschvink, 1980) to compensate for resistant secondary magnetization, apparently due to lightning.

Table 3. Directions of remanent magnetization, rotation parameters, and error estimates from new sampling of the Peach Spring Tuff, California, Arizona, and Nevada

Location Name	Site Id.	Site Lat	Site Long	Strike	Dip	N	Alpha-95	K	I	D	Ic	Dc	F ± DF	R ± DR
Alvord Mtn	0J001	35.069	243.432	55	8 SE	10	4.5	117	40.2	83.7	36.1	89.1	0.3 ± 4.5	56.1 ± 5.6
Halloran Hills	0J011	35.474	244.154	254	34 N	9	9.1	33	64.9	346.5	31.0	345.2	5.4 ± 7.8	-47.8 ± 9.2
Daggett Ridge	6J101A	34.81	243.01	259	23 N	7	6.7	82	44.8	20.9	24.1	13.3	12.3 ± 6.0	-19.7 ± 6.8
	6J101B	34.81	243.01	285	25 N	4	6.2	221	45.3	56.5	24.8	45.8	11.6 ± 5.7	12.8 ± 6.4
Outlet Center	9J803	34.812	242.910	0	0	8	4.1	186	35.4	26.7	35.4	26.7	1.0 ± 4.3	-6.3 ± 5.3
West Gem	9J824	34.807	243.116	275	41 N	8	14.4	17	14.8	135.8	37.7	117.9	-1.3 ± 11.8	86.5 ± 15.0
Stoddard Valley	9J835	34.748	243.119	191	6 W	8	13.3	*	40.3	15.8	39.0	11.1	-6.0 ± 11.0	-21.9 ± 14.2
Baxter Wash	9J844	34.987	243.712	65	10 S	9	5.2	97	36.9	55.3	37.9	63.0	-1.5 ± 5.0	30.0 ± 6.3
Harvard Hill	9J855	34.944	243.330	108	3 S	5	6.3	151	24.2	9.3	27.1	9.0	9.3 ± 5.7	-24.0 ± 6.1

Notes: N, number of specimens averaged; Alpha-95, 95% confidence limit, degrees; K, Fisher (1953) precision parameter; I, D, inclination and declination of in situ directions of magnetization, degrees; Ic, Dc, inclination and declination after correction for tilt of the beds, degrees; F ± DF, inclination flattening relative to Colorado Plateau reference direction (I=36.4°, D=33.0°, alpha-95=3.4°; Walls and Hillhouse, 1989) and 95% confidence limit, degrees; R ± DR, vertical-axis rotation (positive clockwise) relative to Colorado Plateau reference direction and 95% confidence limit, degrees.

The second exception is “Daggett Ridge,” which gave two groups of magnetization directions separated in declination by 32°. The different groups (designated as 6J101A and B) correspond to two areas of outcrop, 80 m apart and across an intervening gully. We conclude that the declination difference at this site is caused by slumping or drag adjacent to the Lenwood fault. Sediment filling the gully conceals any evidence of a structural discontinuity between the two outcrops.

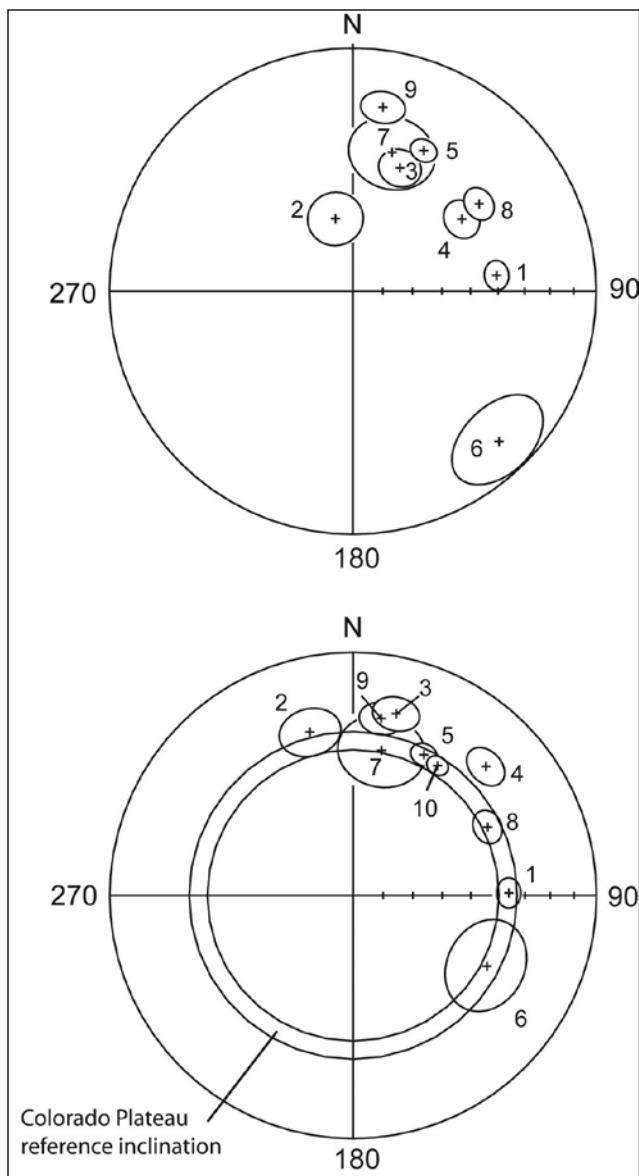


Figure 4. Lower hemisphere equal-area plots of in situ (upper diagram) and tilt-corrected directions (lower diagram) of magnetization, Peach Spring Tuff. Site means (+) with 95% confidence circles are shown. New paleomagnetism sites (Figure 2; Table 3) and Colorado Plateau reference direction: 1, Alvord Mt. (0J001); 2, Halloran Hills (0J011); 3, Daggett Ridge (6J101A); 4, Daggett Ridge (6J101B); 5, Outlet Center (8J803); 6, West Gem (9J824); 7, Stoddard Valley (9J835); 8, Baxter Wash, (9J844); 9, Harvard Hill (9J855); 10, Peach Spring Tuff: Colorado Plateau reference magnetization direction (Wells and Hillhouse, 1989).

Applying tilt corrections to the site means reduced the spread in magnetic inclination considerably, resulting in inclinations ranging from 24.1° to 39.0° (Figure 4). The inclinations are shallow relative to the Miocene dipole inclination (54.4°) and are generally consistent with the mean inclination of the Peach Spring Tuff (36.4°) as measured on the Colorado Plateau (Wells and Hillhouse,

1989). As shown in Figure 4, the 95% confidence circles of six sites intersect the annulus that is defined by the inclination confidence limit from the Plateau reference direction. The results from Harvard Hill and the two outcrops from “Daggett Ridge” give magnetization directions that are near the annulus, but with somewhat shallower inclinations by 9° to 12°. We favor upholding the correlation of Harvard Hill and Daggett Ridge with the Peach Spring Tuff, despite the relatively small inclination differences, because the correlation is supported by dating and at Stoddard Wash, zircon dating and chemistry at Harvard Hill (Leslie et al., this volume; Miller et al., this volume), mineralogy, and mapping near Daggett Ridge (Dibblee, 1970). Overall, the sites reveal large variations in declination relative to the reference declination (33.0°) from the Colorado Plateau, ranging from 48° counter-clockwise at Halloran Hills to 86° clockwise at West Gem. We interpret the declination differences as evidence for vertical-axis rotation, either by regional tectonism (crustal extension and strike-slip shear) or, in some cases, by drag near major faults.

Discussion

Hector Formation, northern Cady Mountains

Our paleomagnetic result from the rhyolite flow in Baxter Wash, northern Cady Mountains (Figure 2), confirms the proposed correlation with the Peach Spring Tuff and provides a revised control point for the magnetostratigraphy of the Hector Formation (MacFadden et al., 1990a; Woodburne, 1998). The Baxter Wash inclination is a good match to the Plateau reference inclination, and the declination, which indicates clockwise rotation of $30^\circ \pm 6^\circ$, is consistent with the rotation ($21^\circ \pm 8^\circ$) that MacFadden et al. (1990a) determined from the mean declination of the Hector Formation. The Peach Spring Tuff is near the top of a relatively thick normal-polarity magnetozone; it is overlain in turn by a thin reversed zone and another thick normal zone (Figure 5). We propose an age model that places the tuff just below the C6–C5E transition (18.748 Ma) of the astronomically tuned Geomagnetic Polarity Time Scale (ATS: Ogg and Smith, 2004). This model is the optimum fit to the time scale, considering the magnetozone relative thicknesses and the model $^{40}\text{Ar}/^{39}\text{Ar}$ age determinations (18.5–18.8 Ma) from the Peach Spring Tuff (Table 2). The next younger match to the C5E–C5D boundary (18.056 Ma) is less favorable, placing the tuff near 18.1 Ma and significantly younger than $^{40}\text{Ar}/^{39}\text{Ar}$ dating of the tuff in other areas. MacFadden et al. (1990a) proposed the younger correlation on the basis of the basalt age (18.6 ± 0.2 Ma, K–Ar, whole rock) and lower tuff (22.9 ± 0.4 Ma; K–Ar, biotite). Below

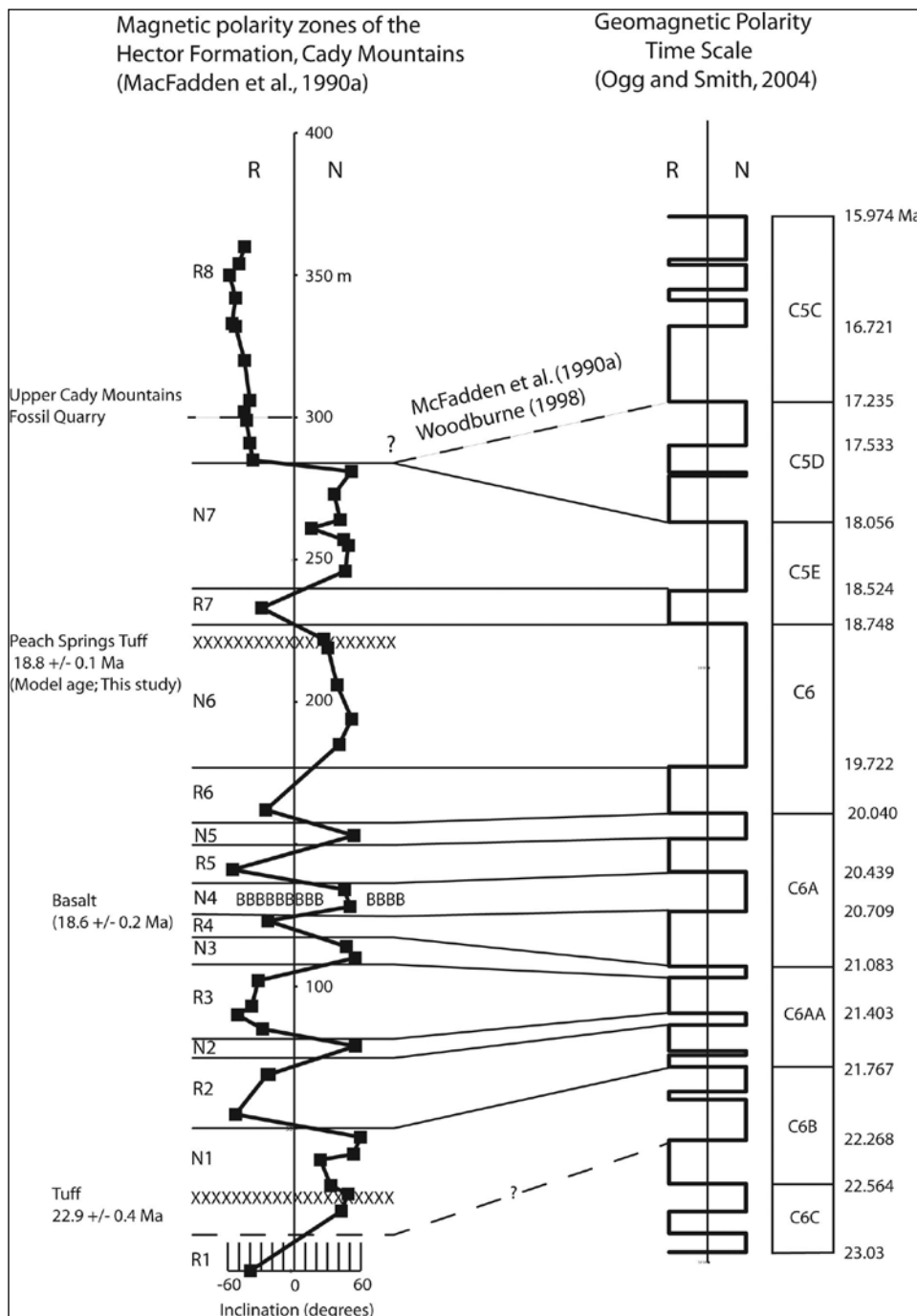


Figure 5. Reinterpreted model for correlation of the Hector Formation magnetostratigraphy, northern Cady Mountains, with the astronomically tuned Geomagnetic Polarity Time Scale.

the Peach Spring Tuff, the alternating pattern of thin magnetozone matches reasonably well with the ATS from 19.7 Ma to 22.5 Ma and with the K–Ar age of the lower tuff. However, the whole-rock age of the intervening basalt (18.6 ± 0.2 Ma) does not fit our proposed correlation model, being ~ 1.8 m.y. younger than the proposed model age. Woodburne (1998) reached the same conclusion that the basalt age was underestimated through regional analysis of Arikareean mammalian fossils and isotopic

dating in the type Hector Formation (Woodburne et al., 1974). He proposed that the basalt correlated to the normal magnetozone of C6, between 19 Ma and 20 Ma. Our interpretation places the basalt in the middle of C6A or approximately 20.5 Ma.

A rich assemblage of Late Hemingfordian vertebrate fossils occurs in sediments in the uppermost reversed-polarity magnetozone (R8: Figure 5) of the Hector Formation, and this assemblage provides a temporal link to the Barstow Formation in the region. MacFadden et al. (1990a) noted the similarity of these fossils to ones found in the lower reversed-polarity magnetozone (R2) of the Barstow Formation, 50 km northwest in the Mud Hills (Figure 2). Two $^{40}\text{Ar}/^{39}\text{Ar}$ -dated tuffs within the Barstow Formation constrain the age of the Late Hemingfordian fossils between 16.56 ± 0.34 Ma (Rak Tuff) and 19.3 ± 0.02 Ma (Red Tuff: MacFadden et al., 1990b). Dating of the Barstow Formation in the Mud Hills favors correlation of the reversed-polarity magnetozone (R2 of the Barstow Formation from MacFadden et al., 1990b) with the bottom part of C5C (ATS: 16.721–17.235 Ma).

Our interpretation of the Hector Formation magnetostratigraphy places the uppermost fossil-bearing reversed magnetozone (R8) in C5D, approximately 0.8 m.y. older than the placement proposed by MacFadden et al. (1990a) and Woodburne (1998). We include a queried alternative correlation to the younger magnetozone (Figure 5), if in fact, R2 of the Barstow Formation (Owl Conglomerate Member) correlates with R8 of the Hector Formation. If the younger correlation is correct, a hiatus or much reduced deposition rate is

implied between the Peach Spring Tuff and magnetozones R8 of the Hector Formation.

Tectonic rotations

The central Mojave Desert (Figure 2) was subject to two main tectonic modes during the Neogene: extension and tilting of crustal blocks, sometimes above detachment faults, followed later by strike-slip faulting (Dokka, 1989). Within the Eastern California Shear Zone (ECSZ), many Tertiary stratigraphic sections show compelling evidence for extensional tectonics prior to the emplacement of

the Peach Spring Tuff (Miller, 1994; Cox, 1995; Glazner et al., 2002), although Gans et al. (2005) have modeled thermal evolution of the central Mojave metamorphic core complex as indicating extension continued to about 17.5 Ma. Early Miocene (~24 Ma to 22 Ma) volcanism accompanied development of extensional basins across a broad area between Barstow and the eastern margin of the ECSZ (Glazner et al., 2002). Volcanism farther north, in the Fort Irwin and China Lake military areas, was younger (16 Ma to 12 Ma) and was not accompanied by significant extension (Schermer et al., 1996). Subsequent to the period of tectonic extension, probably during late

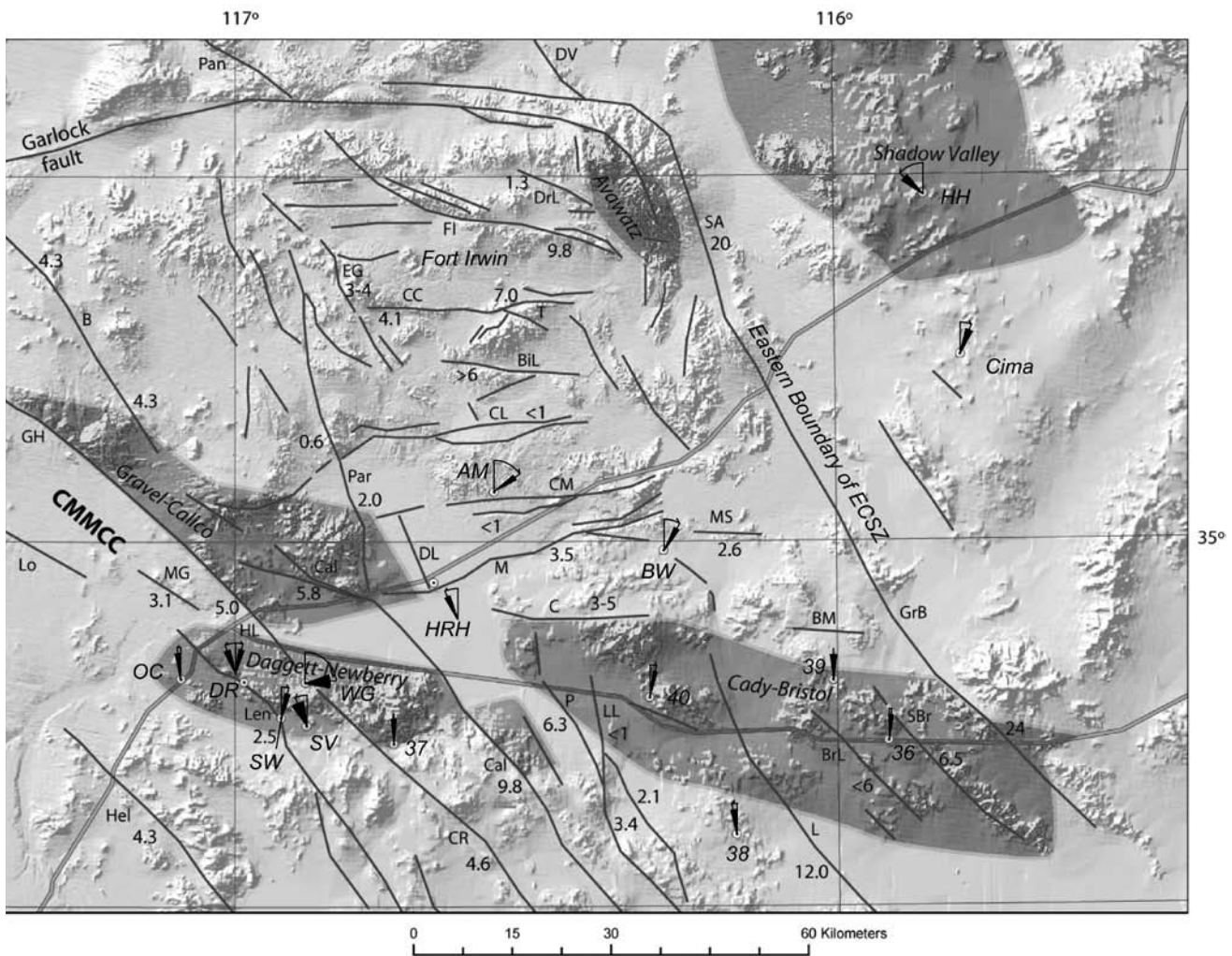


Figure 6. Map of Quaternary faults (after Miller et al., 2007) showing fault offsets in km (from Jachens et al. (2002), Schermer et al. (1996), Pavlis et al. (1998), Oskin and Iriondo (2004), and unpublished mapping by Miller, K.M. Schmidt, and V.E. Langenheim (2009); and general locations of large early Miocene extensional basins in gray (blue) shades (all except the Shadow Valley area, which is middle and late Miocene). Vertical-axis rotations determined by paleomagnetism of the Peach Spring Tuff from this study, Wells and Hillhouse (1989), and Hillhouse and Wells (1991) are shown with 95% confidence fans in black with labels as in Figure 2. Fault names: B, Blackwater; BiL, Bicycle Lake; BM, Broadwell Mesa; BrL, Broadwell Lake; Bw, Blackwater; C, Cady; Cal, Calico; CC, Coyote Canyon; CL, Coyote Lake; CM, Cave Mountain; CR, Camp Rock; DL, Dolores Lake; DrL, Drinkwater Lake; DV, Death Valley; EG, East Goldstone; FI, Fort Irwin; G, Garlock; GH, Gravel Hills; GrB, Granite-Bristol; Hel, Helendale; HL, Harper Lake; L, Ludlow; Len, Lenwood; LL, Lavic Lake; Lo, Lockhart; M, Manix; MG, Mt. General; MS, Mesquite Spring; P, Pisgah; Pan, Panamint; Par, Paradise; SA, Soda-Avawatz; SBr, South Bristol; T, Tiefort Mountain. CMMCC, Central Mojave Metamorphic Core Complex.

Miocene to early Pliocene time, strike-slip faulting of the ECSZ commenced. There are two domains of strike-slip faulting in our area of interest, one governed by east-striking sinistral faults, and the other by northwest-striking dextral faults (Fig. 2 and 6). The strike-slip faulting continues today, manifested as belts of active faults within a much broader area of dormant Pleistocene faults (Miller et al., 2007).

Both tectonic modes, extension and strike-slip faulting, produced vertical-axis rotations that are indicated by paleomagnetism. Ross et al. (1989) and Ross (1995) reported significant clockwise rotations in moderately tilted volcanic rocks of early Miocene age in the Daggett-Newberry and Cady-Bristol areas (Figure 6). The clockwise-rotated domain that Ross et al. (1989) discovered in the Newberry, Rodman, Lava Bed, southern Cady, and Bristol Mountains was interpreted by Dokka (1989) as part of an extensional complex, designated the "Daggett terrane." In these same areas, the Peach Spring Tuff is generally tilted less and shows little vertical-axis rotation except for sites that are very close to the Lenwood and Camp Rock faults (Wells and Hillhouse, 1989). A prime example of this tectonic relationship occurs in the Newberry Mountains near Barstow. Based on geologic mapping and paleomagnetic studies near Minneola Ridge and Kane Wash, Hillhouse et al. (this volume) reported clockwise rotation (43°) in moderately tilted, 22-19 million-year-old volcanic rocks that are capped by subhorizontal and essentially unrotated (CCW: $1.6^\circ \pm 4.3^\circ$) Peach Spring Tuff. Similar relations are evident in the Bristol Mountains, where the Peach Spring Tuff and associated strata angularly overlie steeply tilted volcanic rocks, 23 Ma to 20 Ma (Miller et al., 1994). Ross et al. (1989) described clockwise rotations of 30° in the tilted volcanic strata of the Bristol Mountains. In contrast, rotation of the Peach Spring Tuff is 4° or less in the Bristol Mountain block, which is cut by the dextral and northwest-striking Broadwell Lake, South Bristol, and Granite-Bristol faults (Figure 6).

Before discussing the sinistral-fault rotations, we turn to the issue of drag-induced rotations in fault zones. Wells and Hillhouse (1989) postulated that shear adjacent to the Lenwood fault was responsible for the 13° of clockwise rotation observed at the Stoddard Wash site, an observation consistent with the easterly progression of strikes shown by tuff-bearing, en echelon ridges near the fault (Dibblee, 1970). Similarly, the study by Oskin et al. (2007) showed distributed shear within 0.5 km of the Calico fault. Several of the new sites we report may record rotations due to fault-proximal shear: our Daggett Ridge site lies within a few hundred meters of the Lenwood fault zone and Harvard Hill is in an uplifted block showing complex deformation (Leslie et al., this volume). In ad-

dition, the West Gem site is a steeply dipping lacustrine section that lies about 2 km from folded Miocene strata associated with the termination of the Camp Rock fault; its clockwise (87°) rotation may be due to non-cylindrical folding.

Our paleomagnetic measurements in blocks bounded by east-striking sinistral faults show clockwise rotations similar to values published by Ross et al. (1989) for Alvord Mountain and by MacFadden et al. (1990a) for Baxter Wash, northern Cady Mountains. To provide a context for discussing the rotation kinematics, Figure 6 shows a compilation of Quaternary faults (Miller et al. 2007), vertical-axis rotations measured on tectonic blocks, and published offset estimates for faults bounding the blocks. This fault map differs from previous maps, such as Jennings (1994), in showing a few newly discovered Quaternary faults: the Paradise fault system bounds the sinistral blocks on the west, the Cave Mountain fault (noted in part by Glazner et al., 2002, as their Afton fault) lies between the Coyote Lake and Manix faults. Two 5–10 km long east-striking faults lie well to the east of northwest-striking faults in the region south of Soda Lake: the Mesquite Springs and Broadwell Mesa faults. These newly mapped faults clarify geometries of fault blocks and add a few complexities for fault models.

Schermer et al. (1996) proposed a simple rigid block model (bookshelf faulting) to explain the rotational kinematics of the sinistral-fault domain of the northeast Mojave Desert. Starting with northeast trending faults, the blocks rotate clockwise while being subjected to dextral shear to reach the current east-striking fault pattern. Schermer et al. (1996) measured clockwise rotation averaging 64° in middle Miocene volcanic rocks bounded by the Bicycle Lake and Fort Irwin sinistral faults (Figure 6). They noted that measured fault offsets accounted for about a third of the paleomagnetically-inferred rotation, and appealed to ductile deformation near the block ends to explain the excess rotation.

According to the rigid-block, bookshelf model of Schermer et al. (1996), the width of the tectonic block, rotation of the block, and fault offsets are related geometrically if there is no strain within the blocks. From the perspective of an observer at the Cima site, the tectonic blocks near Barstow move north and west with time, and the Fort Irwin area undergoes clockwise tectonic rotation and elongates east-west with time. This rotation is balanced by bending of the Garlock fault and by shear along dextral faults north of the Garlock and south of the Cady Mountains. East-west extension north of the Garlock fault balances the lengthening caused by block rotation south of the fault, but similar extension is not known south of the Cady Mountains. The only major Miocene and

younger extensional basin is Bristol Lake basin; it may accommodate part or all of the east extension required. If the model holds, all fault panels in the sinistral domain must rotate by similar amounts, because the faults, with the exception of the Manix fault, all strike due east. As a result, fault offset is directly proportional to block width. However, our new sites in the Cady Mountains and Alvord Mountain validate previous work by MacFadden et al. (1990a) and Ross et al. (1989) in showing that vertical-axis rotation is about 25° less in the Cady Mountains compared to all tectonic blocks farther north. Fault offsets are comparable throughout the sinistral domain, although they remain poorly determined in several places. The average rotation of northern blocks is about 65°; this decreases to 30° at the Baxter Wash site and to 12° at Sleeping Beauty in the south margin of the Cady Mountains. The latter site may have some fault-proximal strain if the steep linear front of the Cady Mountains is tectonic, but no fault has been mapped by several generations of geologists.

The puzzles associated with the southern part of the sinistral domain are several: (1) The southward decrease in vertical-axis rotation is not mirrored by progressive change in bounding fault strikes; (2) The Cave Mountain, Manix, and Cady faults show evidence for Holocene activity (Miller et al., 2007), while faults farther north do not (Schermer et al., 1996; Miller and Yount, 2002), suggesting that all faults are not equally active; (3) East-striking faults (Mesquite Springs and Broadwell Mesa) lie east of the sinistral domain, where they are bounded on all sides by northwest-striking dextral faults, apparently in violation of the Schermer model. Our new paleomagnetic data help to constrain models that account for these puzzles, but more data are needed. Our geologic mapping in progress indicates that small (10 km x 10 km) tectonic blocks lie between the sinistral and dextral domains, and in these areas faults are probably dynamically forming and being abandoned to accommodate space changes and overall shear. In addition, the east-striking sinistral faults appear to be a stable mode of faulting; faults in this orientation may not initiate as northeast-striking faults and rotate to east strike.

Geodetic observations of the current deformation in the sinistral-fault domain provide clues to the Neogene tectonic history of the region. Using the GPS network at Fort Irwin, Savage et al. (2004) investigated interseismic strain and rotation in the area of east-striking sinistral faults. They concluded that bookshelf strain on east-trending faults is not a suitable release mechanism for the current strain rates. Instead, they found that the Fort Irwin velocity field is best explained by clockwise rotation of the northeast Mojave block as a whole with strain

release occurring on the northwest-trending bounding faults. The rate of rotation, 4°/m.y., is sufficient to accumulate the paleomagnetically-derived rotation amounts if the same rate persisted since middle Miocene time. Density of the GPS network is not sufficient to investigate individual block movements, however, and rather represents the average kinematic picture for a broad region.

We analyzed paleomagnetism of the Peach Spring Tuff at two sites east of the ECSZ: Halloran Hills and Cima. At Halloran Hills, the result shows rotation of 48° counter-clockwise, and the beds tilt 34° to the north. This site lies within the strongly extended hanging wall of the Shadow Valley detachment system (Davis et al., 1991; Davis and Friedmann, 2005; Reynolds, 1993) that formed about 14 to 12 Ma. Although possibility of strike-slip faulting associated with or postdating extension (Prave and McMackin, 1999) has not been entirely discounted, we consider that the significant vertical-axis rotation of this site was aided by the shallow decollement surface beneath Shadow Valley. The tectonic situation may be similar to the Colorado River extensional corridor, where significant vertical-axis rotation, as well as steep tilt, affected the Peach Spring Tuff in hanging walls above major detachment faults (Howard and John, 1987; Wells and Hillhouse, 1989).

South of I-15 in the Cima area, where extension and tilt of Miocene sediments are more subtle, a detachment fault has not been demonstrated. Nevertheless, the Cima site declination indicates $21.2^\circ \pm 8.0^\circ$ of clockwise rotation, which affected a relatively small block that dips 40° southeastward (Hillhouse and Wells, 1991). Quaternary alluvium, basalt, and cinder cones obscure the contact relationships and structural context of the tuff-bearing block.

Conclusions

Application of single-crystal $^{40}\text{Ar}/^{39}\text{Ar}$ dating methods at eight sites distributed across the broad extent of the Peach Spring Tuff yields a regional mean age of 18.7 ± 0.1 Ma. The dated material consisted of clear sanidine crystals handpicked from crushed whole specimens that were used in earlier paleomagnetic studies of the ash-flow tuff. Adjusting the mean age to a common monitor standard age (FCT-sanidine = 28.02 Ma) gives a model age of 18.8 Ma. The emplacement age reported by Nielson et al. (1990) from sanidine in juvenile pumice also adjusts to 18.8 Ma, after the monitor age is set to 28.02 Ma. Regional correlation of many isolated outcrops to the Peach Springs eruption, as originally proposed by Glazner et al. (1986), is strengthened by the remarkably consistent $^{40}\text{Ar}/^{39}\text{Ar}$ ages measured at sites from Kingman to

Barstow. In particular, dating of the Stoddard Wash and Kane Wash localities confirms that “pink tuffs” (unit Tst of Dibblee, 1964, 1970) surrounding Daggett Ridge near Barstow are part of the Peach Spring Tuff, and paleomagnetic inclination data from the tuff outcrops are also consistent with the correlation.

A rhyolitic ash-flow tuff in the middle part of the Hector Formation, northern Cady Mountains, is confirmed by paleomagnetism to be part of the Peach Spring Tuff. The tuff is now firmly tied to the magnetostratigraphy of the Hector Formation (MacFadden et al., 1990a; Woodburne, 1998), prompting reinterpretation of the correlation of Hector magnetozones with the astronomically tuned Geomagnetic Polarity Time Scale (Ogg and Smith, 2004). Our favored model, based on the isotopic age and normal polarity of the Peach Spring Tuff, places the C6–C5E polarity boundary (18.75 Ma) a few meters above the ash bed.

The pattern and timing of vertical-axis rotations in the central Mojave Desert is becoming better understood with the addition of new paleomagnetic data. Clockwise rotation of the Miocene Spanish Canyon Formation at Alvord Mountain, originally proposed by Ross et al. (1989), is determined to be $56^\circ \pm 6^\circ$ from our paleomagnetic sampling of the Peach Spring Tuff. Further south at Baxter Wash in the Cady Mountains, our paleomagnetic declination measurement from the tuff indicates clockwise rotation of $30^\circ \pm 6^\circ$. This measurement is consistent with the rotation estimate ($21^\circ \pm 8^\circ$) given by MacFadden et al. (1990a) for the same stratigraphic section, using the mean declination of the Hector Formation. Within the Eastern California Shear Zone, significant vertical-axis rotations of blocks younger than the Peach Spring Tuff are confined to the region of east-striking sinistral faults. Clockwise rotation of these blocks has been ascribed to right-lateral shear within the ECSZ. However, the amounts of rotation are not uniform; instead, we have measured progressively larger rotations from the southern Cady Mountains (12° CW) northward to the Alvord Mountains (56° CW). Clearly, kinematic models of the region, such as the rigid-block model of Schermer et al. (1996), require further refinements to explain non-uniform rotation of the northeast Mojave domain and the non-uniform fault activity, both in terms of timing and displacement.

Extensional faulting is another mechanism that has produced large-scale vertical-axis rotation in the Mojave Desert. Early paleomagnetic studies (Ross et al., 1989; Ross, 1995) showed that tilted lower Miocene sections, presumably underlain by detachment faults, underwent clockwise rotations on the order of 30° to 60° . Most of the data are from the Cady–Bristol and Daggett–Newberry

areas (Figure 6), where the rotations are constrained to have occurred before deposition of the Peach Spring Tuff. A more recent example of the paleomagnetic evidence for this tectonic relationship is presented by Hillhouse et al. (this volume) for the Newberry Mountains. Extension is much younger in the Shadow Valley basin, where we measured a large counterclockwise rotation in the upper plate above the Kingston Range–Halloran Hills detachment fault of Davis and Friedmann (2005). Although the mechanisms for the rotations during extension are not well understood, components of shear across the region during extension (Bartley and Glazner, 1991) and dismemberment of the hanging wall into small blocks could produce the rotations.

Acknowledgments

The authors are indebted to Ray Wells (USGS) for his contributions during the early phase of this study, and for reviewing this manuscript. Victoria Pease (Stockholm University) provided valuable assistance with the fieldwork and paleomagnetism measurements. We also wish to thank Shannon Leslie for geologic guidance and help during recent fieldwork. Over the years we have benefited greatly from the geologic expertise generously offered by Keith Howard, Jane Nielson, and Robert Reynolds in our Mojave studies. We also thank Ed Mankinen (USGS) for his comments on improving the original manuscript.

References

- Bartley, J.M., and Glazner, A.F., En-echelon Miocene rifting in the southwestern United States and model for vertical-axis rotations in continental extension, *Geology*, 19, 1165–1168, 1991.
- Beckerman, G. M., J. P. Robinson, and J. L. Anderson, The Teutonia batholith: A large intrusive complex of Jurassic and Cretaceous age in the eastern Mojave Desert, California, in Frost, E. G., and D. L. Martin, eds., *Mesozoic-Cenozoic tectonic evolution of the Colorado River region, California, Arizona, and Nevada*, San Diego, Calif., Cordilleran Publishers, 205–220, 1982.
- Buesch, D. C., Feldspar geochemistry of four Miocene ignimbrites in southeastern California and western Arizona, in D. R. Sherrard, and J. E. Nielson, eds., *Tertiary stratigraphy of highly extended terranes, California, Arizona, and Nevada*, U. S. Geol. Surv. Bull. 2053, 55–69, 1994.
- Buesch, D. C., and G. A. Valentine, Peach Springs Tuff and volcanic stratigraphy of the southern Cerbat Mountains, Kingman, Arizona, in Nielson, J. E., and A. F. Glazner, eds., *Cenozoic stratigraphy, structure, and mineralization in the Mojave Desert*, Geol. Soc. Am. Guidebook, Cordilleran Section, Field Trip 5, 7–14, 1986.
- Byers, F. M., *Geology of the Alvord Mountain Quadrangle, San Bernardino County, California*: U. S. Geol. Surv. Bull. 1089-A, 71 p., 1960.
- Cande, S. C., and D. V. Kent, Revised calibration of the geomagnetic polarity timescale for the Late Cretaceous and Cenozoic, *J. Geophys. Res.*, 100, 6093–6095, 1995.
- Cox, B.F., Structure and tectonics at the south edge of an Early Miocene rift zone, central Mojave Desert, California, *San Bernardino County Museum Association Quarterly*, 42, 30, 1995.

- Davis, G. A., and S. J. Friedmann, Large-scale gravity sliding in the Miocene Shadow Valley supradetachment basin, eastern Mojave Desert, California, *Earth Sci. Reviews*, 73, 149-176, 2005.
- Davis, G. A., M. A. Parke, K. M. Bishop, T. K. Fowler, and S. J. Friedmann, Grand scale detachment and emplacement of gravity-driven slide sheets into a Miocene terrestrial basin, eastern Mojave Desert, California, *Geol. Soc. Am. Abstr. with Prog.*, 23, A47, 1991.
- Davis, G. A., T. K. Fowler, K. M. Bishop, T. C. Brudos, S. J. Friedmann, D. W. Burbank, M. A. Parke, and B. C. Burchfiel, Pluton pinning of an active Miocene detachment fault system, eastern Mojave Desert, California, *Geology*, 21, 627-630, 1993.
- Dibblee, T. W., Jr., Geologic map of the Barstow quadrangle, San Bernardino County, California, U. S. Geol. Surv. Min. Inv. Field Stud. Map, MF-233, scale 1:62,500, 1960.
- Dibblee, T. W., Jr., Geologic map of the Ord Mountains quadrangle, San Bernardino County, California, U. S. Geol. Surv. Misc. Geol. Inv. Map, I-427, scale 1:62,500, 1964.
- Dibblee, T. W., Jr., Geologic map of the Daggett quadrangle, San Bernardino County, California, U. S. Geol. Surv. Misc. Geol. Inv. Map, I-592, scale 1:62,500, 1970.
- Dibblee, T. W., Jr., and A. M. Bassett, Geologic map of the Newberry quadrangle, San Bernardino County, California, U. S. Geol. Surv. Misc. Geol. Inv. Map, I-461, scale 1:62,500, 1966.
- Dokka, R. K., The Mojave extensional belt of southern California, *Tectonics*, 8, 363-390, 1989.
- Fisher, R. A., Dispersion on a sphere, *Proc. Roy. Soc. London*, A217, 295-305, 1953.
- Gans, P., DeVicchio, E., Singleton, J., Van Pelt, J., Wong, M., Reynolds, J., Cenozoic magmatic and structural evolution of the central Mojave Desert, California: New constraints from $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology and thermochronology, *Geol. Soc. Am. Abstr. with Programs*, 37(4), 103, 2005.
- Glazner, A. F., J. E. Nielson, K. A. Howard, and D. M. Miller, Correlation of the Peach Springs Tuff, a large-volume Miocene ignimbrite sheet in California and Arizona, *Geology*, 14, 840-843, 1986.
- Glazner, A. F., Walker, J. D. Bartley, J. M., and Fletcher, J. M., Cenozoic evolution of the Mojave block of southern California, *Geol. Soc. Am. Memoir* 195, 19-41, 2002.
- Gusa, S., J. E. Nielson, and K. H. Howard, Heavy-mineral suites confirm the wide extent of the Peach Springs Tuff in California and Arizona, USA, *J. Volc. Geotherm. Res.*, 33, 343-347, 1987.
- Hewett, D. F., Geology and mineral resources of the Ivanpah quadrangle, U. S. Geol. Surv. Prof. Pap. 275, 1956.
- Hillhouse, J. W., and R. E. Wells, Magnetic fabric, flow directions, and source area of the lower Miocene Peach Springs Tuff in Arizona, California, and Nevada, *J. Geophys. Res.*, 96, 12,443-12,460, 1991.
- Hillhouse, J. W., R. E. Wells, and B. F. Cox, Paleomagnetism of Miocene volcanic rocks in the Newberry Mountains, California: Vertical-axis rotation and a polarity transition, this volume.
- Howard, K. A., and B. E. John, Crustal extension along a rooted system of imbricate low angle faults: Colorado River extensional corridor, California and Arizona, In Coward, M. P., Dewey, J. F., and Hancock, P. L., eds., *Continental extensional tectonics*, *Geol. Soc. London Spec. Pub.* 28, 299-311, 1987.
- Jachens, R. C., V. E. Langenheim, and J. C. Matti, Relationship of the 1999 Hector Mine and 1992 Landers fault ruptures to offsets on Neogene faults and distribution of late Cenozoic basins in the eastern California shear zone, *Bull. Seis. Soc. Am.*, 92, 1592-1605, 2002.
- Jennings, C.W., Fault activity map of California and adjacent areas with location and ages of volcanic eruptions, California Geologic Data Map Series, Map No. 6, California Division of Mines and Geology, scale 1:750,000, 1994.
- Kirschvink, J. L., The least-squares line and plane and the analysis of palaeomagnetic data, *Geophys. J. R. Astron. Soc.*, 62, 699-718, 1980.
- Kuiper, K. F., A. Deino, F. J. Hilgen, W. Krijgsman, P. R. Renne, and J. R. Wijbrams, Synchronizing rock clocks of Earth history, *Science*, 320, 500-504, 2008.
- Leslie, S.R., D. M. Miller, J. L. Wooden, and J. A. Vazquez, Stratigraphy, age, and depositional setting of the Miocene Barstow Formation at Harvard Hill, central Mojave Desert, California, this volume.
- Lourens, L., F. Hilgen, N. J. Shackleton, J. Laskar, and D. Wilson, The Neogene Period, in *A Geologic Time Scale 2004*, eds. F. M. Gradstein, J. G. Ogg, and A. G. Smith, Cambridge University Press, 409-452, 2004.
- Luyendyk, B. P., M. J. Kamerling, and R. Terres, Geometric model for Neogene crustal rotations in southern California, *Geol. Soc. Am. Bull.*, 91, 211-217, 1980.
- MacFadden, B. J., M. O. Woodburne, and N. D. Opdyke, Paleomagnetism and Neogene clockwise rotation of the northern Cady Mountains, Mojave Desert of southern California, *J. Geophys. Res.*, 95, 4597-4608, 1990a.
- MacFadden, B. J., C. C. Swisher, III, N. D. Opdyke, and M. O. Woodburne, Paleomagnetism, geochronology, and possible tectonic rotation of the middle Miocene Barstow Formation, Mojave desert, southern California, *Geol. Soc. Am. Bull.*, 102, 478-493, 1990b.
- Miller, D.M., Cenozoic deposits in the Lava Hills and southern Bristol Mountains, southeastern California, in Sherrod, D.R., and Nielson, J.E., eds., *Tertiary Stratigraphy of highly extended terranes, California, Arizona, and Nevada*: U. S. Geol. Surv. Bull. 2053, 99-107, 1994.
- Miller, D.M., and Yount, J.L., Late Cenozoic tectonic evolution of the north-central Mojave Desert inferred from fault history and physiographic evolution of the Fort Irwin area, California: *Geol. Soc. Am. Memoir* 195, 173-197, 2002.
- Miller, D.M., Dudash, S.L., Green, H.L., Lidke, D.J., Amoroso, L., Phelps, G.A., and Schmidt, K.M., A new Quaternary view of northern Mojave Desert tectonics suggests changing fault patterns during the late Pleistocene, in Miller, D.M. and Valin, Z.C., eds., *Geomorphology and tectonics at the intersection of Silurian and Death Valleys, southern California*, U. S. Geol. Surv. Open-File Report 2007-1424, 157-171, 2007.
- Miller, D. M., S. R. Leslie, J. W. Hillhouse, J. L. Wooden, J. A. Vazquez, and R. E. Reynolds, Reconnaissance geochronology of tuffs in the Miocene Barstow Formation; implications for basin evolution and tectonics in the central Mojave Desert, this volume.
- Miller, J. S., M. T. Heizler, and C. F. Miller, Timing of magmatism, basin formation, and tilting at the west edge of the Colorado River Extensional Corridor: Results from single-crystal Ar/Ar geochronology of Tertiary rocks in the Old Woman Mountains area, southeastern California, *J. Geol.*, 106, 195-209, 1998.
- Miller, S.T., Geology and mammalian biostratigraphy of a part of the northern Cady Mountains, Mojave Desert, California: U. S. Geol. Surv. Open File Report 80-878, 122 p., 1980.
- Nielson, J. E., and K. K. Beratan, Stratigraphic and structural synthesis of a Miocene extensional terrane, southeast California and west-central Arizona, *Geol. Soc. Am. Bull.*, 107, 241-252, 1995.
- Nielson, J. E., and A. F. Glazner, Miocene stratigraphy and structure: Colorado Plateau to the central Mojave Desert, in Nielson, J. E., and A. F. Glazner, eds., *Cenozoic stratigraphy, structure, and mineralization in the Mojave Desert*, *Geol. Soc. Am. Guidebook, Cordilleran Section, Field Trip 5*, 1-6, 1986.
- Nielson, J. E., D. R. Lux, G. B. Dalrymple, and A. F. Glazner, Age of the Peach Springs Tuff, southeastern California and western Arizona, *J. Geophys. Res.*, 95, 571-580, 1990.

- Ogg, J. G., and A. G. Smith, The geomagnetic polarity time scale, in A Geologic Time Scale 2004, edited by F. M. Gradstein, J. G. Ogg, and A. G. Smith, Cambridge University Press, 63-86, 2004.
- Oskin, M. and Iriondo, A., Large magnitude transient strain accumulation on the Blackwater fault, Eastern California shear zone, *Geology*, 32(4), 313-316, 2004.
- Oskin, M., Perg, L., Blumentritt, D., Mukhopadhyay, S., Iriondo, A., Slip rate of the Calico fault: Implications for geologic versus geodetic rate discrepancy in the Eastern California Shear Zone, *J. Geophys. Res.*, 112, B03402, doi:10.1029/2006JB004451, 2007.
- Pavlis, T. L., L. Serpa, B. Troxel, M. Dean, T. Hartman, and T. Rodosta, Late Cenozoic deformation in eastern Fort Irwin and its significance for the tectonic history of the Garlock fault system, *San Bernardino County Museum Association Quarterly*, 45(2), 77-83, 1998.
- Prave, A. R., and M. R. McMackin, Depositional framework of mid- to late Miocene strata, Dumont Hills and southern margin Kingston Range: Implications for the tectonostratigraphic evolution of the southern Death Valley region, *Geol. Soc. Am. Spec. Pap.* 333, 259-275, 1999.
- Renne, P. R., A. L. Deino, R. C. Walter, et al., Intercalibration of astronomical and radioisotopic time, *Geology*, 22, 783-786, 1994.
- Renne, P. R., C. C. Swisher, A. L. Deino, D. B. Karner, T. L. Owens, and D. J. DePaolo, Intercalibration of standards, absolute ages and uncertainties in $^{40}\text{Ar}/^{39}\text{Ar}$ dating, *Chem. Geol.*, 145, 117-152, 1998.
- Reynolds, R. E., Erosion, deposition, and detachment: The Halloran Hills area, Calif., in D. R. Sherrod and J. E. Nielson, eds., *Tertiary Stratigraphy of Highly Extended Terranes, California, Arizona, and Nevada*, U. S. Geol. Surv. Bull. 253, 21-24, 1993.
- Reynolds, R. E., and M. R. McMakin, Field trip roadlog, Cenozoic tectonics in the Halloran Hills and Kingston Range, in D. L. Weide and M. L. Faber, eds., *This Extended Land: Geological Journeys in the Southern Basin and Range*, *Geol. Soc. Am., Cordilleran Section meeting, Las Vegas, Nevada, Field Trip Guidebook*, 201-206, 1988.
- Reynolds, R. E., and M. A. Nance, Shadow Valley Basin: Late Tertiary deposition and gravity slides from the Mescal Range, in D. L. Weide and M. L. Faber, eds., *This Extended Land: Geological Journeys in the Southern Basin and Range*, *Geol. Soc. Am., Cordilleran Section meeting, Las Vegas, Nevada, Field Trip Guidebook*, 207-209, 1988.
- Ross, T. M., North-south-directed extension, timing of extension, and vertical-axis rotation of the southwest Cady Mountains, Mojave desert, California, *Geol. Soc. Am. Bull.*, 107, 793-811, 1995.
- Ross, T. M., B. P. Luyendyk, and R. B. Haston, Paleomagnetic evidence for Neogene clockwise tectonic rotations in the central Mojave Desert, California, *Geology*, 17, 470-473, 1989.
- Savage, J. C., J. L. Svarc, and W. H. Prescott, Interseismic strain and rotation rates in the northeast Mojave domain, eastern California, *J. Geophys. Res.*, 109, B020406, doi:10.1029/2003JB002705, 2004.
- Schermer, E. R., B. P. Luyendyk, and S. Cisowski, Late Cenozoic structure and tectonics of the northern Mojave Desert, *Tectonics*, 15, 905-932, 1996.
- Spell, T. L., E. I. Smith, A. Sanford, and K. A. Zanetti, Systematics of xenocrystic contamination: preservation of discrete feldspar populations at McCullough Pass caldera revealed by $^{40}\text{Ar}/^{39}\text{Ar}$ dating, *Earth Planet. Sci. Lett.*, 190, 153-165, 2001.
- Steiger, R. H., and Jaeger, E., Subcommission on Geochronology: convention on the use of decay constants in geo- and cosmo-chronology, *Earth Planet. Sci. Lett.*, 36, 359-362, 1977.
- Turrin, B. D., J. C. Dohrenwend, R. E. Drake, and G. H. Curtis, K-Ar ages from the Cima volcanic field, eastern Mojave Desert, California, *Isochron/West*, 44, 9-16, 1985.
- Varga, R. J., A. J. Horst, N. Welty, K. Siegner, and P. Ferringner, Paleomagnetism of the Peach Springs Tuff revisited, *Geol. Soc. Am. Abstr. with Prog.*, 38(7), 352, 2006.
- Wells, R. E., and J. W. Hillhouse, Paleomagnetism and tectonic rotation of the lower Miocene Peach Springs Tuff: Colorado Plateau, Arizona, to Barstow, California, *Geol. Soc. Am. Bull.*, 101, 846-863, 1989.
- Woodburne, M. O., The Mojave Desert Province, in *Inland Southern California: the last 70 million years*, *San Bernardino County Museum Association Quarterly*, 38, 60-78, 1991.
- Woodburne, M. O., Arikareean and Hemingfordian faunas of the Cady Mountains, Mojave Desert province, California, *Geol. Soc. Am. Spec. Pap.* 325, p. 197-210, 1998.
- Woodburne, M. O., R. H. Tedford, M. S. Stevens, and B. E. Taylor, Early Miocene mammalian faunas, Mojave Desert, California, *J. Paleontology*, 48, 65-76, 1974.
- Young, R. A., and W. J. Brennan, Peach Springs Tuff: Its bearing on the structural evolution of the Colorado Plateau and development of Cenozoic drainage in Mohave County, Arizona, *Geol. Soc. Am. Bull.*, 85, 83-90, 1974.

John C. Merriam and the Barstow Syncline

Thomas Schweich

Jepson Herbarium, University of California, Berkeley. tomas@schweich.com

The name of John C. Merriam is often associated with the early description of mammalian fossils from the Tertiary beds of the Mojave Desert. The fossil assemblage at Barstow and attributed to Merriam turned out to be extremely important. It was the basis for the Wood Committee naming the Barstovian Land Mammal Age—a biochronologic term (spanning the period 16.2–12 Ma) that is now used to tell time across the entire continent. But, how did Merriam find out about the deposits? And, did he ever visit the site? Study of Merriam's letters at the Bancroft Library of the University of California shows that he received the material through a student and that he probably never visited the Barstow beds.

It was a family connection that brought the Barstow fossils to Merriam. John R. Suman, a student of the University of California, brought a small collection of fossil bones and teeth to Merriam in the early spring of 1911 (Merriam, 1919). He received them from his uncle, H. S. Mourning of Los Angeles (Suman, 1913). This collection was presented to the University by Mr. Mourning and Mr. Suman, and constituted the basis for the first study of the Mohave faunas (Merriam, 1911). Mourning maintained a residence in Los Angeles and lived there with his wife. It is not known why Mourning travelled frequently to Barstow or how he was acquainted with John T. Reed.

Mourning learned of the fossils from John T. Reed of Barstow (Mourning, 1911). Reed had some association with mining of Fuller's earth (a type of clay mined from Fuller's Earth Canyon in the western Mud Hills) in the area (Mourning, 1913b). Some time earlier, either in 1895 or in the period 1896–1900, Reed had given some fossils, presumably from the Barstow area, to Dr. Stephen Bowers (Merriam, 1915). However, neither the date nor collection locality can be confirmed. Bowers was a self-taught geologist, paleontologist, and archeologist, newspaper editor, and ordained Methodist minister (Smith, 1983). He is also variously described as a curiosity seeker or "controversialist" and apparently did nothing with the fossils.

In a brief article published soon after receiving the collection from Suman, Merriam called attention to the importance of this discovery (Merriam, 1911). Merriam wasted no time in investigating further. He sent Charles L. Baker, then a fellow in paleontology at the University,

to the locality in the spring and early summer of 1911. Baker was joined later by Suman and others, who assisted with the work. In connection with the paleontological study it was necessary to make a geological reconnaissance of the formations concerned. The results of this investigation were published by Baker (Baker, 1911).

In December, 1911, a second expedition led by Baker visited exposures of Tertiary beds near El Paso Peak on the northwestern border of the Mojave Desert, and made additional collections, adding considerably to the fossils obtained by the Barstow expedition (Merriam, 1919). This site became the type locality for the Ricardo Formation. The Barstow beds were not visited again until 1913.

Suman graduated in 1912 and went to work for the Rio Bravo Oil Company, a subsidiary of the Southern Pacific Railroad, in Houston, Texas (Houston Geological Society, 1998). He did not return to work in California.

In January and February, 1913, Mourning and J. P. Buwalda, another of Merriam's students, visited the beds in the area north of Barstow, and obtained an excellent collection of the fauna at localities previously visited by Mr. Mourning (Mourning, 1913a). This collection was supplemented by specimens purchased from Mourning, and by some very useful material which Mourning gave to the University.

Buwalda travelled to Barstow on the railroad, as did most of Merriam's students. At that time Southern Pacific was providing half-price tickets to students at the University of California. In travelling back and forth between Berkeley and the Mojave Desert, Merriam often had his students stop at a fossil locality near Bena, which was also a railroad station on the Southern Pacific about 15 miles southeast of Bakersfield (Buwalda, 1913b).



John C. Merriam. UCMP, Berkeley.

While in Barstow, Buwalda stayed in the Hotel Melrose opposite the railroad station. The hotel described itself as a modern and first-class concrete building. The proprietors were O. M. and O. A. Lowell (Buwalda, 1913a). There is no building currently standing opposite the Barstow railroad station. Mourning provisioned the camp at the Barstow beds by horse and wagon, and presumably the collections were also transported to the railroad station in Barstow by the same means.

It was also during the February 1913 trip that Buwalda and Mourning visited the Manix beds. Being unauthorized to make the trip, Buwalda was careful to minimize disruption to his authorized work by travelling at night and using the daytime hours for field work (Buwalda, 1913a). There is nothing in the letters that would indicate Merriam was unhappy with Buwalda's enterprising work.

In December, 1913, and January, 1914, March, 1915, and summer of 1915, further expeditions were made to the Pliocene beds west of the town of Mohave (Horned Toad Hills) and to the Ricardo beds west of the El Paso Mountain (Merriam, 1919). None of these expeditions went to Barstow. Merriam made the trip only in March, 1915. Thus, he likely never saw the Barstow syncline.

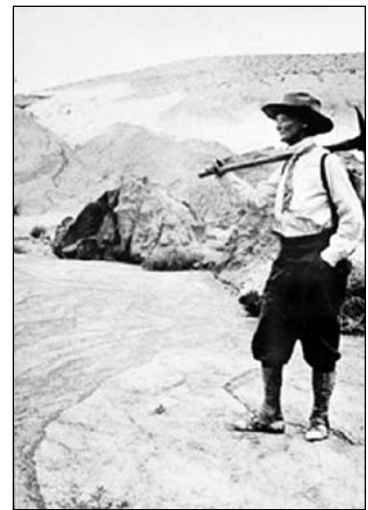
Merriam published his "Tertiary Mammalian Faunas of the Mohave Desert" in 1919, in which he described the Barstow and Ricardo faunas (Merriam, 1919). He did not mention either the Pliocene beds west of Mojave or the Manix beds.

The association of Annie Alexander and John C. Merriam is also well known, as she financed Merriam's 1905 Saurian Expedition and endowed the University of California Museum of Paleontology (UCMP). However, she did not visit the Barstow beds while Merriam was at the University of California. Her displeasure with Merriam when he left the University in 1920 for the Carnegie Institute is well-known.

However, in the spring of 1922 Alexander and Louise Kellogg conducted what would be their most important fossil-hunting expedition during this period, an excursion to the Miocene formations in the Mojave Desert near Barstow. The women worked the entire formation for over a month, collecting a variety of carnivore and ungulate remains—horns, teeth, jaws, skull fragments, and toe bones. The descriptions of her collections generated enough interest that Childs Frick of the American Museum (of Natural History, New York) moved his field party into the region and continued to remove many valuable specimens (Stein, 2001).

Furious at this invasion of her California territory, Alexander wrote directly to Henry Fairfield Osborn, president of the American Museum. Osborn replied promptly saying, "We have all of us felt that important

parts of our collections from the Coast should go to the Museum of Paleontology at the University of California, and I will be very glad to take up with Curator Matthew and Mr. Frick your suggestion." (Stein, 2001) While this may have mollified Alexander, there are very few specimens at UCMP that were collected by Frick.



Annie Alexander. UCMP, Berkeley.

Literature Cited

- Baker, C. L., 1911. Notes on the Later Cenozoic History of the Mohave Desert Region in Southeastern California, Univ. Calif. Publ., Bull. Dept. Geol., vol. 6, pp. 333-383, pls. 34-43, 1911.
- Buwalda, John P., 1913a. Letter to John C. Merriam. February 5, 1913, manuscript. John Campbell Merriam papers, BANC MSS C-B 970, The Bancroft Library, University of California, Berkeley.
- Buwalda, John P., 1913b. Letter to John C. Merriam. February 17, 1913, manuscript. John Campbell Merriam papers, BANC MSS C-B 970, The Bancroft Library, University of California, Berkeley.
- Houston Geological Society, 1998. Our First President: A Tribute to John R. Suman. s. v. <http://www.hgs.org/en/articles/printview.asp?10>, accessed February 7, 2010
- Merriam, J. C., 1911. A Collection of Mammalian Remains from the Tertiary Beds on the Mohave Desert, Univ. Calif. Publ., Bull. Dept. Geol., vol. 6, pp. 163-169. pl. 29.
- Merriam, J. C., 1915. Extinct Faunas of the Mohave Desert, their Significance in a Study of the Origin and Evolution of Life in America. The Popular Science Monthly. Vol. 86, No. 3, March, 1915, pp. 245-264.
- Merriam, J. C., 1919. Tertiary mammalian faunas of the Mohave Desert, Univ. Calif. Publ., Bull. Dept. Geol., vol. 11, 437-595.
- Mourning, H. S., 1911. Letter to John C. Merriam. February 8, 1911, manuscript. John Campbell Merriam papers, BANC MSS C-B 970, The Bancroft Library, University of California, Berkeley.
- Mourning, H. S., 1913a. Letter to John C. Merriam. January 17, 1913, manuscript. John Campbell Merriam papers, BANC MSS C-B 970, The Bancroft Library, University of California, Berkeley.
- Mourning, H. S., 1913b. Letter to John C. Merriam. March 12, 1913, manuscript. John Campbell Merriam papers, BANC MSS C-B 970, The Bancroft Library, University of California, Berkeley.
- Reynolds, R. L., 2010. pers. comm
- Smith, Wallace E., 1983. The Reverend Stephen Bowers: Curiosity Hunter of the Santa Barbara Channel Islands. California History. Vol. 62, No. 1 (Spring, 1983), pp. 26-37
- Stein, Barbara R., 2001. On her own terms: Annie Montague Alexander and the rise of science in the American West. Berkeley, CA: University of California Press, 2001.
- Suman, John R. 1913. Letter to John C. Merriam. February 17, 1913, manuscript. John Campbell Merriam papers, BANC MSS C-B 970, The Bancroft Library, University of California, Berkeley.

The mammalian litho- and biochronology of the Mojave Desert Province

Michael O. Woodburne¹ and Robert E. Reynolds²

¹Department of Geology, Museum of Northern Arizona, Flagstaff, AZ 86001

²Redlands, CA 92373, rreynolds220@verizon.net

Geologic history

The Mojave Desert province is comprised of a diverse group of mountains that range in elevation from about 2,000' (600 m) to 3,000' (900 m), separated by tracts of alluvial cover. The province (Fig. 1) is bounded on the southwest by the San Andreas fault, on the north by the Garlock fault, on the east by the Death Valley–Granite Mountains fault zone, and on the south by the Pinto Mountain fault (e.g., Dokka and Travis, 1990). To paraphrase from Dokka and others (1988), rocks preserved in the Mojave Desert Province collectively record a rich assemblage of Precambrian crystalline basement ranging from 1.87 to 1.2 Ga, upper Precambrian and Paleozoic Cordilleran miogeoclinal sediments and platform rocks, Mesozoic backarc and interarc shallow marine and continental sequences, locally emplaced Permian–Triassic syenitic plutons, extrusion of hypabyssal volcanic rocks of the ?Jurassic “Sidewinder Volcanic Series” of Dibblee (1967), and regionally intruded late Mesozoic granitoid batholithic rocks.

It long has been recognized that the Mojave block must have been a high-standing element for much

of the early Tertiary (e.g. Hewett, 1954), with strata of Late Cretaceous or early Tertiary age preserved only on its margins (examples are the San Francisquito and Cosy Dell formations (Morton and Miller, 2003) of Late Cretaceous and Paleocene age on the southwest in the Devil’s Punchbowl area, and the Paleocene to Eocene (?) Goler

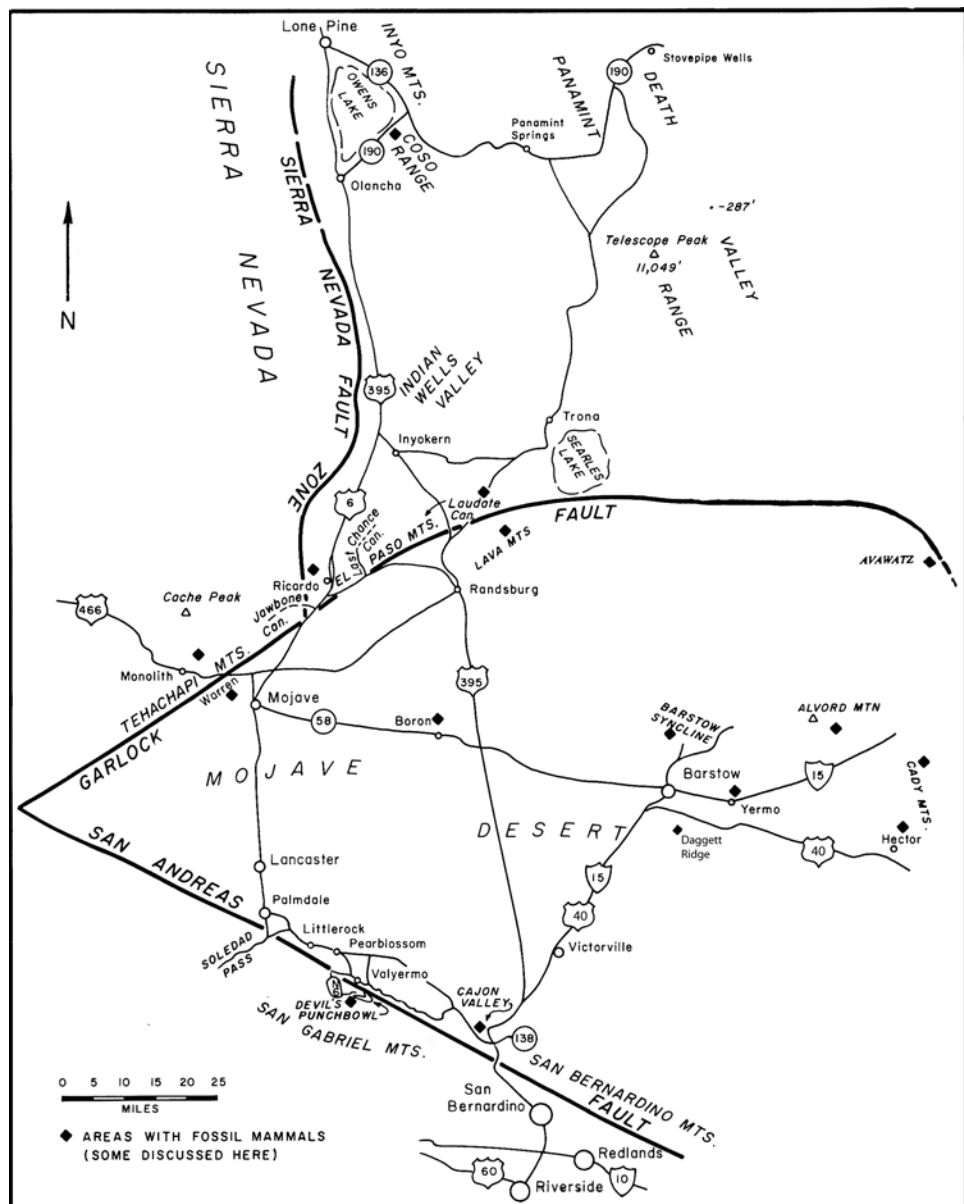


Fig. 1. General map of the greater Mojave Desert (after Woodburne, 1991, Fig. 1)

This report is based on and updated from Woodburne (1991)

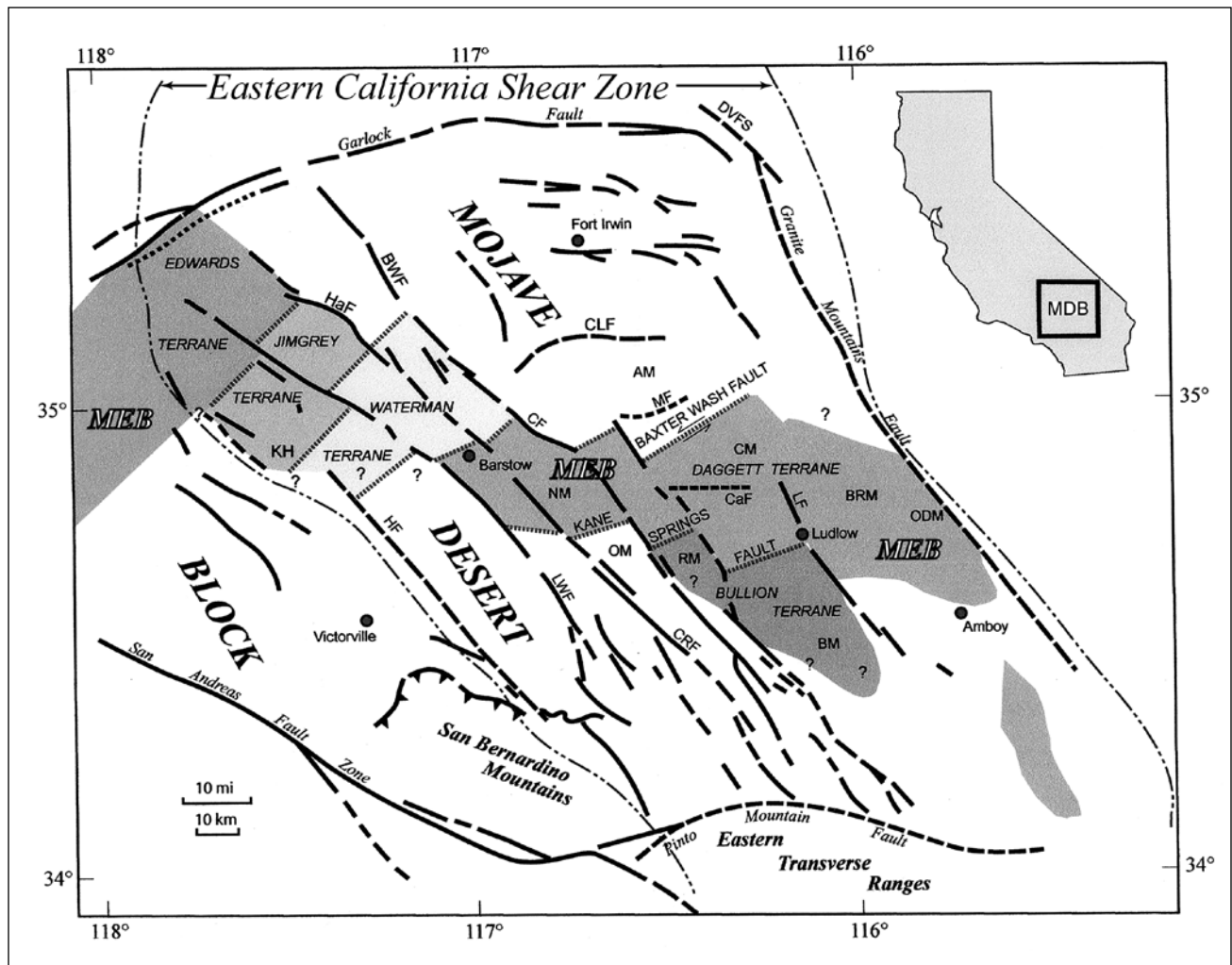


Fig. 2. Mojave Desert block and associated faults and geologic provinces. Boundaries of the Mojave Desert Block include the Garlock Fault to the north, the Death Valley Fault Zone and Granite Mountains Fault to the east, and the San Andreas and Pinto Mountain faults to the south. MEB gray pattern indicates the Mojave Extensional Belt, with the five named terranes (after Ross and others, 1989; Dokka and others, 1998). The barbed line represents the northern frontal fault of the San Bernardino Mountains. Other dashed and solid lines within and north of the Mojave Desert Block represent faults of the Eastern California Shear Zone (Dokka and Travis, 1990a), including BWF, Blackwater fault; CaF, Cady fault; CF, Calico fault; CLF, Coyote Lake fault; CRF, Camp Rock fault; DVFS, Death Valley fault zone; HF, Helendale fault; HaF, Harper fault; LF, Ludlow fault; LWF, Lenwood fault; MF, Manix fault. Geographic features are: AM, Alvord Mountain, BM, Bullion Mountains, BRM, Bristol Mountains, CM, Cady Mountains, KH, Kramer Hills, NM, Newberry Mountains; ODM, Old Dad Mountains; OM, Ord Mountain; WH, Waterman Hills. Inset map shows location in southern California. The assistance provided by Brett F. Cox, USGS Menlo Park, in the production of this figure is greatly appreciated.

Formation on the northwest in the El Paso Mountains (Laudate Canyon, Fig. 1).

As summarized in Bartley and Glazner (1991), Dokka et al. (1989, 1991, 2001), Dokka and Travis (1990a, b), Fillmore et al. (1994) Fillmore and Walker (1996), and Glazner et al., 1989), the Mojave Desert Block has been and continues to be an area of significant accommodation of the interaction between the Pacific and North American tectonic plates (also Dickinson, 1996). In outline form, this response is categorized in three major tectonic provinces during the late Cenozoic: the Mojave Extensional Belt (MEB, including the Central Mojave

Metamorphic Core Complex; CMMCC); the Trans Mojave–Sierran Shear Zone (TM-SMZ); and the Eastern California Shear Zone (ECSZ). Diagrammatic relations of these three tectonic provinces are shown on Fig. 2.

The initial response to plate interaction resulted in a period of extensional tectonics reflected by regional uplift and detachment, bimodal volcanism (basaltic and intermediate composition) and clastic and volcanoclastic sediment accumulation from about 24–19 m.y. (Fillmore and Walker, 1996). The affected area, known as the Mojave Extensional Belt underwent about 30 km of N–S regional extension (Dokka, 1989; about 50 km of extension; Fill-

more and Walker, 1996) and likely was associated with, or just preceded, 30–60° of clockwise rotation (Ross et al., 1989; Ross, 1995; Woodburne, 1996).

A second tectonic episode in the Mojave Desert area is described as a time of E–W dextral shearing in the Mojave Extensional Belt at 21–18 m.y. The region affected is largely congruent areally with the Mojave Extensional Belt, and is dubbed the Trans Mojave–Sierran Shear Zone (Dokka and Travis, 1990a; Dokka et al., 1998). Both the MEB and TM-SSZ are considered to have been a reflection of regional collapse during this interval, with activity of the TM-SSZ manifested over band ca 90 km. wide, and with 80 km or more of net dextral shear (Dokka and Ross, 1995). In the interpretations of Fillmore and Walker (1996) and Ingersoll et al. (1996) the period of extension persisted until about 19 Ma, thus overlapping the age of the TM-MSSZ.

Regardless of these details, the period of extension and associated crustal rotation ceased prior to the time of deposition of the Peach Spring Tuff (18.5 Ma; Neilson et al., 1990) which shows no rotation relative to its source area on the Colorado Plateau (Wells and Hillhouse, 1989).

This was followed by an interval from about 13–10 to 0 m.y. (perhaps 2.5 to 0 m.y.) in which the Eastern California Shear Zone was developed and truncated the prior fabrics (also Dickinson, 1996). The Eastern California Shear Zone likely extends northwestward from the Gulf of California to include the Walker Lane and projected correlates in eastern Oregon (Dokka, 1993; Dokka et al., 1989; 2001; Dickinson, 1996). In the Mojave Desert region, the ECSZ accounts for about 65 km of net dextral shear across its breadth (Dokka, 1993), including significant (up to ca 90° clockwise) rotation of crustal blocks

bounded to the north and south by E–W trending left slip faults, and on the east and west by north-west trending right slip faults (e.g., Kamerling and Luyendyk, 1979; Luyendyk et al., 1980; Carter and others, 1987; Dickinson, 1996). The kinematic link between the San Andreas fault zone and the ECSZ indicates that inboard strike-slip faulting within the continent (Mojave Desert region and beyond) can be related to the transform tectonics of the Pacific and North American plates. The initial age of the ECSZ (ca 6 Ma) is inferred from the age of strike-slip activity on faults in the Death Valley region (Stewart, 1983) that are part of the eastern boundary of the ECSZ (Dokka and Travis, 1990a).

The western border of the Eastern California Shear Zone (ECSZ) in the Mojave Desert Block is located about 40 km west of Barstow, California, and elements of the Barstow Formation exposed about 13 km to the north reflect tectonic and depositional activities prior to the earliest manifestation of the ECSZ.

The Barstow Formation ranges in age from ca 19–13 m.y. (Woodburne et al., 1990) and has been interpreted to interfinger with the Pickhandle Formation on the northern side of the Mud Hills (Woodburne et al., 1990). Ingersoll et al. (1996) assigns parts of the Pickhandle Formation here to the Mud Hills Formation, as a unit that interfingers with, or overlies, the Pickhandle, but unconformably underlies the Barstow Formation. Whereas this may remove elements of the Barstow Formation from that unit on the northern Mud Hills, the ca 19 Ma age for the lowermost part of the Barstow Formation remains intact (contra Ingersoll et al., 1996) as determined from rocks on the southern limb of the Barstow Syncline, and the southern flank of the Mud Hills. In this context the lower part of the Barstow Formation overlaps slightly the age of

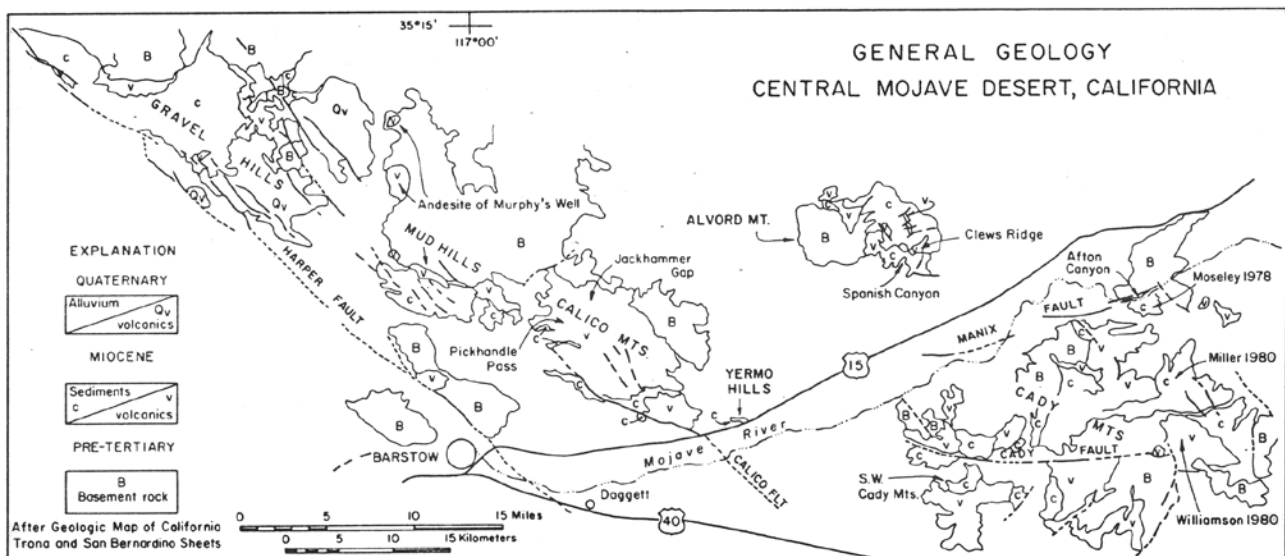


Fig. 3. General geology of the central Mojave Desert (after Woodburne, 1991, Fig. 3).

the Peach Spring Tuff (ca 18.5 Ma; Nielson et al., 1990) considered to post-date the interval of extension and rotation preserved in the Mojave Extensional Belt.

The extensional regime in the MEB lasted from about 22–17 m.y., and is generally divided into two phases. The first, from about 22 to 20 m.y., records intense extension via the activity of crustal-scale, simple shear, low-angle normal faults, high-angle normal faults and extension fracturing along with intrusion of intermediate to silicic volcanic rocks, including episodes of explosive volcanic activity (e.g., Pickhandle Formation in the Mud Hills and Calico Mountains, and the Spanish Canyon Formation of the Alvord Mts; Figs. 1, 4). The Peach Spring Tuff, a major regional marker unit that extends across the Mojave from Arizona to about Barstow (Wells and Hillhouse, 1989; Buesch, 1994) forms an effective stratigraphic lid on the extension interval, and is dated isotopically at about 18.5±0.2 Ma (Nielson, and others, 1990). This extension episode was followed by an interval of high-angle normal faulting and dike emplacement from about 19 to 17 Ma.

The extension interval was accompanied by at least local uplift, with the Waterman Gneiss, in the Waterman Hills south of the Mud Hills which contain the type Barstow Formation, having risen a documented 92 m during this time, possibly significantly more (e.g., Dokka and others, 1988). Beginning about 16 Ma clasts of Waterman Gneiss are contained in type Barstow Formation sediments on the south side of the Mud Hills (e.g. Woodburne and others, 1990).

Most volcanoclastic and epiclastic sedimentary units of the central Mojave block post-date the interval of extension and normal faulting, although the lower parts of the Hector and type Barstow formations apparently were coeval with some or all of this interval. Most of these interiorly accumulated sequences are post-extensional and occur as a complex of marginal alluvial and more interior lacustrine facies that formed as recently as about 16 Ma. These deposits, generally termed Barstow Formation or equivalent, contain numerous beds of air-fall and water-laid tuff that form convenient marker beds as well as being amenable to isotopic age analysis. The rocks are locally fossiliferous, as well, and contain the main record of fossil mammal evolution in the Mojave Desert Province. Collectively, they record interior basin deposition that covered an area of about 4,000 square miles.

Especially in the central area, between the Mud Hills and Gravel Hills on the west and the Cady Mountains on the east, the strata were subsequently deformed, folded, and cut by faults that either trend northwest and show right-lateral separation or trend about east west and show left-lateral separation.

A model of crustal rotation for the central Mojave region originally proposed by Kamerling and Luyendyk (1979) and Luyendyk and others (1980) suggested that crustal blocks in this area that are bounded by east–west trending faults experienced ca 90° of clockwise rotation during the Miocene. Blocks bounded by northwest-trending faults were considered to have undergone no significant rotation (see examples of such blocks in Fig. 2). Further work indicated a much more complex history of crustal activity, however, with the amount, sense, and timing of rotation depending largely on whether the terrane under consideration was west of both the San Gabriel and San Andreas faults, between those faults or east of the San Andreas fault (e.g., Hornafius and others, 1986; Golombeck and Brown, 1988; Carter and others, 1987; MacFadden and others, 1990a, b).

Weldon (1985) has suggested that rocks of the Crowder Formation and younger strata as well in Cajon Valley have undergone no significant rotation since their deposition beginning about 17 Ma. This implies that the adjacent parts of Cajon Valley and the San Bernardino Mountains have undergone no rotation, as well.

MacFadden and others (1990a) analyzed the paleomagnetic properties of the type Barstow Formation in the Mud Hills (Fig. 1), and suggested that this unit (and hence the district in which it lies) has undergone negligible rotation since its deposition, ca 19 Ma. In conformity with the Kamerling and Luyendyk (1979) model, this area is bounded by northwest striking faults.

MacFadden and others (1990b) also analyzed the paleomagnetic signatures of the Hector Formation in the northern Cady Mountains and suggested that these rocks underwent about 260° of clockwise rotation subsequent to their deposition, or after about 16 Ma. Also consistent with the general model, the Cady fault that bounds the Hector rocks on the displays left-lateral separation.

In other studies, Ross and others (1989) suggest that a number of terranes in the central Mojave Desert associated with northwest trending faults underwent about 30–50° of clockwise rotation post-early Miocene, in some places subsequent to about 50° ± 15.6° of clockwise rotation during the time of early Miocene extension.

Among the most interesting aspects of the scenario of deformation across the Mojave Desert is whether or not it deformed homogeneously by simple shear (e.g., Garfunkel, 1974), whether the rotation was mostly counterclockwise (e.g., Garfunkel, 1974), or otherwise, and whether or not the rotation(s) took place mostly in the late Miocene (e.g., Kamerling and Luyendyk, 1979).

Both right-lateral and left-lateral families of faults on the Mojave Desert have been active into Recent or nearly Recent times (such faults include the Cady fault and the

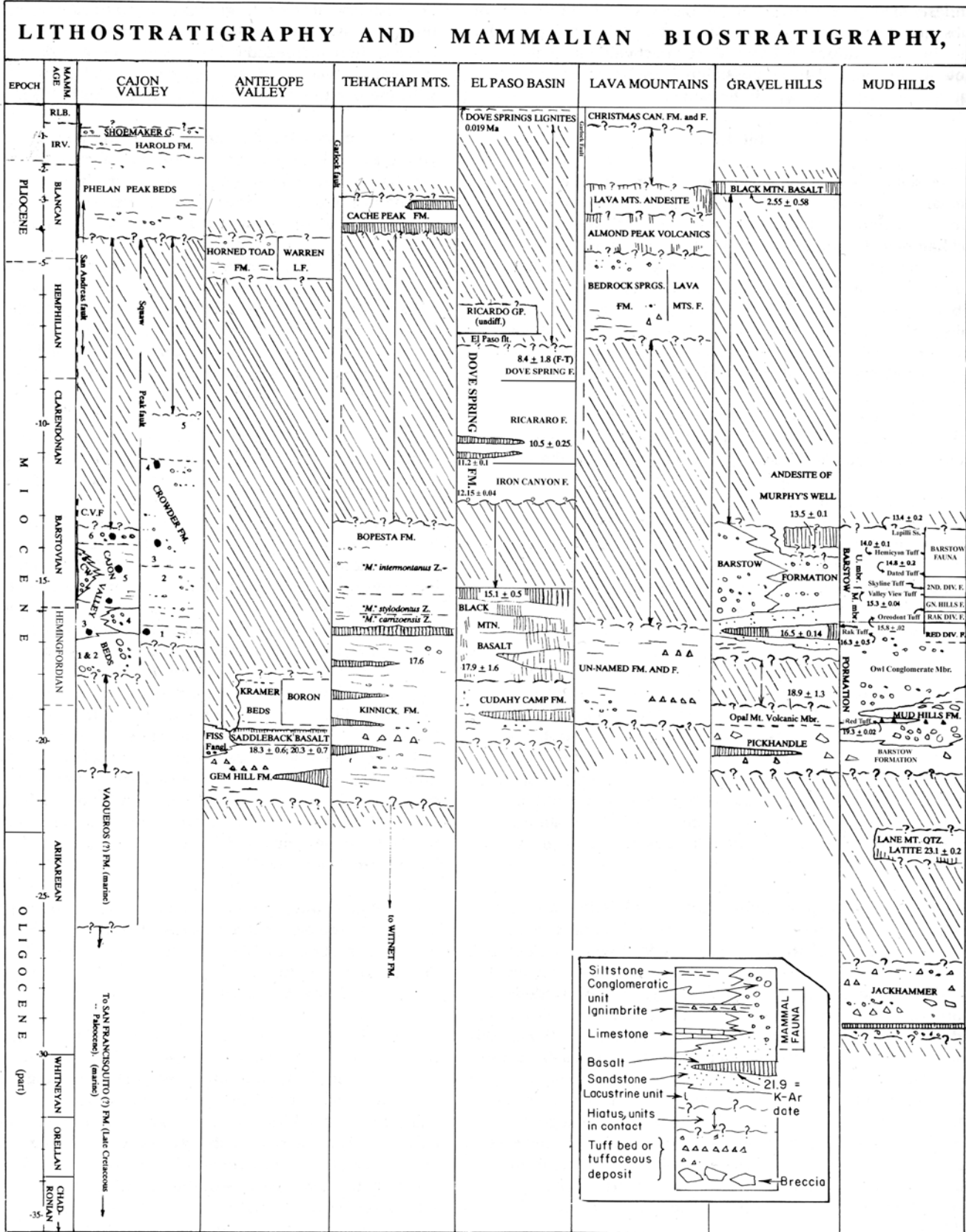
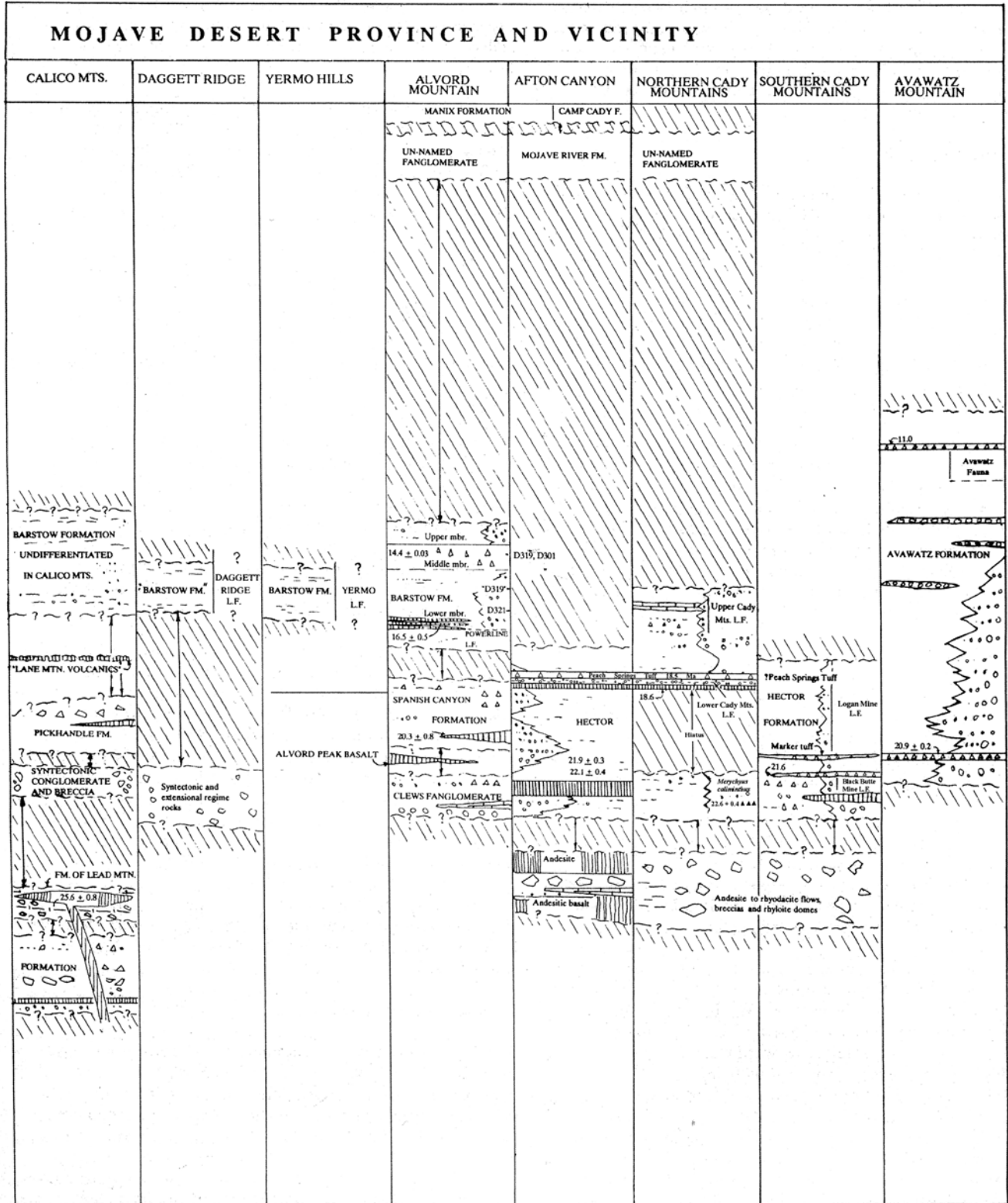


Fig. 4. Correlation chart of rock units (after Woodburne, 1991, Fig. 4).



faults that cut the Mud Hills area; see fig. 3). Meisling and Weldon (1989) indicate that tectonic activity in the San Bernardino Mountains associated with northwest-trending faults along its northern front began between ca 1.5 Ma and 0.7 Ma, and activity continues to the present.

Regional studies also link the onset of the most recent phase of activity along the San Andreas fault with the time of opening of the Gulf of California (e.g., Atwater, 1970). In this context it is relevant that Dokka and Travis (1990) suggest that as much as 65 km of cumulative right shear is distributed across several domains that embrace the Mojave Desert block; that each domain has accommodated the strain independently of the others (some hardly rotated or faulted at all; e.g., the western Antelope Valley; Fig. 1); that virtually none are physically linked to the Garlock fault, and that all domains experiencing activity did so sometime in the past 10 Ma, but more likely that the deformation was considerably younger than that.

Thus, the evidence seems to support the thesis that the Mojave Desert block is among the most recently active

terranes in southern California and that fault-slip and crustal rotation still are taking place.

Central Mojave Desert

Figure 3 shows the districts in the central Mojave Desert having sequences useful to fossil mammal biostratigraphic and faunal analysis. Some of these areas also contain unfossiliferous rock sequences that pertain to extensional episodes that affected this area, as well. The discussion here begins on the west and progresses eastward, after the presentation in Woodburne and others (1982), Woodburne and Tedford (1982), Woodburne and others (1990), MacFadden and others (1990) and Woodburne (1991).

As indicated in the introduction, a diverse basement rock terrane in the central Mojave Desert is overlain unconformably by a thick sequence of Tertiary-age volcanoclastic and epiclastic sediments interbedded with air-fall and some ash-flow tuffs. The Tertiary rocks range in age from more than 26 Ma to about 12 Ma and record the

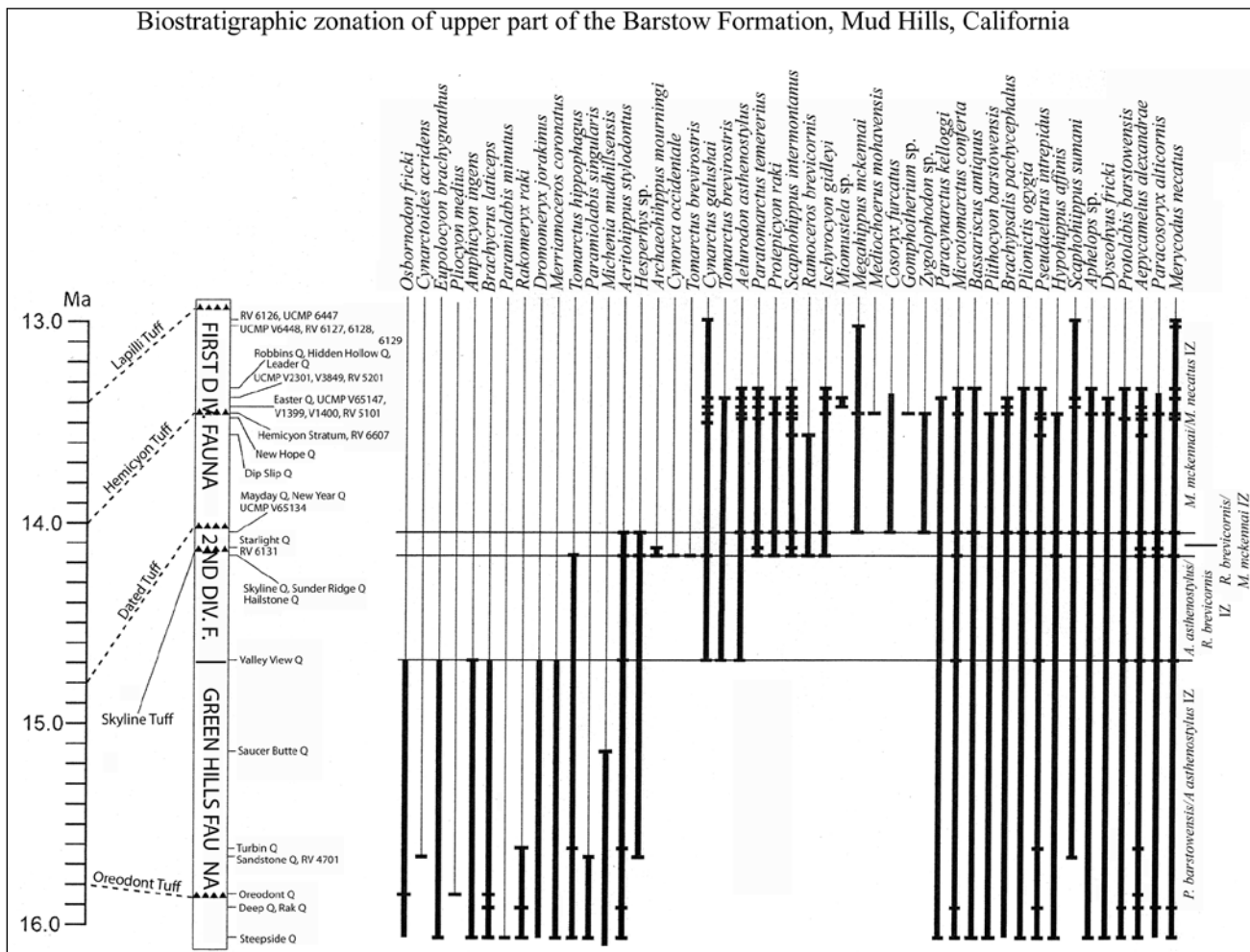


Fig. 5. Biostratigraphic zonation of the upper part of the Barstow Formation, Mud Hills, California (after Pagnac, 2009, Fig. 3).

Magnetostratigraphy and calibration of the Barstow Formation, southern California

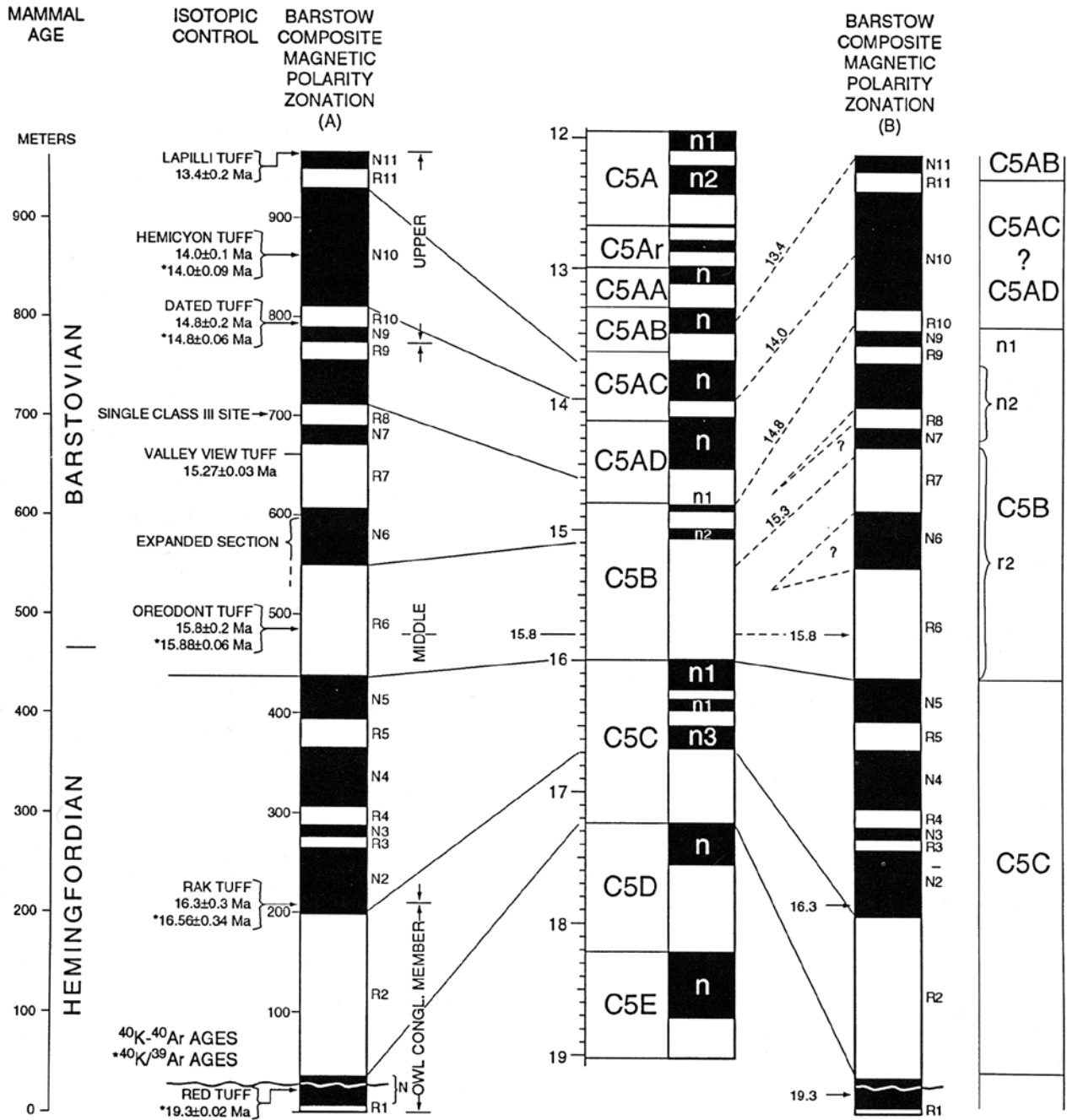


Fig. 6. Composite magnetic polarity stratigraphy, Barstow Formation, Mud Hills, California and its correlation to the MPTS (after Woodburne, 1996, Fig. 5)

onset of internal drainage in the Mojave Desert Province (e.g. Hewett, 1954; Byers, 1960). Lacustrine sediments in the Lead Mtn. area south of the Calico Mts. are somewhat older than 26 Ma (see Fig. 4), and the Jackhammer Formation may be about that old, as well (cut by dikes that old). The apparent absence of Upper Cretaceous to lower Paleogene terrigenous strata on the Mojave Desert Province is striking, with the only hints as to local activities during that time provided by units exposed on the

northwest (Goler Formation) and southwest (Cosy Dell and San Francisquito formations) margins of the block (e.g., Figs. 1, 4).

Most of the mammal-bearing rock units in the central part of the Mojave are referred to the Barstow and Hector formations. The Barstow depocenter apparently was confined mostly to the western part of the district, from the Gravel Hills to the Alvord Mts. (Fig. 3), and ranges in age from about 19–13 Ma (Figs. 4–6). The Hector Formation

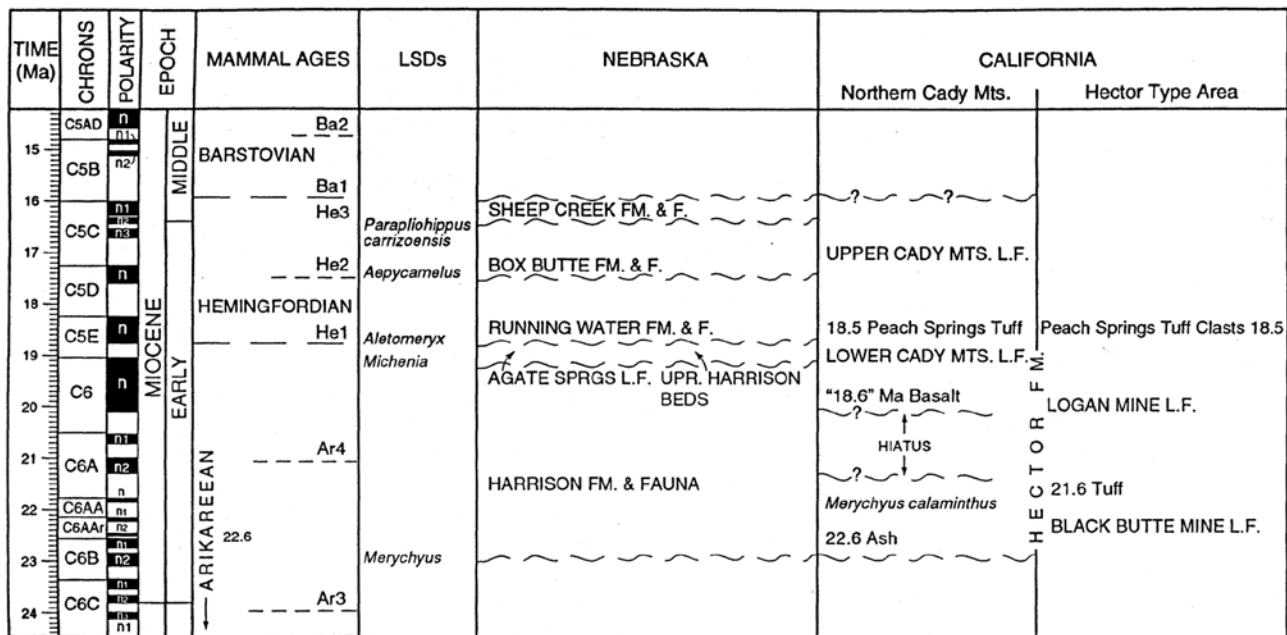


Fig. 7. Regional correlation of the late Arikareean and Hemingfordian mammalian faunas of the Cady Mountains district relative to sequences in Nebraska. After Woodburne (1998; Fig. 5).

flanks the Cady Mountains in the eastern part of the area and ranges in age from about 22–16 Ma (Fig. 7).

The Avawatz Formation (ca 20–11 Ma old) is exposed in the Avawatz Mts. on the eastern margin of the area. It is one of the youngest Miocene rock units exposed in this district, and may more properly be considered as relevant to events on the southern Basin and Range Province near the southern end of the Death Valley fault zone (e.g., Spencer, 1990). The Avawatz Local Fauna (Henshaw, 1939) is of Clarendonian age and is derived from the upper part of the formation (Fig. 4) in association with a tuff dated at ca 11 Ma (e.g., Evernden and others, 1964).

In the Lava Mountains, on the northern edge of the province, and south of the Garlock fault, the Bedrock Springs Formation (Fig. 4) contains fossils of middle to late Hemphillian age (Smith, 1964; Whistler, 1991), and is a little younger than the Dove Spring Formation, but may be about as old as sediments assigned to the Ricardo Group (Whistler and Burbank, 1992; Whistler, 1991), exposed in the El Paso Basin (Fig. 4) on the north side of the Garlock. The Dove Spring Formation of the Ricardo Group ranges in age from early Clarendonian to early Hemphillian, or from about 12.5 to 7 Ma. Whistler et al. (2009) provide a revised geochronology and biostratigraphy of the Dove Spring Formation and its faunas, and provide many more age calibrations than shown on Fig. 4. References for the geologic history of the Garlock fault and adjacent areas include Dibblee (1952), Garfunkel (1974), Cummings (1976), Cox (1987), Loomis and Burbank (1988), Pampeyan and others (1988), and Dokka and Travis (1990).

The Boron Local Fauna (Whistler, 1984) comes from the upper part (Arkose member of the Kramer beds) of the Tropico Group, in the Antelope Valley, on the western part of the province (Fig. 4). The fauna, of early Hemingfordian age, is contained with borate-bearing sediments mined by U.S. Borax. The sediments rest on the Saddleback Basalt, dated (Armstrong and Higgins, 1973) at 20.3 ± 0.7 Ma and 18.3 ± 0.6 Ma (Kistler, in Whistler, 1984).

Farther to the northwest, near the town of Mojave, and yet on the south side of the Garlock fault, the Horned Toad Formation (Dibblee, 1967) has yielded the Warren Local Fauna (May, 1981a, 1981b) of late Hemphillian age (Fig. 4).

Cache Peak is part of the Tehachapi Mountains just north of the Garlock fault adjacent to the western part of the Mojave Province. The range contains three major Tertiary units. The oldest, the Witnet Formation, is unfossiliferous but is lithologically similar to, and commonly correlated with, the Goler Formation of the El Paso Mountains (Dibblee, 1967; Cox, 1982, 1987; Cox and Diggles, 1986; McDougall, 1987; McKenna and others, 1987). The Kinnick Formation unconformably overlies the Witnet, and contains a hornblende biotite tuff breccia dated by Evernden and others (1964) at 17.6 Ma. Quinn (1987) revised the Kinnick and the conformably overlying Bopesta Formation. As revised, the Bopesta Formation consists of a sequence of volcanoclastic, epiclastic, and minor lacustrine strata that ranges in age from late Hemingfordian to late Barstovian.

Many of the units described above are unconformably overlain by unfossiliferous units, and some of these likely

are of Pliocene age. The Black Mountain Basalt that unconformably overlies the Barstow Formation in the Gravel Hills (Fig. 4) has been dated at about 2.55 ± 0.58 Ma (Burke and others, 1982). Nevertheless, throughout the greater Mojave Desert Province there seems to have been a change in regional deposition such that erosion predominated, in contrast to deposition at the scale recorded for the Miocene. This observation leads to the concept of a 5 million year gap (from about 7–2 Ma; Figure 4) in the recorded geologic history of the Mojave Desert Province and contiguous areas. Some record of events that took place during this interval is preserved in locally derived deposits in or adjacent to the San Bernardino and San Gabriel mountain ranges. These include the succession in Cajon Valley (Fig. 4; also see Woodburne, Cajon Valley, 1991), and outcrops of the Old Woman sandstone (e.g., Sadler, 1982) that can be traced along the front of the San Bernardino Mountains. Sequences that are geologically thick, relatively continuous, and temporally extensive that span this gap are rare, but other data may develop for the Mojave region, and help fill in the geological history of this interval (e.g., Miller and Yount, 2002). Probably the best example is the sequence preserved in the San Timoteo Badlands (Frick, 1926; May and Repenning, 1982; Reynolds and Reeder, 1986, 1991; Albright, 1999) but this is south of, rather than on, the Mojave Block. Resolution of what was taking place during the “missing 5 million-year gap” remains one of the important desiderata in contributing to our understanding of the geological and biological evolution of this region.

The post-Miocene geologic milieu also saw the early to late Miocene units being tilted, folded, and cut by faults that generally trend northwest and show right-lateral separation and, more rarely, trend east–west and show left-lateral separation. During the later part of this interval (mostly Pleistocene) internal drainage once again produced major areas of lacustrine deposition, now recorded by the Manly and other lake beds in the southern Death Valley, Tecopa, Soda and Manix basins. These deposits range in age from about 2 Ma to 19,000 yrs and locally record mammalian taxa of Blancan, Irvingtonian to Rancholabrean age (e.g., Jefferson, 1985, 1991; Woodburne and Whistler, 1991; Reynolds and Reynolds, 1985).

Models proposed to relate the geological features summarized above to the general tectonic regime of North America include (1) plate margin subduction reflected by the onset of Tertiary volcanism (Atwater, 1970; Lipman and others, 1972; Christiansen and Lipman, 1972; Armstrong and Higgins, 1973), (2) interplate strain reflected in counterclockwise rotation of fault-bounded blocks (Garfunkel, 1974; Cummings, 1976), (3) clockwise rotation of selected fault-bounded blocks (Kamerling

and Luyendyk, 1979, 1985; Luyendyk and others, 1980, 1985; Terres and Luyendyk, 1985; Carter and others, 1987; Hornafius, 1985; Hornafius and others, 1986; Luyendyk and Hornafius, 1987), and (4) a combination of early Miocene extension and post-extensional deposition followed by post-late Miocene faulting and rotation partitioned across six domains from the San Andreas fault to the Death Valley fault zone (Dokka, 1983, 1986, 1989; Dokka and Glazner, 1982; Dokka and Travis, 1990). Bartley and others (1990) also suggest that the region has been subjected to north–south contraction (see also Smith, 1991; Glazner and others, 1989, 2002).

The following summary sets out the basic stratigraphic and geochronologic relationships of Cenozoic rock units in selected districts of the Mojave Desert Province. Special emphasis is given to those sequences that have produced fossil vertebrates. Figures 1 and 3 show the general location of these areas.

Gravel Hills

The Gravel Hills (Fig. 3), about 35 km northeast of Barstow, record what apparently was the westernmost extent of the Barstow and Pickhandle formations. According to Dibblee (1968), the Cenozoic sequence begins with the **Pickhandle Formation** (V on Fig. 3). This largely pyroclastic unit is about 600 m thick, and unconformably overlies pre-Tertiary quartz monzonite. The Pickhandle is mostly composed of light gray breccia and tuff, interbedded with minor basalt flows, conglomeratic lenses, rhyolitic flow breccia, and beds of white lithic and lapilli tuff. Burke and others (1982) report a whole-rock date of 18.9 ± 1.3 Ma from the Opal Mountain volcanic member at the top of the Pickhandle (formerly considered as a separate unit by Dibblee, 1968).

Dibblee (1968) considered the Pickhandle to have been derived from volcanic centers in the Gravel Hills and in the Opal Mountain area, as well as from loci in the Calico Mts. (vicinity of Pickhandle Pass, Fig. 3), and Byers (1960) proposed that the Spanish Canyon Formation of Alvord Mountain, some 60 km to the east, was genetically related to the Pickhandle, as well (see Figs. 3 and 4). In the Mud Hills the Pickhandle is exposed mostly on the northern side of the Barstow Syncline, but underlies and interfingers with the Barstow Formation there. Walker and others (1990) have correlated volcanoclastic rocks atop the Waterman Hills and the Mitchell Range with the Pickhandle, but the great diversity of exotic clasts (apparently derived from the Alvord Mts.) is unusual for the Pickhandle rocks of the Mud Hills (see Mud Hills Formation of Travis and Dokka in Dokka and others, 1991).

The **Barstow Formation** (C, Fig. 3) is more widely exposed in the central and western parts of the Gravel

Hills than in the east. It consists of a lacustrine sequence overlain (Fig. 4) by basal beds of coarse-grained, fluvialite sandstone and thick units of fanglomerate that contain granitic and volcanic detritus (Dibblee, 1968; pl. 1: 30–31). The total sequence ranges up to 1,500 m in thickness. The basalt flow in Black Canyon near the central Gravel Hills is undated and may not be the same flow as the one from which Burke and others (1982) obtained a K-Ar date of 16.5 ± 0.4 Ma.

Fossil mammals are sparse in the Barstow deposits below the basalt in Black Canyon, but seem to pertain to the Green Hills Fauna of the Mud Hills (Woodburne and others, 1990, Fig. 6), calibrated there at between 15.4–15.8 Ma. Those above the basalt in Black Canyon appear to correlate with elements of the Barstow Fauna of the Mud Hills (ca 14.8–13.4 Ma; First Division Fauna, Fig. 6). The mammal remains are fragmentary, and geochronologic correlations based on them remain tentative (see also Lewis, 1968b, p. 35).

The two fanglomeratic facies—granitic on the west and southwest, volcanic debris on the north (Fig. 4)—reflect deposition from two different sources. Dibblee (1968: 32) suggests that the fanglomerate of granitic and quartz latite clasts were derived from rising basement terranes to the west and southwest; the fanglomerate of volcanic debris apparently accumulated at the same time, but from northwestern sources. These data suggest that the western edge of the Barstow depositional basin lay in the vicinity of the Gravel Hills. Although coarse-grained sediments occur in the western Mud Hills (and generally interfinger eastward with finer-grained units), there is no physical continuity between the Gravel and Mud hills.

The youngest unit of Miocene age in the Gravel Hills area is the “**Lane Mountain Andesite**” mapped by Dibblee (1968, p. 1) in the vicinity of Murphy’s Well (Fig. 3) about 10 km north of the western end of the Mud Hills. This unit is shown on Fig. 4 as the **Andesite of Murphy’s Well**, dated by Burke and others (1982) at 13.5 ± 0.1 Ma and, except for resting unconformably on pre-Tertiary basement, is stratigraphically unconfined. Nevertheless, Woodburne and others (1990) note the presence of a distinctive facies of hornblende-bearing sandstone in the uppermost part of the Barstow Formation in the northwestern Mud Hills, and point out that the age of this part of the formation (ca 13.4 Ma) is very close to the age of the Murphy’s Well andesite.

The **Black Mountain Basalt**, dated at 3.77 ± 0.11 Ma by Oskin and Iriondo (2004) caps the Barstow Formation in the Gravel Hills and is an important marker that demonstrates about 1.8 km of right-lateral offset on the Blackwater fault (Fig. 2) in this area.

Burke and others (1982) and Valentine and others

(1988) investigated the possibility of tectonic rotation (about vertical axes) of late Cenozoic units in this area. They suggested that the Opal Mountain member of the Pickhandle Formation and the Andesite of Murphy’s Well may have experienced up to 30° of subsequent counterclockwise rotation, but that the Barstow Formation and the Black Mountain Basalt were effectively un-rotated. Note, however, that these possibilities are based on very few data points and must remain tentative.

Mud Hills

A relatively complete sequence of deposits of early and medial Miocene age crops out in the Mud Hills, north of Barstow, California (Figs. 1, 3). The sequence consists of the Jackhammer, Pickhandle, and Barstow formations, in ascending order (Fig. 4), and the Lane Mountain Quartz Latite. For geological literature, see also Hershey (1902), Baker (1911), Dibblee (1968), Durrell (1953), and Steinen (1966). Paleontological references include Merriam (1913, 1915, 1919), Hall (1930), Stirton (1930), Wood (1936), Stock (1937), Schultz and Falkenbach (1947, 1949) and Tedford and Alf (1962). Pagnac, (2005, 2006, 2009).

The type locality of the **Jackhammer Formation** is in Jackhammer Gap (Fig. 3), in the northwestern Calico Mountains (McCulloh, 1952, in Dibblee, 1968), but the unit is preserved in its westernmost occurrence on the north limb of the Barstow Syncline in the Mud Hills (Dibblee, 1968, pls. 1, 3). There the Jackhammer unconformably overlies pre-Tertiary quartz monzonite, is about 30 m thick, and is unconformably overlain by the Pickhandle Formation. In the Calico Mts., the Jackhammer is about 240 m thick and includes beds of conglomerate, basalt, tuff, tuff breccia and tuffaceous sandstone.

The Jackhammer apparently was one of the earliest units to be deposited in a basin that extended from the Calico Mts. toward the Gravel Hills. In the Calico Mts. area, Lambert (1987) reports that the Jackhammer is cut by dikes dated at 25.6 ± 0.8 Ma. The lower limit to the age of the formation still is unknown but, if Cenozoic deposition on the Mojave Desert is related regionally to the time of beginning consumption of the subduction zone along the western margin of California (e.g., Atwater, 1970), the Pickhandle may not be older than 30 Ma.

The **Lane Mountain Quartz Latite** is of limited areal extent as restricted by Burke and others (1982) and dated at 23.1 ± 0.2 Ma. In contrast to Dibblee (1968) who considered the unit to post-date the Barstow Formation, these authors suggest that this quartz latite is a remnant of a number of welded quartz latite ash-flow sheets that locally contain unwelded tuff and vitrophyre at the base.

The present areal restriction of this unit suggests considerable erosion subsequent to its extrusion; it also may mark the age of change from a regime that produced mainly andesitic to rhyodacitic extrusive rocks to one that resulted in largely basaltic eruptions and concurrent fluvio-lacustrine deposition and the development (sometimes explosive) of air-fall to ash flow tuffs. Based on data in the Calico Mts. (below), the Lane Mountain Quartz Latite is younger than the Jackhammer Formation and older than the Pickhandle.

The **Pickhandle Formation** crops out mostly on the north limb of the Barstow Syncline and is there about 1,300 m thick. It is composed of a basal interval of gray conglomerate, and greenish-gray to reddish-gray arkosic to tuffaceous sandstone; this is followed by a middle unit of lithic tuff and andesitic tuff breccia, then succeeded by an upper unit of granitic and rhyolitic breccia and megaconglomerate that is interbedded with the Barstow Formation (Woodburne and others, 1990).

The age of the Pickhandle in the Mud Hills is as least as young as the basal part of the Mud Hills Formation, with which it interfingers (below). If the proposed correlation of the Pickhandle and the Spanish Canyon Formation at Alvord Mtn. is valid, then the age shown on Fig. 4 may be reasonable.

The tuffaceous portions of the Pickhandle apparently were the result of episodes of explosive volcanic activity, and the spectacular, thick lenses of monolithologic granitic and rhyolitic breccia exposed in the upper part of the Pickhandle Formation in Owl Canyon may have been deposited as landslides or debris flows. The size of the blocks suggests nearby rather than more distant provenance, and the fact that these overlie tuffaceous units of volcanic origin suggests that the breccias were the result of local tectonic activity that produced a rugged terrane of granitic mountains on the margin of the Pickhandle and, later, the Barstow basin (Dibblee, 1968, p. 22, 23).

Travis and Dokka (in Dokka and others, 1991) remove the breccias and megaconglomerates of Dibblee's (1968) Pickhandle Formation, and term these the **Mud Hills Formation**. The Mud Hills Formation crops out on the north limb of the Barstow Syncline, is about 300 m thick, and overlies and interfingers eastward with the volcanoclastic Pickhandle Formation. The Mud Hills is considered to range in age from about 19.8–18.1 Ma, and is interpreted as a sequence of rock avalanche and debris flow deposits that is unconformably overlain by the Barstow Formation on the north limb of the syncline. We note that the 19.3 ± 0.02 Ma Red Tuff found in the lower part of the Barstow Formation on the south limb of the syncline also occurs within the Mud Hills Formation on the north. Fig. 4 interprets that the two formations interfinger re-

gionally, especially in that the lower part of the Mud Hills Formation. is considered to have had a southern source; the upper part was derived mainly from the north (Travis and Dokka in Dokka and others, 1991).

The lithostratigraphy, paleomagnetism, biostratigraphy and geochronology of the **Barstow Formation** in the Mud Hills has been reviewed by MacFadden and others (1990) and Woodburne and others (1990), and the following is paraphrased from those sources.

The Barstow Formation is composed of a sequence of fluvial and lacustrine sediments, and water-laid air-fall tuff beds; it is about 1,000 m thick, interfingers with the Pickhandle Formation and is unconformably overlain by Quaternary alluvium in the Mud Hills, and by the Pliocene Black Mountain Basalt in the Gravel Hills (Fig. 4). The sediments have been folded into a syncline (Barstow Syncline) that trends about east–west and broken by several faults that generally trend northwest–southeast and show right-lateral separation (e.g., Fig. 3).

Overall, the Barstow Formation is interpreted as being a generally upward-fining sequence of basin fill deposits, with coarser-grained marginal facies being preserved on the north, west and southwest, and more distal (basinal) facies preserved to the northeast, east and southeast. Based on lithologic data the Owl Conglomerate Member deposits now exposed on the north and south limbs of the Barstow Syncline, accumulated on the north and south sides of the Barstow depositional basin, respectively.

In the following discussion the Barstow faunal succession is divided into four major intervals, based largely on terminology employed by members of the Frick organization (F:AM; Frick American Mammals, associated with the American Museum of Natural History, New York) which collected fossil mammals from the Barstow Formation mostly during the 1930s to 1950s. These are, from lowest to youngest, the Red Division, the Rak Division, the Green Hills Fauna, the Second Division Fauna, and the First Division (Fig. 5). Recently, Pagnac (2009) has proposed a system of biostratigraphic Interval Zones to describe the succession beginning at the base of the Green Hills Fauna, as shown in Fig. 7. Lindsay (1972) also proposed Assemblage Zones for the same span based on rodents and other small mammals. The following discussion is designed as a brief overview of the faunal mammal succession of the Barstow Formation in terms of these chronologic systems.

Red Division Fauna

The basal Owl Conglomerate Member (Fig. 4) is largely unfossiliferous, but the Red Division Fauna occurs in the uppermost part of this unit (Fig. 4), about 30 m below the Rak Tuff that defines (in part) its upper boundary. The

Red Division Fauna contains the mesodont horse *Parapliohippus carrizoensis* (Kelly, 1995), the oreodont *Merychys relictus fletcheri* Schultz and Falkenbach, 1947 (holotype), the miolabine camels *Paramiolabis tenuis* and *P. singularis* (Pagnac, 2005), and the pronghorn *Meryceros* sp. (referred to *M. joraki*, but not represented by horn cores).

The Red Division Fauna is one of the main late Hemingfordian reference faunas of this part of the Mojave Desert. The Rak Tuff that caps the unit is dated at about 16.3 ± 0.3 Ma; the Red Tuff near the local exposed base of the Owl Conglomerate Member is dated at about 19.3 ± 0.02 Ma. A fault separates the Red Tuff from overlying parts of the formation, however (Woodburne et al., 1990). Based on paleomagnetic analysis, the bulk of the Owl Conglomerate Member apparently correlates to the reversed (lower) part of chron C5C in the Magnetic Polarity Time Scale (MPTS) (see Fig. 6).

Rak Division Fauna

The middle member of the Barstow Formation is between 470 and 570 m thick. It begins just above the Rak Tuff and its equivalents and, in the main part of the outcrop area is composed of finer-grained, basinal, facies relative to the underlying Owl Conglomerate Member. A number of tuff beds occur in the middle member, with the Oreodont Tuff being an important regional marker for mapping purposes. As shown in Figures 4 and 6, the Rak Division Fauna occurs in the lower 265 m of the middle member. The Rak Division Fauna was newly recognized by Woodburne and others (1990), and separated from the upper part of the Red Division Fauna as used by Woodburne and Tedford (1982).

The Rak Division Fauna includes the canid *Tomarctus* cf. *T. rurestris*; the amphicyonid *Amphicyon* cf. *A. sinapius*; a rhino, *Aphelops*; the miolabine camel *Paramiolabis tenuis* (the range zone of which continues upward from the Owl Conglomerate Member), associated with *P. singularis*, a larger form of *Paramiolabis* (Pagnac, 2005). Protolabine camels make their first appearance in the local section, with both *Protolabis* and *Michenia mudhillensis* present, and a large and small species of *Aepycamelus* also appear, along with the continued presence of pronghorns referred to *Meryceros joraki* (but still no horn cores). Alf (1970) described leaf impressions of oak and palm in the Rak Division shale beds, indicating the riparian nature of the vegetation in the areas where the mammals lived.

Lindsay (1991; 1995) reports the presence of *Copemys* sp. and other taxa of late Hemingfordian age in sediments of the Rak Division in sites west of the Rainbow exit loop road.

Woodburne and others (1990; Fig. 6) suggest that the Rak Division Fauna ranges in age from about 16.3–15.9

Ma. MacFadden and others (1990) suggest that it correlates to the upper part of chron C5C of the MPTS (Fig. 6, here) as also in Lourens et al. (2004).

Green Hills Fauna

The upper part of the middle member is more fossiliferous than the lower, and taxa referred to the Green Hills Fauna are obtained here. The Green Hills Fauna is based on taxa that occur in rocks beginning with Steepside Quarry to just below Valley View Quarry, 60 m below the Skyline Tuff, an interval of about 280 m. The Oreodont Tuff, with an isotopic age of 15.8 ± 0.09 Ma, occurs stratigraphically 40 m above Steepside Quarry. This unit is about equivalent to the *Plithocyon barstowensis*/*Aelurodon asthenostylus* Inverval Zone of Pagnac (2009).

Steepside Quarry contains taxa (*Copemys*, *Hemicyon* (*Plithocyon*)) that define the Green Hills Fauna and the beginning of the Barstovian mammal age (Tedford and others, 1987). Note, however, that E.H. Lindsay (1995) has since recovered remains of *Copemys* from the Rak Division Fauna and that Tedford and others (2004) removed *Copemys* from the Barstovian definition.

Taxa of the Green Hills Fauna include the rodents *Copemys* and *Peridiomys*; the bear, *Hemicyon* (*Plithocyon*); the canids *Euoplocyon*, *Cynarctoides*, “*Tomarctus*” *kellogae*, *T. confertus*, and *T. rurestris*; the amphicyonid *Amphicyon* cf. *A. ingens*; the pliohippine *Acritohippus styloodontus* (Pagnac, 2009) and the merychippine *Merychippus* cf. *M. insignis*; the peccaries *Hesperhys*, *Dyseohyus fricki* and *Cynorca occidentale*; the oreodont *Brachycrus buwaldi*; the cervoid *Rakomeryx*; the pronghorns *Merriamoceros* and *Meryceros* cf. *M. joraki* (still no horn cores); and miolabine camels, one closely related to *Miolabis fissidens* and the other species related to “*Miolabis*” *tenuis*. Protolabine camels are represented by *Protolabis* cf. *P. barstowensis* and *Michenia mudhillensis*, both of which are parts of lineages that extend upward from the Rak Division; large and small *Aepycamelus* also continue upward from underlying beds. The strata containing the Green Hills Fauna also record the last local occurrences of *Michenia* (*M. mudhillensis*) and the *Paramiolabis singularis* clade (Pagnac, 2009).

The Green Hills Fauna ranges in age from about 15.9 to 15.4 Ma (Woodburne and others, 1990; Fig. 6), and correlates to chron C5B and the lower part of chron C5AD (MacFadden and others, 1990). Pagnac (2009) indicates that the *Plithocyon barstowensis*/*Aelurodon asthenostylus* Inverval Zone ranges in age from about 16 to 15.2 Ma, and is comparable stratigraphically to the Green Hills Fauna.

Second Division Fauna

The Second Division Fauna occurs in beds that extend from about 60 m below the Skyline Tuff (Figs. 4, 5) to

about 20 m above the tuff, and thus spans the boundary between the middle and upper members of the Barstow Formation. The name is taken from the Frick Laboratory designation for this interval. This unit is about equivalent to the *Aelurodon asthenostylus/Ramoceros brevicornis* Interval Zone of Pagnac (2009; Fig. 5 here).

The Second Division Fauna includes the canids *Aelurodon asthenostylus*, *Cynarctus galushai*, and *Tomarctus brevirostris*, and the pliohippine *Acritohippus stylodontus*. This taxon persists to the Skyline Quarry level, just below the Skyline Tuff, where it coexists with the lowest stratigraphic occurrence of *Scaphohippus intermontanus* (Pagnac, 2009). This represents the last local appearance of pliohippine equids that are so characteristic of the Barstow Formation below the Skyline Tuff, and occurs within the *Ramoceros brevicornis/Megahippus mckennai* Interval Zone of Pagnac (2009). This zone occurs within the uppermost part of Second Division (Fig. 5). The tiny anchitheriine horse *Archaeohippus mourningi* is virtually restricted to the lower part of the *R. brevicornis/M. mckennai* Interval Zone (locally being present to about 10 m above the Skyline Tuff). *Miolabis* and *Protolabis barstowensis* continue through this interval (Pagnac, 2009). Significantly, the Green Hills Fauna taxa *Brachycrus*, *Rakomeryx*, and *Merriamoceros* are lacking here. Merycodont horn cores are present for the first time in the *R. brevicornis/M. mckennai* interval.

The Second Division Fauna ranges in age from about 15.4–14.8 Ma (Woodburne and others, 1990; Fig. 6), and correlates to chron C5AD in the MPTS (MacFadden and others, 1990). C.C. Swisher (personal commun., 1991) reports a $40\text{Ar}/39\text{Ar}$ age of 15.26 ± 0.04 Ma on four plagioclase crystals and an age of 15.28 ± 0.01 Ma on two anorthoclase crystals from the Valley View Tuff (new name) at about the 660 m elevation in the Barstow composite magnetic polarity zonation (MacFadden and others, 1990). This is represented in Woodburne (1996) as 15.27 ± 0.03 Ma. The Valley View Tuff is stratigraphically just below the R7/N7 boundary (see Fig. 6).

The *Aelurodon asthenostylus/Ramoceros brevicornis* Interval Zone ranges from about 15.3 to 15.0 Ma. The *Ramoceros brevicornis/Megahippus mckennai* Interval Zone ranges from 15.0 to 14.8 Ma (Fig. 5).

Barstow Fauna

The youngest assemblage of the Barstow Formation, the Barstow Fauna, begins with the lowest stratigraphic occurrence of proboscideans at or but a short stratigraphic distance below the Dated Tuff, about at the level of the Frick Laboratory New Year Quarry (First Division Fauna; Fig. 5) and extends to the top of the Barstow Formation, an interval of about 225 m. The Barstow Fauna thus begins about 10–20 m above the base of the upper member

of the Barstow Formation. This unit is equivalent to the *Megahippus mckennai/Merycodus necatus* Interval Zone of Pagnac (2009).

The upper member includes the conspicuous Skyline Tuff at its base and continues upward for about 200 m. Along with coarse-grained distal facies on the northeast, the upper member typically is composed of finer-grained sandstones, mudstones, local beds of limestone and tuff, an overall more basinal facies. Alf (1970) reports fossil wood from the Skyline Tuff as representing a chaparral flora with juniper, poison oak, and buckbrush.

The Skyline Tuff (Sheppard and Gude, 1969; = “lower marker tuff” of Dibblee, 1968) is too fine-grained and altered to be amenable to isotopic age analysis, but the “Dated” and Hemicyon tuffs and the “Lapilli Sandstone” have yielded valuable isotopic ages. The Dated Tuff of Sheppard and Gude (1969) is a brown biotite-crystal tuff with characteristic casts of mud cracks on its lower surface. This tuff occurs 10–30 m above the Skyline Tuff, and yields an isotopic age of 14.8 ± 0.09 Ma (contra 15.5 Ma; Evernden and others, 1964). The Hemicyon Tuff occurs about 80 m stratigraphically above the Skyline and is dated at 14.0 ± 0.1 Ma. The Lapilli Sandstone (Lindsay, 1972) occurs near the top of the Barstow Formation, about 150 m stratigraphically above the Skyline Tuff, and is 13.4 ± 0.2 Ma old (Figs. 4, 6).

Taxa of the Barstow Fauna include the first local occurrence of the rodent *Pliosaccomyx*; the horses *Megahippus mckennai* and *Scaphohippus* (formerly “*Merychippus*”) *sumani* (Pagnac, 2009); the mastodons *Miomastodon* and *Gomphotherium*; and the canids *Epicyon* and “*Tomarctus*” *temerarius*, and the extended range of *Aelurodon*, *Cynarctus*, and a *Tomarctus paulus*; the ursid *Hemicyon barstowensis* (high in the unit only); the upper part of the range of the peccary *Dyseohyus fricki*; the restricted occurrence of the oreodont *Mediochoerus mohavensis*; and the pronghorns *Ramoceros* and *Cosoryx*, with *Meryceros joraki* continuing from lower levels (type specimen from the upper unit). The miolabine camels, represented by a large form, *Miolabis* cf. *M. fissidens*, occur in the lower part of the unit and represent the last local occurrence of the group. The protolabine camels are represented by *Protolabis* cf. *P. barstowensis* and a larger form, “*Protolabis*” cf. “*P. inaequidens*.” *Michenia* is not present; giraffe-camels are represented only by *Aepycamelus alexandrae*, and metapodials suggest the appearance of *Procamelus* at these levels.

Hemingfordian—Barstovian Boundary

The position of the Hemingfordian—Barstovian boundary in the Barstow Formation is discussed by Woodburne and others (1990). Tedford and others (1987) defined the beginning of the Barstovian on the basis of the im-

migrants *Copemys* and *Hemicyon* (*Plithocyon*) which occur together at Steepsides Quarry at the base of the Green Hills faunal interval. Lindsay (1995) has subsequently found remains of *Copemys* in the Rak Division. Tedford and others (2004) defined the Barstovian on the first occurrence of *Plithocyon* and the definite occurrence of *Zygodon*. It is important to note that other taxa found in the Green Hills Fauna, such as *Brachycrus*, *Rakomeryx*, *Dyseohyus*, and *Merriamoceros*, characterize the earlier part of the Barstovian mammal age as originally proposed by Wood and others (1941).

The Red Division Fauna contains no taxa in common with the typical late Hemingfordian faunas of Nebraska. The Sheep Creek Tuff that overlies the Sheep Creek Fauna in Nebraska has been dated at 16.3 ± 0.13 Ma (C.C. Swisher, unpubl. data), a date that corresponds with that for the Rak Tuff in the Barstow Formation. Many taxa of the Rak Division Fauna continue into, or show affinities with, those of the Green Hills, whereas others have affinities with those of, or continue upward from, the Red Division.

At the moment, the best evidence for considering the Rak Division Fauna as equivalent in age to late Hemingfordian is the presence of *Paramiolabis tenuis* (which follows up from the Red Division) and *P. singularis*, and the absence of taxa (see above) characteristic of the Barstovian. Although *Copemys* can no longer be used to help define the beginning of the Barstovian, *Hemicyon* (*Plithocyon*) still can and the other taxa of the Green Hills Fauna cited by Wood and others (1941) as characteristic of, or first occurring in, the Barstovian support the proposal that the Green Hills Fauna be included in the Barstovian mammal age.

Calico Mountains

This range is located immediately east and southeast of the Mud Hills (Fig. 3). The Calico Mountains contain a thick sequence of eruptive and intrusive volcanic rock and associated sedimentary deposits of early to medial Miocene age. The units, which locally are complexly folded and are cut by northwest-trending faults, unconformably overlie both intrusive rocks of Mesozoic age and Paleozoic carbonate rocks. The Cenozoic units include the Jackhammer Formation, "Formation of Lead Mountain," syntectonic conglomerates and breccias, Pickhandle Formation, "Lane Mountain volcanics," Barstow Formation, and various alluvial units.

The type locality of the **Jackhammer Formation** is located near Jackhammer Gap in the northwestern Calico Mountains (Fig. 3), and has been discussed above (Mud Hills). The unit also crops out in the Lead Mountain

area, about 5 km east of Barstow (Fig. 3) where Lambert (1987) shows that it is cut by intrusive dikes dated 25.6–0.8 Ma (Fig. 4).

Lambert (1987) also reports that the Jackhammer is unconformably overlain by a unit, indicated here as the "**formation of Lead Mountain**." This unit is composed of a sequence of tuffs, tuff breccias, lava flows, mudflows, and beds of limestone, sandstone, and conglomerate. The interval is about 2,500' (760 m) thick, and composed of a basal unit of rhyodacitic tuff and tuff breccia with individual blocks up to 1 m in diameter; a middle unit of interbedded lapilli ash and tuff, and minor felsite lavas, welded tuff, volcanic mudflows, and tuffaceous sandstones; an upper tuff unit, with interbedded mudflows, basalt, and sandstone. Each of the three units is separated from the others by beds of limestone.

Another unit, here designated as "**syntectonic conglomerate and breccia**" after Dokka and others (1988, Fig. 9) overlies the other units and apparently is considered to relate to tectonic activities associated with the early Miocene interval of extension in this area. If so, the unit could be significantly younger than the other "Lead Mountain" deposits, as shown on Fig. 4.

The **Pickhandle Formation** has been discussed above (Gravel Hills, Mud Hills). In the Calico Mts. the Pickhandle undergoes rapid lateral changes in both lithology and thickness, and becomes more pyroclastic as it is traced from the Mud Hills eastward into its type area, near Pickhandle Pass (Fig. 3). Both McCulloh (1952:123) and Dibblee (1968:20, 22) suggest that the Pickhandle reflects a violent interval of volcanic activity and that the "eruptions were separated by intervals of relative quiescence, during which tuffaceous sandstone, conglomerate, and mudflow deposits accumulated" (McCulloh, 1952:123).

Another unit of "**Lane Mountain Volcanics**," separated from the Lane Mountain Quartz Latite by Burke and others (1982), is dated at 18.1 ± 0.5 Ma. The age derives from each of the two units; one overlies both the Jackhammer and Pickhandle formations; the other intrudes the Pickhandle.

The **Barstow Formation** in the Calico Mountains is exposed in the northwestern, southern, and southeastern parts of the range, and Lambert (1987) maps it in the Lead Mountain area to the south, as well. The unit reaches a thickness of about 1,000 m and consists of a folded sequence of beds in which lacustrine sandstone and shale predominate. Basal beds of limestone are locally present, and intervals of granitic conglomerate occur throughout. Fossil mammals are rare in these deposits, but Dibblee (1970, explanation sheets) notes the local presence of *Scaphohippus intermontanus*, a taxon typical of the Bar-

stow Fauna in the Mud Hills (see above). *Parapliohippus carrizoensis* also is recorded in strata in Sunrise Canyon on the southern part of the range so that, collectively, the Barstow Formation in this range has about the same geochronologic span as the fossiliferous part of the unit in the Mud Hills. Interestingly, Reynolds and Woodburne (2001) report the presence of proboscidean trackways here that pre-date body fossils found in the Mud Hills. The Calico Mountains Barstow Formation appears to represent the lacustrine, basinal, facies of the unit, comparable to the conditions inferred for eastern and northern outcrops in the Mud Hills. We note that the biostratigraphy of the Calico Mountains may be equivalent to the abbreviated sequence in the Yermo Hills.

Daggett Ridge

Rocks considered to be part of the extensional, and syntectonic, regime are found in the western Newberry Mountains (Dokka and others, 1988). Unconformably overlying these are fossil-bearing beds of varicolored sandstone, siltstone, and limestone, capped by a Miocene silcrete paleosol (Reynolds, p. c. 2009). The Daggett Ridge Local Fauna is derived from a lenticular mud-flow breccia with clasts up to 10 cm. in diameter. Dokka and Glazner (1982) refer to the strata as Barstow Formation. We note here that these outcrops are physically well removed from the districts that typically contain this formation (see Fig. 1) and hesitate to assign this name to them.

Reynolds (1991, revisions to taxa in Lindsay and Reynolds, 2008) reports the following taxa from the Daggett Ridge Local Fauna. *Miospermophilus* sp; the heteromyid rodents *Mookomys altifluminus*, *Perognathus minutus*, *P. furlongi*, *Cupidinimus lindsayi*, and *Mioheteromys crowderensis*; a ?amphicyonid carnivore; a ?canid; a felid; the horses *Archaeohippus mourningi*, *Parapliohippus carrizoensis*, and a large equid, cf. *Acritohippus stylodontus*; a camel; and a pronghorn. The co-occurrence of the horses and *Mioheteromys* suggest a correlation to late Hemingfordian to earliest Barstovian faunas of the Mud Hills, or an age of circa 16 Ma (Fig. 4).

Yermo Hills

These limited and stratigraphically poorly confined deposits of the Barstow Formation crop out in small hills north of Interstate 15, about 6 km east of the Calico Mts. (Figs. 1, 3, 4). The sequence is about 200 m thick, slightly to strongly tilted toward the northeast, and is overlain by early Quaternary alluvium.

The Barstow sediments consist mostly of drab-colored, fine-grained claystone, with minor beds of sandstone, tuff, and limestone. The Argonaut fauna in the middle of the section (Reynolds, 1991, revisions to taxa in Lindsay and

Reynolds, 2008) includes a gasterosteid fish, *Miospermophilus* sp., the heteromyid rodents *Mookomys altifluminus*, *Proheteromys sulculus*, *Perognathus minutus*, *P. furlongi*, *Cupidinimus lindsayi*, *C. halli*, *Peridomys* sp., cf. *P. oregonensis*, *Mioheteromys crowderensis*, *Diprionomys* sp., cf. *D. parvus*, and *Mojavemys lophatus*. Large fossil mammals in the Yermo Local Fauna are of late Hemingfordian and early Barstovian age. The horse, *Parapliohippus carrizoensis*, occurs throughout the unit, but about 10 m below its stratigraphic top, taxa characteristic of the Green Hills Fauna of the Mud Hills are associated with it: *Acritohippus stylodontus*, *Amphicyon* cf. *A. ingens*, *Merriamoceros*, and *Rakomeryx yermoensis*. An age about the same as that of the Daggett Ridge Local Fauna seems likely.

The fine-grained nature of the clastic units exposed here, coupled with the presence of limestone and water-laid tuff, indicate that the Barstow Formation of the Yermo Hills was deposited in a largely lacustrine environment, comparable to the conditions recorded for much of the formation in the Calico Mountains and eastern Mud Hills.

Alvord Mountain

A relatively complex sequence of deposits of early and medial Miocene age is exposed on the eastern flank of a terrane of Mesozoic plutonic and pre-Cretaceous metamorphic rock in the Alvord Mountain, located about 20 km northeast of the Yermo Hills (Fig. 3). The Cenozoic sequence was studied by Byers (1960), who indicates that the late Pleistocene Manix lake beds of the Coyote Lake embayment also crop out in this district. Major reference to the fossil mammals from the Barstow Formation here is Lewis (1964; 1968a; 1968b). The Miocene rock units are: the Clews Fanglomerate, the Alvord Peak Basalt, the Spanish Canyon Formation, and the Barstow Formation (oldest to youngest).

The base of the Cenozoic sequence is the **Clews Fangerlomerate**, the type locality of which is Clews Ridge (Fig. 3). The Clews is about 200 m thick and unconformably overlies plutonic basement rock. The unit was deposited on a former erosional surface with about 300 m of relief, and pinches out westward beneath the Alvord Peak Basalt. This basalt is only of local occurrence and pinches out to the east, where the Clews Fangerlomerate is overlain with possible unconformity by the Spanish Canyon Formation (Fig. 4).

The Clews is mostly composed of red-brown conglomerate with blocks of mafic plutonic rock up to 3 m wide. Most of the unit appears to have been derived locally from the nearby range and to reflect tectonic events in the immediate area.

The age of the Clews is unknown. That shown on Fig. 4 is based on general stratigraphic relationships and the suggestion that the pyroclastic materials of the Spanish Canyon Formation are correlative with those of the Pickhandle Formation to the west.

The **Alvord Peak Basalt**, about 100–125 m thick, is located mainly west of Spanish Canyon (Fig. 3), and thins eastward. The Alvord Peak Basalt conformably overlies the Clews Fanglomerate, and dikes and pipes of basaltic rock locally cut the Clews. The Alvord Peak Basalt is composed of multiple flows, and is usually overlain by the Spanish Canyon Formation but, locally, by the Barstow Formation.

The age of the Alvord Peak Basalt is unknown.

The **Spanish Canyon Formation** is about 100 m thick. It conformably overlies the Alvord Peak Basalt and the Clews Fanglomerate; it is locally conformably overlain by the Barstow Formation. Unconformities found in other places (Byers, 1960:22) indicate that there is a hiatus between the two units. The Spanish Canyon Formation is of limited areal extent and pinches out to the west.

In addition to locally-derived detritus, such as granitic and mafic boulder conglomerate, the Spanish Canyon Formation contains upper flows of olivine basalt, and tuff beds, of which one is correlated with the Peach Spring Tuff (Hillhouse, this volume). Byers (1960) interpreted the tuffs to record a major explosive volcanic event or events in the central Mojave Desert, and suggested on the basis of microscopic analysis that they correlate with the Pickhandle Formation to the west.

The **Barstow Formation** crops out in the eastern part of the Alvord Mountain district and consists of “clastic and tuffaceous beds that are characterized by lithology and vertebrate fossils similar to or identical with those found in the Barstow Formation at the type locality [and] are assigned to [that formation]...” (Byers, 1960:26). The lithic features characteristic of the Barstow Formation in more western districts do, however, differ in many ways from the rocks correlated with the Barstow Formation in Alvord Mountain, and the differences are especially striking when these deposits are compared to the primarily lacustrine sediments in the Yermo Hills, the geographically closest sequence. The Barstow Formation becomes coarser-grained to the northeast (MOW, pers. observation) and in Alvord Mountain is probably a more marginal setting relative to the depositional setting found in sites to the west. In fact, the Barstow becomes coarser-grained to the northeast.

The Barstow Formation in the Alvord Mountain district is about 400 m thick, and is divided into three members. The formation overlaps the Alvord Peak Basalt and is basically unconformable on the Spanish Canyon

Formation as well as on older units; the upper contact is locally gradational and conformable with overlying Pliocene granitic fanglomerate (Byers, 1960:27), but in places there is a slight angular unconformity, and local channels also are present.

The lower member of the formation consists of about 175–200 m of interbedded sandstone and pebble conglomerate, overlain by 50–175 m of tuff, tuffaceous sandstone, siltstone, and volcanic pebble conglomerate. Three flows of olivine basalt occur in the lower 60–100 m of the unit, adjacent to Clews Ridge. C.C. Swisher (personal commun., 1991) has obtained $40\text{Ar}/39\text{Ar}$ ages of 16.47 ± 0.5 Ma on the lower of these basalts near Clews Ridge and 16.6 ± 0.2 Ma on a correlative unit in Spanish Canyon.

P. carrizoensis (= *M. tehachapiensis*) listed in Lewis, 1968b) and *Merychius* (*Metorodon*) *relictus* have been recovered from USGS Loc. D321 (Fig. 4), about 170 m below the middle member of the Barstow Formation, and thus about 30 m above the dated basalt near Clews Ridge. Based on the age of this basalt, this part of the Barstow Formation appears comparable in taxonomy and age to the Red Division in the Mud Hills.

The remains of *Brachycrus buwaldi* have been obtained from a site about 50 m stratigraphically below the middle member, in the general area of USGS D319. See “D319” on Fig. 4. Although reported as occurring at USGS D319 in the middle member of the Barstow Formation by Byers (1960:33) and Lewis (1968b:C78), in association with *A. styloodontus*, R.H. Tedford, who collected the specimens attests to their stratigraphic separation, as shown here (Fig. 4) and in Woodburne and others (1982). *Brachycrus buwaldi* is one of the characteristic elements of the Green Hills Fauna in the Mud Hills (see above), so this part of the formation in the Alvord Mountain district appears to be early Barstovian in age.

The middle member of the Barstow Formation is 7–30 m thick and consists of two or three beds of white lapilli tuff separated by clastic (tuffaceous sandstone) sediments. This section is one of the most fossiliferous in the Barstow Formation of Alvord Mountain.

Based on samples from USGS localities D319 and D301, and on UCR localities, the middle member contains elements of *Brachypsalis* cf. *B. pachycephalus*, *Acritohippus styloodontus*, and *Protolabis barstowensis*. The horse and camel are comparable to those in the Mud Hills that extend from the Green Hills into the Second Division faunal levels which are stratigraphically below the Skyline and Dated tuffs. C.C. Swisher (personal commun., 1991) reports a mean age of 14.4 ± 0.03 Ma on plagioclase and biotite from the upper middle member tuff, suggesting that these species range somewhat later in the Alvord Mountain sequence than in the Mud Hills. Interestingly,

none of the other horses typical of the Barstow Fauna in the Mud Hills (*Scaphohippus intermontanus* or *S. sumani*) have been found in the Alvord succession.

Other fossils, from UCR localities clustered in and within a few meters stratigraphically above and below the tuffaceous middle member, contain *Acritohippus styloodontus*, *Merycodus*, various Camelidae (not all necessarily *Protolabis barstowensis*), and rabbits.

Cady Mountains

Dibblee and Bassett (1966a, b) mapped the regional geology of the area, and indicate that the Tertiary succession here (Fig. 3) is arranged largely peripheral to a central core of pre-Tertiary plutonic rock and metamorphic rock of limited areal extent. Major outcrops of Miocene sedimentary rock occur on the northern and western flanks of the range but, toward the center, extensive units of volcanic rock are shown. As discussed below, these rocks are likely mostly older than the main sedimentary succession, and pertain to a volcanic center that was active as early as 23–24 m.y. ago. Nonmarine fluvial, volcanoclastic, and pyroclastic deposition (Hector Formation) began locally by about 23 Ma, and persisted until at least 16 Ma, prior to the development of a major hiatus. The Peach Spring Tuff, a regional marker unit dated at about 18.5 Ma (Nielsen and others, 1990), is present in this sequence as well.

Subsequent to Hector Formation deposition, the region underwent at least 26° of clockwise rotation (MacFadden and others, 1990), but geological evidence younger than the Hector deposits is sparse or, if present, poorly dated. The Mojave River Formation and the Manix Formation (Jefferson, 1999) and precursors show that the Central Mojave Desert was occupied by one or more regional playas or lakes from 2 Ma to about 19,000 yrs. Dokka and Travis (1990) point to the recency of geologic events in this area. The Manix fault, one of a family of west-trending faults with left-lateral separation, experienced a major (ca 6.2) earthquake in 1947 (Richter, 1947; Real and others, 1987).

The following discussion begins with the Tertiary succession of the southwestern Cady Mountains, which contains the geologically oldest biochronological data, and is followed by comments on successions exposed on the northern and eastern flanks of the range.

Southwestern Cady Mountains—Woodburne and others (1974) indicate that the base of the Tertiary succession of the southwestern Cady Mountains begins with andesitic lahar and agglomerate and interbedded tuffaceous sediments that rest unconformably on pre-Tertiary basement rock. As mapped, this succession is not named,

but seems comparable to pre-Hector units mapped in the northeastern Cady Mountains (Williamson, 1980).

The Miocene strata of the **Hector Formation** are about 500 m thick; they unconformably overlie the agglomerate units and are unconformably overlain by Quaternary alluvium. The Hector sediments (Woodburne and others, 1974) are divided into a largely tuffaceous and volcanoclastic lower sequence (below the marker tuff bed on Fig. 4) and an upper succession that, although of tuffaceous matrix, included many fewer beds of tuff. The Peach Spring Tuff may be present in this section, however (Woodburne, 1998). A tuff near the top of the lower sequence has been dated (revised [Dalrymple, 1979] from Evernden and others, 1964) at about 21.6 Ma (Fig. 4). If the Peach Spring Tuff is present in the upper sequence, that would indicate an age of about 18.5 Ma for that part of it, as shown relative to the Hector beds in the northern Cady Mountains (Woodburne, 1998; Figs. 4 and 7 here).

Fossil mammals from the **Black Butte Mine Local Fauna** occur in the lower part of the succession, below the marker tuff (Figs. 4, 7) and appear to be of late Arikareean (Ar3) age (Woodburne, 1998). Taxa include the oreodont *Merychius calaminthus* and a camel, *Stenomylus* cf. *S. hitchcocki*, which ranges above the dated tuff, but not above the marker tuff.

The **Logan Mine Local Fauna** occurs in beds that stratigraphically overlie those of Black Butte Mine, and includes camels, *Protolabis* and *Michenia*, cf. *M. agatensis*; an oreodont, *Phenacocoelus* cf. *P. leptoscelos*; the amphicyonid carnivore cf. *Daphoenodon*; and the weasel, *Promartes*. Woodburne and others (1974) originally considered these taxa to be of early Hemingfordian age, and although revisions in the stratigraphy of the pertinent beds and faunas in Nebraska (Hunt, 1978; 1981) suggested that all of the Hector Formation local faunas were of late Arikareean age Woodburne (1998) re-affirmed the early Hemingfordian (He1) age of the Logan Mine L.F. (Fig. 7).

Regionally, faunas of late Arikareean age are calibrated as ranging from 23–18 Ma (Tedford and others, 2004). Woodburne (1998) indicated that the Black Butte Mine Local Fauna is correlated to Ar3, the Logan Mine Local Fauna to Ar4.

Northern and eastern Cady Mountains—The geology, paleontology, and isotopic age of the Hector Formation in the northern Cady Mountains have been described by Miller (1980), Moseley (1978), and Williamson (1980; see Fig. 3 for locations). Williamson (1980) studied the dominantly extrusive volcanic succession north of the Cady fault in the eastern part of the range. A major unconformity apparently separates this volcanic succession from the Hector Formation, as mapped by Williamson (1980; Fig. 4 here).

Miller (1980) indicates that the Hector Formation as exposed on the northern flank of the range contains a succession of tuffaceous and volcanoclastic sedimentary rock, interbedded with a flow of olivine basalt and a distinctive ignimbrite unit. The deposits have been gently deformed into a shallow syncline that trends east–west and plunges shallowly to the east. The beds have been cut by a number of faults that trend generally northward and show dip-slip throw of about 100 m at most.

The two volcanic units are used to divide the sequence into three intervals. The lowest begins with a succession of tuffaceous sediment, tuff, and derivative clastic rock that is about 100 m thick. Tuffs near the local base of this sequence have been dated at 22.6 ± 0.4 Ma (Miller, 1980), and a basalt that overlies this interval has been dated at 18.6 Ma.

The middle part of the Hector Formation in this area continues upward from the just-mentioned basalt for about 85 m, where it is capped by a distinctive ignimbrite dated at 17.9 ± 0.3 Ma (Miller, 1980), but now considered to belong to the Peach Spring Tuff, dated at 18.5 Ma (Woodburne, 1998; Fig. 7 here). The rocks of this interval are generally similar to those of the lower part, and are mostly thin-bedded intervals plausibly of lacustrine accumulation. To the southwest, however, coarser-grained, more marginal facies prevail. Based on an analysis of the paleomagnetic succession reported for the Hector Formation in the northern Cady Mountains (MacFadden and others, 1990), Woodburne (1998) suggested that the middle part of the Hector Formation in this area is separated from its lower part by a hiatus (Figs. 4, 7) that accounts for missing parts of chrons C6, C6A and C6AA, or from about 19 to 21.8 Ma.

The **Lower Cady Mountains** Local Fauna, of early Hemingfordian age, is represented by sparse but significant fossil mammals that occur in this part of the section about 15 m above the basalt dated at 18.6 ± 0.2 Ma, but considered to actually be between 19 and 20 m.y. old (Woodburne, 1998). This fauna, which contains sparse remains of Aletomerycinae and Merychyinae, is separated from specimens of *Merychyus* cf. *M. calaminthus*, which occur about 21–24 m below the basalt. Woodburne (1998) removed these oreodont specimens from the Lower Cady Mountains Local Fauna, but did not provide a separate faunal name for them. *M. calaminthus* is compatible with a late Arikareean age, correlated to Ar3 by Woodburne (1998), and occurs about 70 m stratigraphically above an ash dated at 22.6 Ma.

This Lower Cady Mountains L.F. material is only generally diagnostic, but the association is also found in the early Hemingfordian Boron Local Fauna to the west (Whistler, 1984; Whistler, 1991), and correlated as He1

(Woodburne, 1998; Fig. 7 here).

The upper part of the Hector Formation begins with a brown to brownish gray to orange-brown and locally bluish-gray ignimbrite about 17m thick that is correlated with the Peach Spring Tuff (Woodburne, 1998). The upper unit that rests conformably on the unit is about 130 m thick, and consists of tan to greenish gray and tan tuffaceous siltstone, mudstone, tuff, and local lenses of conglomerate. A thick unit of nonmarine limestone occurs near the top of the sequence.

The **Upper Cady Mountains** Local Fauna occurs just below the massive limestone interval, about 65 m stratigraphically above the ignimbrite. The faunal elements, that occur within a range of a few tens of meters, include: *Paraplihippus carrizoensis*; the beaver cf. *Anchitheriomys*; a rhino, cf. *Diceratherium*; the dog *Tomarctus* cf. *T. hippophagus*; a rodent, *Proheteromys sulculus*; the camels *Paramiolabis tenuis* (Kelly, 1992) and cf. *Aepyamelus*; and the antelope *Merycodus*. The assemblage as a whole suggests a late Hemingfordian age for the Upper Cady Mountains Local Fauna, or about 16–17 Ma old (Woodburne, 1998; Fig. 7 here).

The **Afton Canyon** district, mapped by Moseley (1978), consists of a 570 m thick sequence that correlates best with the lower and middle parts of the Hector Formation of Miller (1980). The rocks form a sequence of alluvial and lacustrine strata, interbedded with laharic breccia, and a sequence of volcanic rocks, air-fall tuff, a major ignimbrite, and flows of basaltic to andesitic lava. No fossils have been reported, but coarser-grained facies suggest that the local basin margin was not far away to the east.

Avawatz Mountains

The Avawatz Mountains are north of Baker (Fig. 1) at the junction of the Garlock fault and the Death Valley fault zone (Fig. 2). The range contains a variety of Precambrian metasedimentary, Paleozoic clastic and carbonate units, Mesozoic granitic rock, and Tertiary sediments (e.g., Spencer, 1990a). Along the southern and western flank, the Avawatz Formation crops out and locally contains terrestrial mammal fossils.

The Avawatz Formation is several thousand feet thick. Spencer (1990b) divides the unit into an upper and lower sandstone and siltstone facies and a middle facies of conglomerate. The lower part of the formation, at least 1,000 m thick, contains a tuff near its base that has been dated at 20.9 ± 0.2 Ma, and accumulated mostly west of the Arastre Spring fault. This fault now bounds the western part of the range, and apparently was active during the deposition of the lower Avawatz Formation. Those beds include clasts not derivable from the present range, but must have come from sources to the east, now buried or displaced.

The Avawatz thus records events both on and adjacent to the present boundary of the Mojave Desert Province.

The upper part of the Avawatz Formation contains sandstone, locally gypsiferous siltstone, and beds of thin, unwelded tuff. Spectacular areally extensive sheets of breccia are also interbedded in the section and can form convenient lithic as well as temporal markers.

The **Avawatz** Local Fauna of Clarendonian age has been known since about 1937 (Henshaw, 1939). Evernden and others (1964) recorded a K-Ar date of about 11.0 Ma from a tuff that apparently is stratigraphically about 30 m above C.I.T. 267, the main fossil-producing locality in the upper part of the formation.

According to Henshaw (1939, with revisions by Barnosky, 1986), the Avawatz Local Fauna consists of the felid *Pseudaelurus intrepidus*; the rodents *Cupidinimus avawatzensis* and *Copemys dentalis*; the rabbit *Hypolagus* sp; the horse *Pliohippus* sp; the camel *Procamelus coartatus*; and the antelope *Meryceros* cf. *M. cerroensis*.

References cited

- Albright, L.B. 1999. Biostratigraphy and vertebrate paleontology of the San Timoteo Badlands, Southern California. University of California Publications in Geological Sciences, 144.
- Alf, R.M. 1970. A preliminary report on a Miocene flora from the Barstow Formation, Barstow, California. Southern California Academy of Sciences, Bulletin 69:183-188.
- Armstrong, R.L., and R.E. Higgins. 1973. K-Ar dating of the beginning of Tertiary volcanism in the Mojave Desert, California. Geological Society of America Bulletin 84:1095-1100.
- Atwater, T. 1970. Implications of plate tectonics for the Cenozoic tectonic evolution of western North America. Geological Society of America Bulletin 81:3513-3536.
- Baker, C.L. 1911. Notes on the later Cenozoic history of the Mohave Desert region in southeastern California. University of California Publications in Geological Sciences 6:333-383.
- Barnosky, A.D. 1986. New species of the Miocene rodent *Cupidinimus* (Heteromyidae) and some evolutionary relationships within the genus. Journal of Vertebrate Paleontology 6(1): 46-64.
- Bartley, J.M., and A.F. Glazner. 1991. En echelon Miocene rifting in the southwestern United States and model of vertical-axis rotation in continental extension. Geology 19:1165-1168.
- Bartley, J.M., A.F. Glazner, and E.R. Schermer. 1990. North-south contraction of the Mojave Block and strike-slip tectonics in southern California. Science 248:1398-1401.
- Buesch, D.C. 1994. Feldspar geochemistry of four Miocene ignimbrites in southeastern Calif. and western Ariz., in Sherrod, D.R., and Nielson, J.E., eds., Tertiary stratigraphy of highly extended terranes, California, Arizona, and Nevada. U.S. Geological Survey Bulletin 2053: 55-69.
- Burke, D.B., J.W. Hillhouse, E.H. McKee, S.T. Miller, and J.L. Morton. 1982. Cenozoic rocks in the Barstow basin area of southern California—Stratigraphic relations, radiometric ages, and paleomagnetism. U.S. Geological Survey Bulletin 1529-E: E1-E16.
- Byers, F.M., Jr. 1960. Geology of the Alvord Mountain quadrangle, San Bernardino County, California. U.S. Geological Survey Bulletin 1089-A: 1-71.
- Carter, J.N., B.P. Luyendyk, and R.R. Terres. 1987. Neogene clockwise rotation of the eastern Transverse Ranges, California, suggested by paleomagnetic vectors. Geological Society of America Bulletin 98:199-206.
- Christiansen, R.L., and P.W. Lipman. 1972. Cenozoic volcanism and plate tectonic evolution of the western United States. II. Late Cenozoic. Royal Society of London Philosophical Transactions, A, 271:249-284.
- Cox, B.F. 1982. Stratigraphy, sedimentology, and structure of the Goler Formation (Paleocene), El Paso Mountains, California: implications for Paleogene tectonism on the Garlock fault zone. University of California, Riverside, Department of Earth Sciences, unpubl. Masters thesis.
- _____. 1987. Stratigraphy, depositional environments, and paleotectonics of the Paleocene and Eocene Goler Formation, El Paso Mountains, California—Geologic summary and road log, pp. 1-30, in Cox, B.F., ed., Basin analysis and paleontology of the Paleocene and Eocene Goler Formation, El Paso Mountains, California. Society of Economic Paleontologists and Mineralogists, Book 57.
- Cox, B.F. and M.F. Diggles. 1986. Geologic map of the El Paso Mountains Wilderness Study Area, Kern County, California. U.S. Geological Survey Miscellaneous Field Studies Map MF-1827.
- Cummings, D. 1976. Theory of plasticity applied to faulting, Mojave Desert, southern California. Geological Society of America Bulletin 87: 720-725.
- Dalrymple, G.B. 1979. Critical tables for conversion of K-Ar ages from old to new constants. Geology 7:558-560.
- Dibblee, T.W., Jr. 1952. Geology of the Saltdale quadrangle, Kern County, California. California Division of Mines and Geology Bulletin 160:7-43.
- _____. 1967. Areal geology of the western Mojave Desert, California. U.S. Geological Survey Professional Paper 522:153p.
- _____. 1968. Geology of the Opal Mountain and Fremont Peak quadrangles, California. California Division of Mines and Geology Bulletin 188: 64p.
- _____. 1970. Geologic map of the Daggett Quadrangle, San Bernardino County, California. U.S. Geological Survey Miscellaneous Geological Investigations Map I-592, 1:62,500.
- Dibblee, T.W. Jr. and A.M. Bassett. 1966a. Geologic Map of the Newberry Quadrangle, San Bernardino County, California. U.S. Geological Survey Miscellaneous Geological Investigations Map I-461, 1:62,500.
- _____. 1966b. Geologic Map of the Cady Mountains quadrangle, San Bernardino County, California. U.S. Geological Survey Miscellaneous Geological Investigations Map I-467, 1:62,500.
- Dickinson, W.R. 1996. Kinematics of transrotational tectonism in the California Transverse Ranges and its contribution to cumulative slip along the San Andreas transform fault system. Geological Society of America Special Paper 305:1-46.
- Dokka, R.K. 1983. Displacements on late Cenozoic strike-slip faults of the central Mojave Desert, California. Geology, 11:305-308.
- _____. 1986. Patterns and modes of early Miocene crustal extension, central Mojave Desert, California, pp. 75-95, in L. Mayer, ed., Extensional tectonics of the southwestern United States: a perspective on processes and kinematics. Geological Society of America Special Paper 208.
- _____. 1989. The Mojave Extensional Belt of southern California. Tectonics, 8:363-390.
- Dokka, R.K. and A.F. Glazner. 1982. Aspects of early Miocene extension of the central Mojave Desert, pp. 31-45, in Cooper J.D. (compiler), Volume and guidebook, "Geologic Excursions in the California Desert," 78th Annual Meeting, Cordilleran Section, Geological Society of America

- Dokka, R.K., D.J. Henry, T.M. Ross, A.K. Baksi, J. Lambert, C.J. Travis, S.M. Jones, C. Jacobson, M.M. McCurry, M.O. Woodburne, and J.P. Ford. 1991. Aspects of the Mesozoic and Cenozoic geologic evolution of the Mojave Desert. Geological Society of America, San Diego, Annual Meeting, field trip guidebook.
- Dokka, R.K., M. McCurry, M.O. Woodburne, E.G. Frost, and D. Okaya. 1988. Field guide to the Cenozoic crustal structure of the Mojave Desert, pp. 21-44, in *This extended land, geological journeys in the southern Basin and Range*, Weide D.L and Faber M.L., (eds.), Cordilleran Section, Geological Society of America, Las Vegas.
- Dokka, R.K., and Macaluso, K.Y. 2001. Topographic effects of the Eastern California Shear Zone in the Mojave Desert. *Journal of Geophysical Research* 106 (B12):30,625-30,644.
- Dokka, R.K., and Ross, T.M. 1995. Collapse of southwestern North America and the evolution of early Miocene detachment faults, metamorphic core complexes, the Sierra Nevada orocline, and the San Andreas fault system. *Geology* 23 (12):1075-1078.
- Dokka, R.K., T.M. Ross, and G. Lu. 1998. The Trans Mojave-Sierran shear zone and its role in early Miocene collapse of southwestern North America, in Holdsworth, R.E., Strachan, R.A., and Dewey, J.F., eds., *Continental Transpressional and Transtensional Tectonics*. Geological Society, London, Special Publications 135:183-202.
- Dokka, R.K. and C.J. Travis. 1990a. Late Cenozoic strike-slip faulting in the Mojave Desert, California. *Tectonics* 9:311-340.
- Dokka, R.K., and C.J. Travis. 1990b. Role of the Eastern California shear zone in accommodating Pacific-North American plate motion. *Geophysics Research Letters* 19:1323-1326.
- Durrell, C. 1953. Geological investigations of strontium deposits in southern California. California Division of Mines and Geology Special Report 32:23-36.
- Evernden, J.F., D.E. Savage, G.H. Curtis, and G.T. James. 1964. Potassium-argon dates and the Cenozoic mammalian geochronology of North America. *American Journal of Science* 262:145-198.
- Fillmore, R.P. 1993. Sedimentation and extensional basin development in a Miocene metamorphic core complex setting, Alvord Mountain, central Mojave Desert, California, USA. *Sedimentology* 40:721-742.
- Fillmore, R.P., and J.D. Walker. 1996. Evolution of a supradetachment extensional basin: the Lower Miocene Pickhandle basin, central Mojave Desert, California, in Beratan, K.D., ed., *Reconstructing the history of Basin and Range extension using sedimentology and stratigraphy*. Geological Society of America Special Paper 303: 107-126.
- Frick, C. 1926. The Hemicyoninae and an American Tertiary Bear. *American Museum of Natural History Bulletin* 69: 669p.
- Garfunkel, Z. 1974. Model for the late Cenozoic tectonic history of the Mojave Desert, California, and for its relation to adjacent regions. *Geological Society of America Bulletin* 85:1931-1944.
- Glazner, A.F., J.M. Bartley, and J.D. Walker. 1989. Magnitude and significance of Miocene crustal extension in the central Mojave Desert, California. *Geology* 17:50-53.
- Glazner, A.F., J.D. Walker, J.M. Bartley, and J.M. Fletcher. 2002. Cenozoic Evolution of the Mojave Block of Southern California. *Geological Society of America Memoir* 195: 19-41.
- Golombek, M.P. and L.L. Brown. 1988. Clockwise rotation of the western Mojave Desert. *Geology* 16:126-130.
- Hall, E.R. 1930. Rodents and lagomorphs from the Barstow beds of southern California. University of California Publications in Geological Sciences 19:313-318.
- Henshaw, P.C. 1939. A Tertiary mammalian fauna from the Avawatz Mountains, San Bernardino County, California. Carnegie Institute of Washington Publication 514:1-30.
- Hewett, R.H. 1954. General geology of the Mojave Desert region, California, pp. 5-20, in Jahns, R.H., ed., *Geology of Southern California*. California Division of Mines and Geology Bulletin 170, Ch. II.
- Hershey, O.H. 1902. Some Tertiary formations of southern California. *American Geologist* 29:349-372.
- Hillhouse, J.W., D.M. Miller, and B.D. Turrin. 2010. Correlation of the Miocene Peach Spring Tuff with the geomagnetic polarity time scale and new constraints on tectonic rotations in the Mojave Desert, California, in *Overboard in the Mojave: 20 million years of lakes and wetlands*, R.E. Reynolds and D.M. Miller, eds. California State University Fullerton, Desert Studies Consortium, this volume.
- Hornafius, J.S. 1985. Neogene Tectonic Rotation of the Santa Ynez Range, western Transverse Ranges, California, suggested by paleomagnetic investigation of the Monterey Formation. *Journal of Geophysical Research* 90, B14:12,505-12,522.
- Hornafius, J.S., B.P. Luyendyk, R.R. Terres, and M.J. Kamerling. 1986. Timing and extent of Neogene tectonic rotation in the western Transverse Ranges, California. *Geological Society of America Bulletin* 97:1476-1487.
- Hunt, R.M., Jr. 1978. Depositional setting of a Miocene mammal assemblage, Sioux County, Nebraska. *Paleogeography, Palaeoclimatology, Paleoecology* 24:1-52.
- _____. 1981. Geology and vertebrate paleontology of the Agate Fossil Beds National Monument and surrounding region, Sioux County, Nebraska. *National Geographic Society Research Reports* 13:263-285.
- Ingersoll, R.V., Devaney, K.A., Geslin, J.K., Cavazza, W., Diamond, D.S., Heins, W.A., Jagiello, K.J., Marsaglia, K.M., Paylor, E.D., II, and Short, P.F. 1996. The Mud Hills, Mojave Desert, California: structure, stratigraphy, and sedimentology of a rapidly extended terrain. *Geological Society of America Special Paper* 303:61-84.
- Jefferson, G.T. 1985. Stratigraphy and geologic history of the Pleistocene Lake Manix formation, central Mojave Desert, California, pp. 157-170, in R.E. Reynolds, compiler, *Geologic Investigations along Interstate 15; Cajon Pass to Manix Lake, California*. San Bernardino County Museum Special Publication.
- Jefferson, G.T. 1999. Age and stratigraphy of Lake Manix Basin. pp. 109-112, in R.E. Reynolds and J. Reynolds, compilers, *San Bernardino County Museum Assn, Quarterly*, Vol 46(3).
- Kamerling, M.J., and B.P. Luyendyk. 1979. Tectonic rotations of the Santa Monica Mountains Region, western Transverse Ranges, California, suggested by paleomagnetic vectors. *Geological Society of America Bulletin* 90:1331-1337.
- _____. 1985. Paleomagnetism and Neogene tectonics of the northern channel islands, California. *Journal of Geophysical Research* 90, B14:12,485-12,502.
- Kelly, T.S. 1992. New middle Miocene camels from the Caliente Formation, Cuyama Valley badlands, California. *PaleoBios* 13(32):1-22.
- _____. 1995. New Miocene horses from the Caliente Formation, Cuyama Valley Badlands, California. *Natural History Museum of Los Angeles County Contributions in Science* 455: 1-33.
- Lambert, J.R. 1987. Middle Tertiary structure and stratigraphy of the southern Mitchell Range, San Bernardino County, California. Louisiana State University, Department of Geology and Geophysics (Master's thesis).
- Lewis, G.E. 1964. Miocene vertebrates of the Barstow Formation in southern California. U.S. Geological Survey Professional Paper 475-D: D18-D23.
- _____. 1968a. Stratigraphic paleontology of the Barstow Formation in the Alvord Mountain area, San Bernardino County, California.

- U.S. Geological Survey Professional Paper 600-C: C75-C79.
- _____. 1968b. Stratigraphic paleontology of the Barstow Formation in the Mud Hills area, San Bernardino County, California, in Dibblee, T.W., Jr., 1968. *Geology of the Opal Mountain and Fremont Peak quadrangles, California*. California Division of Mines and Geology Bulletin, 188:64p.
- Lindsay, E.H. 1972. Small mammal fossils from the Barstow Formation, California. University of California Publications in Geological Sciences 93:104p.
- _____. 1991. Small mammals near the He/Ba boundary. San Bernardino County Museum Association Quarterly 38(3,4):78-79.
- _____. 1995. Copemys and the Barstovian/Hemingfordian boundary. *Journal of Vertebrate Paleontology* 15(2):357-365.
- _____. and R. E. Reynolds, 2008. Heteromyid rodents from Miocene faunas in the Mojave Desert, Southern California. *Natural History Museum of Los Angeles County Science Series No 41*: 213-236.
- Lipman, P.W. and others. 1972. Cenozoic volcanism and plate tectonic evolution of the western United States. I. Early and middle Cenozoic. *Royal Society of London Philosophical Transactions A*, 271:217-248.
- Loomis, D.B., and D.W. Burbank. 1988. The stratigraphic evolution of the El Paso basin, southern California: Implications for the Miocene development of the Garlock fault and uplift of the Sierra Nevada. *Geological Society of America Bulletin* 100:12-28.
- Lourens, L., Hilgen, F., Shackleton, N.J., Laskar, J., and Wilson, D. 2004. The Neogene Period, pp. 409-440, in Gradstein, F., Ogg, J., and Smith, A., eds., *A geologic time scale*. Cambridge University Press, Cambridge.
- Luyendyk, B.P., and J.S. Hornafius. 1987. Neogene crustal rotations, fault slip, and basin development in southern California, pp. 259-283, in *Cenozoic Basin Development of Coastal California*, Ingersoll, R.V. and Ernst, W.G., eds. Englewood Cliffs, Prentice-Hall.
- Luyendyk, B.P., M.J. Kamerling, and R. Terres. 1980. Geometric model for Neogene crustal rotations in southern California. *Geological Society of America Bulletin* 91:211-217.
- Luyendyk, B.P., M.J. Kamerling, R.R. Terres, and J.S. Hornafius. 1985. Simple shear of southern California during Neogene time suggested by paleomagnetic declinations. *Journal of Geophysical Research*, 90; B14:12,454-12,466.
- May, S.R. 191a. *Geology and mammalian paleontology of the Horned Toad Hills, Mojave Desert, California*. University of California, Department of Earth Sciences, unpub. Masters thesis.
- _____. 1981b. *Repomys (Mammalia: Rodentia Gen. Nov.) from the late Neogene of California and Nevada*. *Journal of Vertebrate Paleontology* 1:219-230.
- May, S.R., and C.A. Repenning. 1982. New evidence for the age of the Mount Eden Fauna, southern California. *Journal of Vertebrate Paleontology* 2:109-113.
- McCulloh, T.H. 1954. *Geology of the southern half of the Lane Mountain quadrangle, California*. University of California, Los Angeles, Department of Geology (Ph.D. Thesis).
- McDougall, K. 1987. Foraminiferal biostratigraphy and paleoecology of marine deposits, Goler Formation, California, pp. 43-67, in Cox, B.F., ed., *Basin analysis and paleontology of the Paleocene and Eocene Goler Formation, El Paso Mountains, California*. Society of Economic Paleontologists and Mineralogists, Book 57.
- McKenna, M.C., J.H. Hutchison and J.H. Hartman. 1987. Paleocene vertebrates and nonmarine mollusca from the Goler Formation, California, pp. 31-42, in Cox, B.F., ed., *Basin analysis and paleontology of the Paleocene and Eocene Goler Formation, El Paso Mountains, California*. Society of Economic Paleontologists and Mineralogists, Book 57.
- MacFadden, B.J., N.D. Opdyke, C.C. Swisher, III, and M.O. Woodburne. 1990a. Paleomagnetism, geochronology and possible tectonic rotation of the middle Miocene Barstow Formation, Mojave Desert, California. *Geological Society of America Bulletin* 102: 478-493.
- MacFadden, B.J., N.D. Opdyke, and M.O. Woodburne. 1990b. Paleomagnetism and Neogene Clockwise Rotation of the Northern Cady Mountains, Mojave Desert of Southern California. *Journal of Geophysical Research* 95(B4):4597-4608.
- Meisling, K.E., and R.J. Weldon. 1989. Late Cenozoic tectonics of the northwestern San Bernardino Mountains. *Geological Society of America Bulletin* 101:106-128.
- Merriam, J.C. 1913. *New Anchitheriine horses from the Tertiary of the Great Basin area*. University of California Publications in Geological Sciences 7:419-434.
- _____. 1915. *New horses from the Miocene and Pliocene of California*. University of California Publications in Geological Sciences 9:49-58.
- _____. 1919. *Tertiary mammalian faunas of the Mohave Desert*. University of California Publications in Geological Sciences 86:437-585.
- Miller, D.M., S.L. Dudash, H.L. Green, D.J. Lidke, L. Amoroso, G.A. Phelps, and K.M. Schmidt, 2007. A new Quaternary view of northern Mojave Desert tectonics suggests changing fault patterns during the late Pleistocene, pp. 157-171, in Miller, D.M. and Valin, Z.C., eds., *Geomorphology and tectonics at the intersection of Silurian and Death Valleys, southern California*: U.S. Geological Survey Open-File Report 2007-1424.
- Miller, D.M., and J.L. Yount. 2002. Late Cenozoic tectonic evolution of the north-central Mojave Desert inferred from fault history and physiographic evolution of the Fort Irwin area, California: *Geological Society of America Memoir* 195, p. 173-197.
- Miller, S.T. 1980. *Geology and mammalian biostratigraphy of a part of the northern Cady Mountains, California*. U.S. Geological Survey Open File Report 80-978:121p.
- Morton, D.M., and Miller, F.K. 2003. Preliminary geologic map of the San Bernardino 30'x60' quadrangle, California. United States Geological Survey Open-File Report 03-293, <http://geopubs.wr.usgs.gov/open-file/of03-293/>.
- Moseley, C.G. 1978. *Geology of a portion of the northern Cady Mountains, Mojave Desert, California*. University of California, Riverside, Department of Earth Sciences (Master's thesis).
- Nielsen, J.E., D.R. Lux, G.B. Dalrymple, and A.F. Glazner. 1990. Age of the Peach Spring Tuff, southeastern California and western Arizona. *Journal of Geophysical Research*, 95:571-580.
- Oskin, M., and Iriondo, A. 2004. Large-magnitude transient strain accumulation on the Blackwater fault, eastern California shear zone. *Geology* 32(4): 313-316.
- Pagnac, D. 2005. *New camels (Mammalia: Artiodactyla) from the Barstow Formation (middle Miocene), San Bernardino County, California*. *PaleoBios* 25: 19-31.
- Pagnac, D. 2006. *Scaphohippus, a new genus of horse (Mammalia: Equidae) from the Barstow Formation of California*. *Journal of Mammalian Evolution* 13: 37-61.
- Pagnac, D. 2009. Revised large mammal biostratigraphy and biochronology of the Barstow Formation (middle Miocene), California. *PaleoBios* 29(2): 48-59.
- Pampeyan, E.H., T.L. Holzer, and M.M. Clark. 1988. Modern ground failure in the Garlock fault zone, Fremont Valley, California. *Geological Society of America Bulletin* 100:677-691.
- Quinn, J.P. 1987. *Stratigraphy of the middle Miocene Bopesta Formation, southern Sierra Nevada, California*. *Natural History Museum (Los Angeles), Contributions in Science* 393:1-31.

- Real, C.R., T.R. Topozader and D.L. Parke. 1987. Earthquake epicenter map of California, 1900-1974. California Division of Mines and Geology, Map Sheet 39.
- Reynolds, R.E., 1991. Hemingfordian/Barstovian Land Mammal Age faunas in the central Mojave Desert, exclusive of the Barstow Fossil Beds, in *Inland Southern California: the last 70 million years*, Woodburne, M.O., Reynolds, R.E., and Whistler, D.P., eds. San Bernardino County Museum Association Quarterly 38(3, 4):88-90.
- and R.L. Reynolds. 1985. Late Pleistocene faunas from Daggett and Yermo, San Bernardino County, California, pp. 175-191, in Reynolds, R.E., compiler, *Geologic Investigations along Interstate 15; Cajon Pass to Manix Lake, California*. San Bernardino County Museum Special Publication.
- ., and Woodburne, M.O., 2001. Review of the Proboscidean datum within the Barstow Formation, Mojave Desert, California. *Journal of Vertebrate Paleontology, Abstracts of Papers*, 21(3): 93A.
- Richter, C.F. 1947. The Manix (California) earthquake of April 10, 1947. *Seismological Society of America Bulletin*, 37, 171-179, 1949.
- Ross, T.M. 1995. North-south directed extension and timing of extension and vertical-axis rotation in the southwestern Cady Mountains, Mojave desert, California. *Geological Society of America Bulletin* 107:793-811.
- Ross, T.M., B.P. Luyendyk, and R.B. Haston. 1989. Paleomagnetic evidence for Neogene tectonic rotations in the central Mojave Desert, California. *Geology*, 17:470-473.
- Sadler, P.M. 1982. Provenance and structure of late Cenozoic sediments in the northeast San Bernardino Mountains, pp. 83-91 in Cooper, J. (compiler), *Geologic Excursions in the Transverse Ranges, Volume and Guidebook 78th Annual Meeting, Cordilleran Section, Geological Society of America, Anaheim*.
- Schultz, C.B., and C.H. Falkenbach. 1947. Merychyinae, a subfamily of oreodonts. *American Museum of Natural History Bulletin*, 88:157-256.
- _____. 1949. Promerycochoerinae, a new subfamily of oreodonts. *American Museum of Natural History Bulletin*, 93:69-168.
- Sheppard, R.A., and A.J. Gude, 3rd. 1969. Diagenesis of tuffs in the Barstow Formation, Mud Hills, San Bernardino County, California. U.S. Geological Survey Professional Paper 634:1-34.
- Smith, G.I. 1964. Geology and volcanic petrology of the Lava Mountains, San Bernardino County, California. U.S. Geological Survey Professional Paper 457:97p.
- _____. 1991. Anomalous folds associated with the east-central part of the Garlock fault, southeast California. *Geological Society of America Bulletin* 103:615-624.
- Spencer, J.E. 1990a. Geologic Map of southern Avawatz Mountains, northeastern Mojave Desert region, San Bernardino County, California. U.S. Geological Survey Miscellaneous Field Studies Map MF-2117, 1:24,000.
- _____. 1990b. Late Cenozoic extensional and compressional tectonism in the southern and western Avawatz Mountains, southeastern California, pp. 317-333, in *Geological Society of America Memoir* 176.
- Steinen, R.P. 1966. Stratigraphy of the middle and upper Miocene Barstow Formation, San Bernardino County, California. University of California, Riverside, Department of Earth Sciences (Master's thesis).
- Stewart, J.H. 1983. Extensional tectonic in the Death Valley area, California, Transport of the Panamint Range structural block 80 km northwest. *Geology* 11:153-157.
- Stirton, R.A. 1930. A new genus of Soricidae from the Barstow Miocene of California. *University of California Publications in Geological Sciences* 19:217-228.
- Stock, C. 1937. A peccary skull from the Barstow Miocene, California. *Proceedings National Academy of Sciences* 23:398-404.
- Tedford, R.H., and R.M. Alf. 1962. A new Megahippus from the Barstow Formation, San Bernardino County, California. *Southern California Academy of Sciences* 61:113-122.
- Tedford, R.H., M.F. Skinner, R.W. Fields, J.M. Rensberger, D.P. Whistler, T. Galusha, B.E. Taylor, J.R. Macdonald, and S.D. Webb. 1987. Faunal succession and biochronology of the Arikarean through Hemphillian interval (late Oligocene through earliest Pliocene epochs) in North America, pp. 153-210, in Woodburne, M.O., ed, *Cenozoic Mammals of North America, geochronology and biostratigraphy*. University of California Press, Berkeley.
- Tedford, R.H., L.B. Albright, A.D. Barnosky, I. Ferrusquia-Villafranca, R.M. Hunt, Jr., J.E. Storer, C.C. Swisher, III, M.R. Voorhies, S.D. Webb, and D.P. Whistler. 2004. Mammalian biochronology of the Arikarean through Hemphillian interval (late Oligocene through early Pliocene epochs), pp. 169-231, in Woodburne, M.O., ed., *Late Cretaceous and Cenozoic Mammals of North America: Biostratigraphy and Biochronology*. Columbia University Press, New York.
- Terres, R., and B.P. Luyendyk. 1985. Neogene tectonic rotation of the San Gabriel region, California, suggested by paleomagnetic vectors. *Journal of Geophysical Research* 90, B14, 12,467-12,484.
- Valentine, M.J., L. Brown, and M. Golombek. 1988. Counterclockwise Cenozoic rotations of the Mojave Desert. *Geological Society of America Abstracts with Programs, Cordilleran Section*, 20, 239.
- Walker, J.D., J.M. Bartley, and A.F. Glazner, 1990. Large-magnitude Miocene extension in the central Mojave Desert: implications for Paleozoic to Tertiary paleogeography and tectonics. *Journal of Geophysical Research* 95:557-569.
- Weldon, Ray. 1985. Implications of the age and distribution of the late Cenozoic stratigraphy in Cajon Pass, southern California. *Cajon Pass to Manix Lake; geological investigations along Interstate 15*. San Bernardino County Museum Special Publication: 59-68.
- Wells, R.E., and J.W. Hillhouse. 1989. Paleomagnetism and tectonic rotation of the lower Miocene Peach Spring Tuff: Colorado Plateau, Arizona to Barstow, California. *Geological Society of America Bulletin* 101: 846-863.
- Whistler, D.P. 1984. An early Hemingfordian (early Miocene) fossil vertebrate fauna from Boron, western Mojave Desert, California. *Natural History Museum (Los Angeles) Contributions in Science* 355:1-36
- _____. 1991. Geologic History of the El Paso Mountains Region, pp.108-113, in Woodburne, M.O., Reynolds, R.E., and Whistler, D.P., eds., *Inland Southern California: the last 70 million years*. San Bernardino County Museum Quarterly 38(3&4).
- Whistler, D.P., and D.W. Burbank. 1992. Miocene biostratigraphy and biochronology of the Dove Spring Formation, Mojave Desert, California and a characterization of the Clarendonian age (late Miocene) in California. *Geological Society of America Bulletin* 104: 644-658.
- Whistler, D.P., R.H. Tedford, G.T. Takeuchi, X. Wang, Z.J. Tseng, and M. Perkins. 2009. Revised Miocene biostratigraphy and biochronology of the Doved Spring Formation, Mojave Desert, California, in Albright, L.B., III, ed., *Papers on geology, vertebrate paleontology, and biostratigraphy in honor of Michael O. Woodburne*. *Museum of Northern Arizona Bulletin* 65: 331-362.
- Williamson, D.A. 1980. The geology of a portion of the eastern Cady Mountains, Mojave Desert California. University of California, Riverside, Department of Earth Sciences (Master's thesis).

- Wood, A.E. 1936. Fossil heteromyid rodents in the collections of the University of California. *American Journal of Science* 32:112-119.
- Wood, H.E., II, R.W. Chaney, J. Clark, E.H. Colbert, G.L. Jepsen, J.B. Reeside, Jr., and C. Stock. 1941. Nomenclature and correlation of the North American continental Tertiary. *Geological Society of America Bulletin* 52:1-48.
- Woodburne, M.O. 1991. The Mojave Desert Province, in. Woodburne, M.O., Reynolds, R.E., and Whistler, D.P., eds, *Inland Southern California: the last 70 million years*. San Bernardino County Museum Quarterly 38(3&4): 60-77.
- _____. 1996. Precision and resolution in mammalian chronostratigraphy: principles, practices, examples. *Journal of Vertebrate Paleontology* 16(3):531-555.
- _____. 1998. Arikareean and Hemingfordian faunas of the Cady Mountains, Mojave Desert province, California. In. Terry, D.O., Jr., LaGarry, H.E., and Hunt, R.M., Jr, eds, *Depositional environments, lithostratigraphy, and biostratigraphy of the White River and Arikaree Gouups (late Eocene to early Miocene, North America)*. Geological Society of America Special Paper 325: 197-201.
- _____ and R.H. Tedford. 1982. Litho and biostratigraphy of the Barstow Formation, Mojave Desert, California, p. 65-76, in J.D. Cooper (compiler), *Volume and guidebook, "Geologic Excursions in the California Desert," 78th Annual Meeting, Cordilleran Section, Geological Society of America:65-76.*
- _____, R.H. Tedford and S.T. Miller. 1982. Stratigraphy and Geochronology of Miocene strata in the central Mojave Desert, California, p. 47-64, in J.D. Cooper (compiler), *Volume and guidebook, "Geologic Excursions in the California Desert," 78th Annual Meeting, Cordilleran Section, Geological Society of America:47-64.*
- _____, R.H. Tedford, M.S. Stevens, and B.E. Taylor. 1974. Early Miocene Mammalian faunas, Mojave Desert, California, *Journal of Paleontology*, 48:6-26.
- _____, R. H. Tedford, and C. C. Swisher, III. 1990. Lithostratigraphy, Biostratigraphy and Geochronology of the Barstow Formation Mojave Desert, southern California. Geological Society of America, 102: 459-477.
- _____ and Whistler, D.P. 1991. The Tecopa lake beds, San Bernardino Museum Association Quarterly, SP91:155-157.

Extending the boundaries of the Barstow Formation in the central Mojave Desert

R. E. Reynolds¹, D. M. Miller², M. O. Woodburne³, and L. B. Albright⁴

¹Redlands, CA. RReynolds220@verizon.net

²US Geological Survey, 345 Middlefield Road, Menlo Park CA 94025

³Professor Emeritus University of California, Riverside,

⁴University of North Florida, Jacksonville, Fl. 32240

Abstract

A sequence of marker beds is recognized in the lower Barstow Formation in the central Mojave Desert. The consistent lacustrine sequence includes (upward) beds of stromatolitic limestone, brown platy limestones, and a horizon rich in strontium sulfate and borate minerals. Where it has been dated, the marker bed sequence represents approximately 0.8 million years of deposition. The sequence can be followed from outcrops in the Harvard Hill area, east of Barstow, westerly through Daggett Ridge, Stoddard Valley, Calico Mountains, Lenwood, and into the type section of the Barstow Formation at the Mud Hills. Recent identification of the Peach Spring Tuff (PST, 18.7 Ma) in several stratigraphic sections relative to the marker sequence shows that the time of deposition of the marker bed sequence varies considerably, generally younging from east to west. These data allow preliminary interpretation of the extent and time of deposition within the early Miocene Barstow basin and the age of the fossil faunas from the Barstow Formation.

Background

Vertebrate fossils from Miocene sediments in the central Mojave Desert were reported in the early 20th century (Baker, 1911) and initially described by Merriam (1919). This assemblage of fossils was used to define the Barstovian Land Mammal Age (Wood and others, 1941). The type section for the Barstow Formation was described in the Mud Hills, east of Owl Canyon and west of Copper City Road in Solomon's Canyon (Durrell, 1953). Substantial mapping of members of the Barstow Formation was completed in the 1960s (Steinen, 1966; Dibblee, 1968). Near the end of the 20th century, the biostratigraphy and geochronology was summarized (Woodburne and others, 1990), which led to recognition that the type section was early and middle Miocene in age. The "undeformed" upper part of the Miocene Barstow Formation was suggested as a temporal and stratigraphic cap constraining the end of crustal extension in the central Mojave Desert (Dokka & Glazner, 1982; Dokka and Travis, 1990; Dokka and others, 1991). In contrast, Fillmore and Walker (1996) and Glazner and others (2002) have stated that the entire Barstow Formation represents post-extension deposition. In the

present chapter we indicate that the early Miocene part of the Barstow succession is largely syn-extensional, with the units younger than the PST deposited during waning extension in coalescing basins that had shallow marginal relief. The magnetostratigraphy of the section and its relationship to global and continental events was described by MacFadden and others (1990) and Woodburne and Swisher (1995). Small mammal assemblages in the dated stratigraphic framework of the Barstow Formation (Lindsay, 1972) have assisted in chronological analysis of other Miocene deposits (Reynolds, 1991a, 1991b; Lindsay and Reynolds, 2008; Reynolds and others 2008; Burbank and Barnosky, 1990; Reynolds and Woodburne, 2001a, b). Additional small mammal faunas have been recovered from the lower portion of the section in the Mud Hills, constrained with magnetostratigraphy (Lindsay, 1991, 1995), and described within a biochronologic framework (Tedford and others, 2004). The dated stratigraphy of the Barstow Formation has also allowed fossil trackways to be grouped by age (Alf, 1966; Sarjeant and Reynolds, 1999; Reynolds and Woodburne, 2001).

Recently, a consistent sequence of marker beds has been recognized in the lower Barstow Formation in the

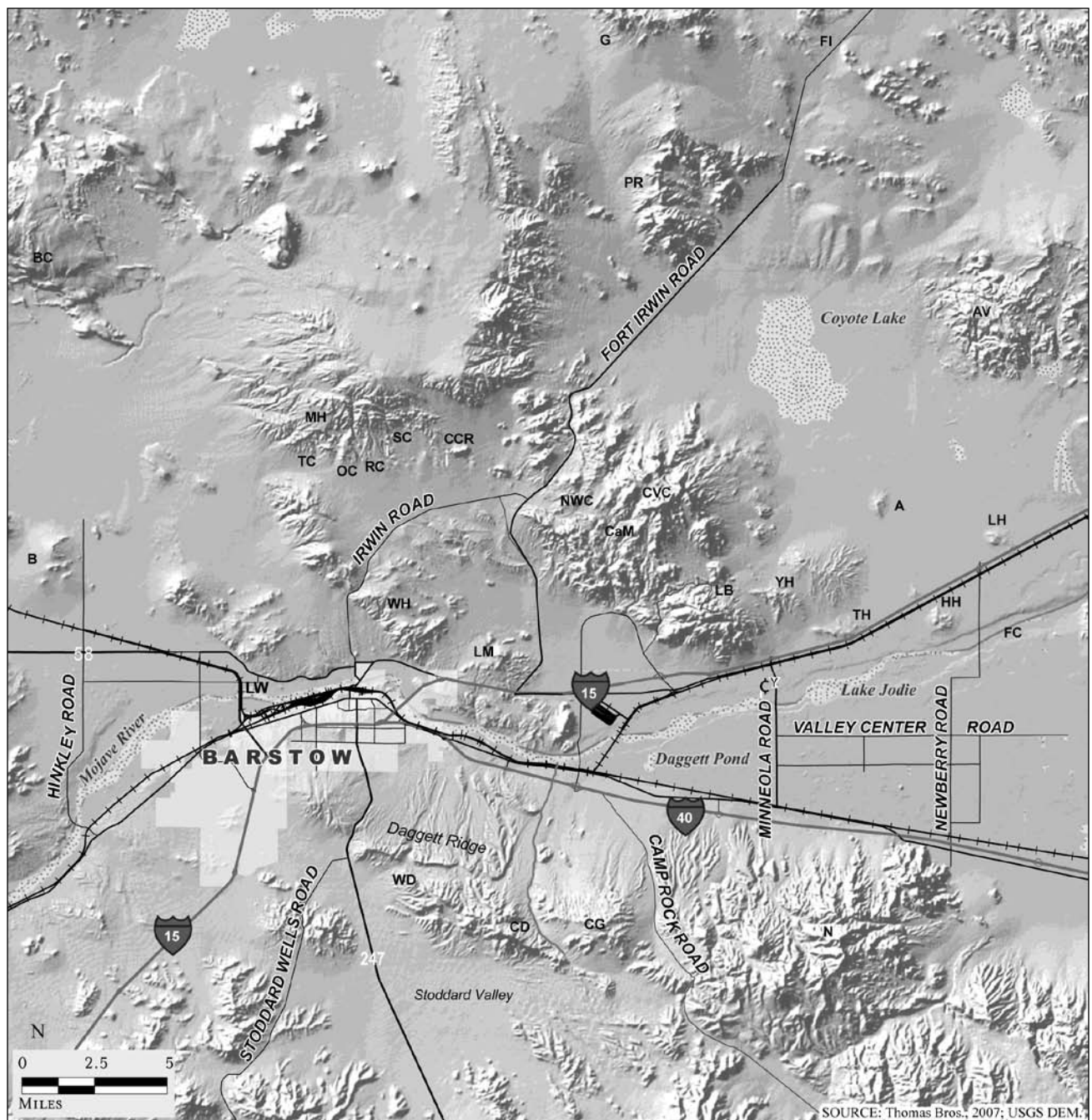


Figure 1a. Locality and outcrop reference map.

central Mojave Desert (Reynolds, 2000, 2001, 2003a, 2003b, 2004; Reynolds and Woodburne, 2002a). The appearance of each marker bed facies changes laterally according to depositional conditions, but the consistent stratigraphic order of the sequence is recognized from west to east from Lenwood to Ludlow, and from south to north from Stoddard Valley through the Mud Hills, and Calico Mountains to the hills west of Main Base Fort Irwin (Figure 1). Marker beds in the Barstow Formation include, from lowest: massive stromatolitic limestone (MSL); brown-platy limestone (BPL), and the strontium–borate horizon (SrB). Locally, other distinctive fa-

cies help correlate between outcrops. Where dated on the south limb of the Barstow syncline in the Mud Hills, the marker bed sequence spans approximately 0.8 Ma (Figure 2). In outcrops beyond the Mud Hills and Calico Mountains, the 18.7 Ma Peach Spring Tuff (informally called the “pink tuff”) is recognized at several stratigraphic positions in relation to the marker sequence, and reinforces the time-transgressive nature of the marker facies in the developing basin. The corpus of this article is focused on describing the stratigraphic and geographic occurrence of this marker bed sequence, and its relation to a sequence of conglomeratic units in order to gain a better understand-

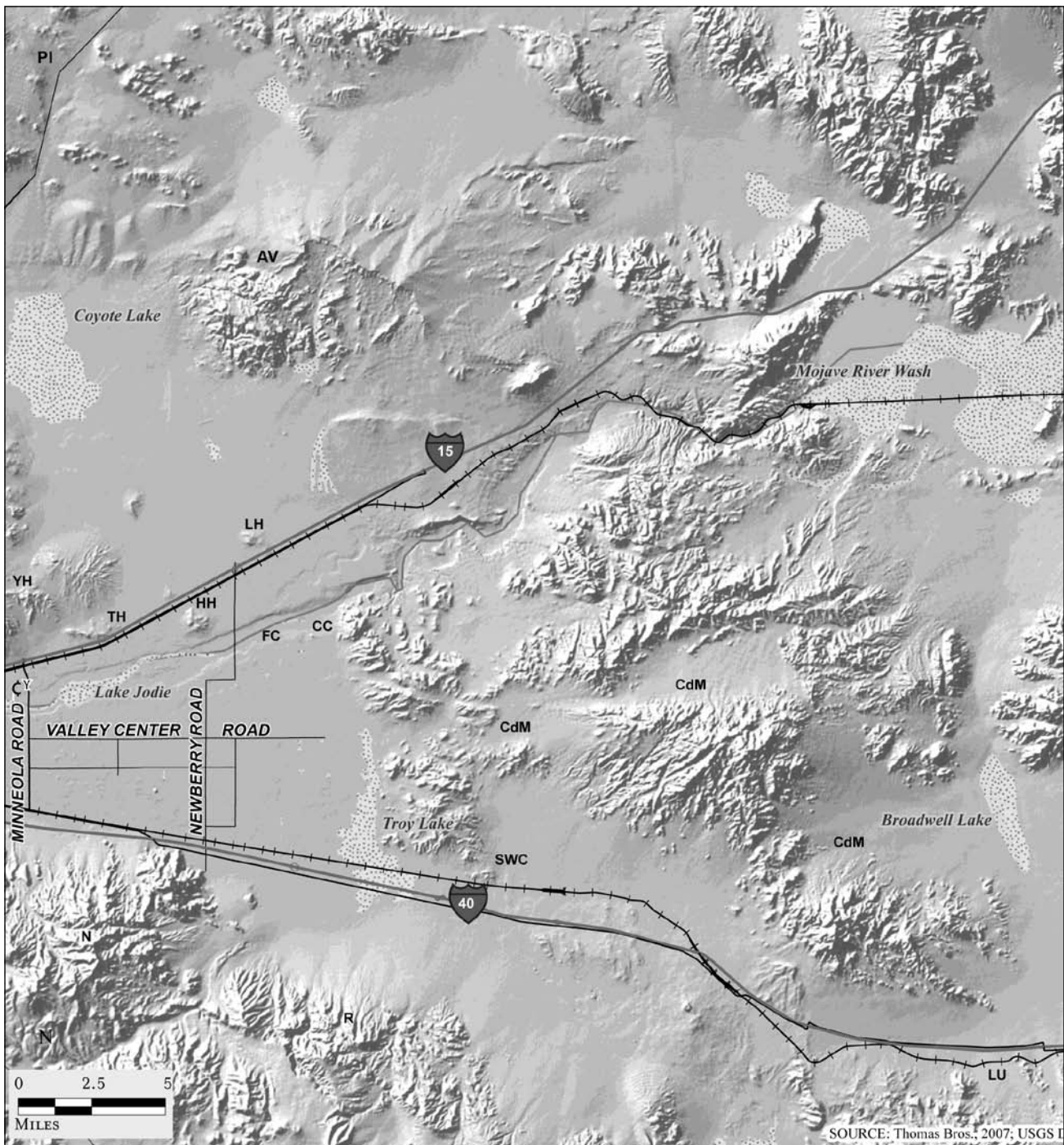


Figure 1b. Locality and outcrop reference map.

ing of the evolution of the Barstow Basin during the early Miocene.

Marker beds of the sequence

Massive Stromatolitic Limestone (MSL)

The massive stromatolitic limestone is named for its massive appearance in the eastern portion of the early Barstow basin, and for the stromatolite structures (tufa deposited around plants; phytoherms, Cole and others,

2005; onchoids of Awramik and others, 2000), in the western part of the basin. Stromatolites and massive limestone with organic and stromatolitic structures appear in the central part of the basin. The MSL occurs below the brown platy limestone (BPL). It can be found in western outcrops at Lenwood (Grandview), Mud Hills, west and east Calico Mountains, Main Base Fort Irwin, west, central and east Daggett Ridge, Stoddard Valley, Yermo, Toomey, Agate, Harvard, Lime and Camp Cady hills, and in the southwestern Cady Mountains and at the DuPont

strontium prospects (Durrell, 1953) in the southeastern Cady Mountains northwest of Ludlow (Figures 1 and 3).

In the northwestern Calico Mountains, and southwestern Cady Mountains and DuPont claims in the southeastern Cady Mountains (Figure 1), the massive MSL is 3 to 4 four feet thick. At American Borate prospects on the north side of Lead Mountain, the “basal limestone” (=MSL of Dibblee, 1970) is 70 feet thick. The color is light tan to dark gray, and the surface is eroded by chemical weathering so that silicified organic structures stand out. The massive limestone often looks like gray, weathered, cherty Paleozoic limestone, a useful field characteristic. In other places, it may contain black and white thin beds of silicified limestone, or even white, platy to fissile limestone.

Stromatolite columns occur as structures in, and at the same horizon as, the massive limestone. The stromatolite columns are simple if tufa is deposited around water reeds, or complex if tufa is deposited around branching bushes. Simple stromatolites are up to five inches in diameter in the Mud Hills and Calico Mountains, and approach 12 inches in diameter in northern Stoddard Valley and Agate Hill.

Stromatolite formation is similar to that described for onchoids in Pleistocene Lake Manix (Awramik and others, 2000; Reheis and others, 2007). For cyanobacteria to precipitate carbonate, lake water must be fresh and clear enough (shallow, low sediment content) to permit sunlight to reach bottom-dwelling or wall-dwelling cyanobacteria. The cyanobacteria react with groundwater containing concentrated calcium bicarbonate to precipitate calcium carbonate on objects such as the upper surface of pebbles. Lacustrine limestone is precipitated when lake

water is calcium-rich. In the central to eastern portion of the Barstow lake, thick siliceous lacustrine has been deposited, suggesting calcium-rich lake water that was a host for later silicification. In the western and younger portion of the Miocene lake, carbonate was deposited on plants instead of on cobbles, suggesting low-carbonate water reacting with high calcium bicarbonate from the fine or coarse substrate. Tubular stromatolites from the south limb of the Barstow Syncline of the Mud Hills are approximately 96 weight percent calcium carbonate.

The simple and branching stromatolites of the Mud

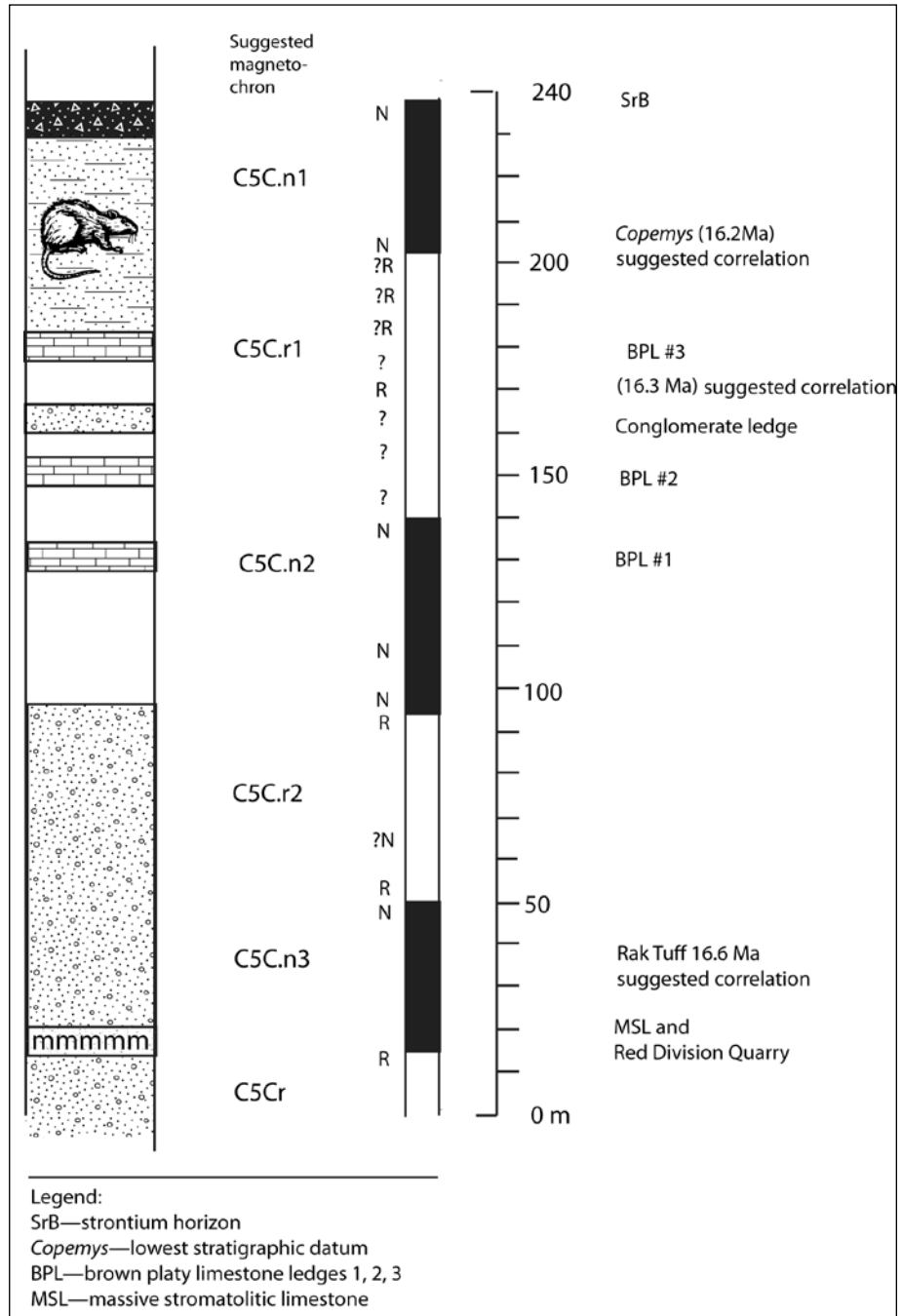


Figure 2 Magnetic polarity profile, Trident Canyon section, Barstow Formation, Mud Hills.

Hills are located in a thin bed of gray silty lacustrine sand that interfingers with the Owl Conglomerate. Therefore, simple tubular or complex tubular stromatolites within conglomerates may indicate deposition during a shallow lake incursion that crossed alluvial fans. In contrast, thick limestones at Harvard Hill were probably deposited in perennial shallow water that precluded growth of reeds and branching plants, but maintained enough clarity that cyanobacteria were actively precipitating thick layers of calcium carbonate.

Tubular stromatolites in silty sands mark the level of Red Division Quarry in Trident Canyon, where mammal fossils are exposed along the south limb of the syncline. The magnetostratigraphic section from Trident Canyon indicates that the age of Red Division Quarry is about 16.8 Ma (Figure 2).

In addition to data from volcanic events, the age and distribution of stromatolitic units assist in reconstructing the depositional history of the lower part of the Barstow Formation in the Barstow Basin which is otherwise is not well calibrated and contains few fossil vertebrates. Cole and others (2005) used U-Pb methods to date branching tufa mounds, which they called phytoherms, in Owl Canyon (Mud Hills). Results from three dated MSL localities are 15.39 ± 0.15 Ma and 15.30 ± 0.25 Ma at Owl Canyon Campground (Fig. 1), and 16.25 ± 0.25 Ma one mile north northeast of the campground in Owl Canyon on the north limb of the Barstow Syncline. Although the dates may be less precise than magnetostratigraphic ages, they give the relative age differences indicating an 0.5 to 1.0 Ma difference in age for the MSL in the two locations in the Mud Hills.

Dates on tubular stromatolites and phytoherms of the MSL from Trident Canyon in the Mud Hills are 16.8 Ma (Fig. 2). The stromatolites dates of the MSL from the Mud Hills (15.3–16.8 Ma) can be compared to a few other locations. At Harvard Hill the MSL overlies the PST, so is about 18.7 to 18.5 Ma (Leslie and others, this volume). At the Columbus/Gem Mine and DuPont Claims (Figures 1 and 3), the MSL is stratigraphically far below the 18.7 Ma PST. These dates suggest that across the Barstow basin the MSL formed as much as 3 million years earlier in the south and east, as compared to the Mud Hills. Additional stratigraphic observations on the position of the tufa (=MSL, RER pers. obs.) in the Mud Hills indicate that the stromatolitic structures occur far below the top of the Owl Conglomerate in many places (Loop Entrance, Rainy Day Canyon, Trident Canyon). Although not directly dated, these occurrences are consistent with the 19.3 Ma initial deposition of the Barstow Formation indicated by the Red Tuff in the Rainbow Loop Entrance area (Woodburne et al., 1990). The MSL

units thus provide empirical evidence for lacustrine sedimentation coeval with the development of distal alluvial fans as generally reconstructed by Woodburne et al. (1990), and as documented for Pleistocene Lake Manix by Reheis and Miller (this volume).

Brown-platy limestone (BPL)

Brown-platy limestone (BPL) is named because it is brown in color and generally occurs as flat sheets or plates. It commonly has a crinkly bedding surface and contains rip-up clasts and fossils of water reed and grasses are preserved at the base. In many sections (southern Mud Hills, Daggett Ridge; Fig. 2) it appears as resistant, ledge forming beds that are approximately 30 cm thick. The near shore outcrop containing dacite pebbles in the northwestern Calico Mountains (Fig. 1) is 4 m thick. A single BPL in central basin facies on the north side of Lead Mountain (Fig. 1) is 4 m thick and contains no clastic material. On the south side of the Mud Hills, at Grandview, in Central Daggett Ridge, in the central and eastern Calico Mountains, in Stoddard Valley, and in the southwestern Cady Mountains, brown platy limestone was deposited in siltstone (Figure 1, 2, 3). In the northwestern Calico Mountains and at Lead Mountain, the BPL was deposited over volcanic breccias, and below shale beds. At Columbus/Gem on eastern Daggett Ridge (Fig 1, 3), BPL beds interfinger with volcanoclastic (Pickhandle-like) sediments and are overlain by silts. Three resistant ledge-forming layers of limestone are often apparent (southern Mud Hills, northeastern Mud Hills, Lenwood, Central Calico Mountains, Central Daggett Ridge, Columbus/Gem), although in the Mud Hills within the Rainbow Loop road, there are at least five layers.

Toward basin center (Lead Mountain) there is a thick section of platy, brown limestone above MSL and below Strontium-Borate horizon (described below). The calcite content for BPLs at three localities was determined by weighing, dissolving in acid, and weighing the insoluble fraction. The BPL at Lead Mountain is mottled brown and tan, dense, porcelaneous, and without solution cavities. It contains 99% CaCO_3 by weight. The BPL in the southern Mud Hills has minor pyrolusite stains on alternating brown and tan layers of limestone with minor solution cavities. It contains 95.7% CaCO_3 by weight. The BPL in the northwestern Calico Mountains grades upward from a conglomerate containing sub-rounded dacite clasts (2+ cm) to a “spongy” textured carbonate. This upper section contains vertical solution cavities reminiscent of those in the tufa at Travertine Point in Imperial County, California. The vertical cavities are lined with pyrolusite? and clear hyaline opal. The BPL specimens from low in the section at the northwestern Calico

Mountains site contains 28% CaCO₃ by weight due to the high percentage of included dacite clasts.

The BPL hosts light metals (Sb, As, Ag, Hg) deposited by sub-aqueous hot springs in the vicinity of the Sulfur Hole (Fig. 1; Cooper, and others 2002; Schuiling, 1999) in the eastern Calico Mountains. Pyritic hot spring deposits at the Sulfur Hole relate to ongoing activity in the Calico volcanic center.

Layers of BPL, whether separate or combined, are in places silicified at the base by brown, red or yellow jasper. The silicification is a post-depositional and post structural event. Jasper deposition is always greater in the topographically lowest BPL bed. This can be seen in sections that have normal stratigraphic orientation (Trident and Little Borate canyons, Grandview) and in those sections that have been overturned (Columbus/Gem, DuPont).

Insect-bearing, small discoidal stromatolites are found in several horizons between the BPL and the overlying strontium-borate horizon, and are described below under Paleontology. The BPL and associated siltstones have produced avian eggshells, including those of flamingos, which help describe a lake margin mud flat (Miller, 1952, 1956; Reynolds, 1998; Leggitt, 2002).

In summary, massive limestone (MSL) precipitated in shallow saline waters of the central basin while tubular stromatolites were deposited under conditions of rising shallow lakes inundating plants. BPL was deposited as biohermal mats that coarsen toward the shore and thin toward basin center. Both the MSL and BPL developed at lake margins under transgressive conditions in the early stages of basin formation before lake salinity had stabilized.

Strontium–Borate horizon (SrB)

The Strontium-Borate horizon (SrB) is named for the layers of borate and strontium minerals that have been mined from sediments of the Barstow Formation in the Mud Hills, Lead Mountain, Calico Mountains, Daggett Ridge, the DuPont Claims at the southeastern edge of the Cady Mountains northwest of Ludlow and that occur at Agate Hill (Fig. 1, 3) and in hills west of Main Base, Fort Irwin (Fig. 1) (Byers, 1960; see saline minerals p. 65, and stratigraphic section p. 41; Dibblee, 1968, 1970; Durrell, 1953; Weber, 1966, 1967; Link, 1980). The SrB horizon occurs stratigraphically higher than the BPL layers (Figures 2, 3). The minerals are found as nodules or geodes in limestone layers, and economic quantities were mined from lenticular pods (up to 3m thick) consisting of colemanite. Colemanite, howlite, celestine, calcite and quartz occur in cavities within petroliferous, siliceous lacustrine limestone. The distribution of borate minerals appears to be restricted to lacustrine limestone and lacustrine silt-

stone of the central and eastern Calico Mountains, north Lead Mountain, central and eastern (Columbus/Gem) Daggett Ridge and at northern Agate Hill. Original borate minerals in the lacustrine deposits were probably sodium and sodium/calcium borates (borax and ulexite). These soluble borates have been altered by ground water to the more stable calcium borate (colemanite) and calcium borosilicate (howlite; Housley, p. c. to RER), the calcium ion being obtained during ground water etching of the lacustrine limestone.

The Strontium-Borate horizon contains sulfate and carbonates of strontium, in addition to borate minerals. Celestine (strontium sulfate) is often associated with the borate minerals, but is not confined to borate localities, and appears to have a much wider distribution. It is found on the south flank of the Mud Hills at Loop Exit and Ross deposits as layers of strontianite (strontium carbonate, Durrell, 1953) concretions and on the north flank at the Solomon deposits as strontianite rock (Durrell, 1953), and as far east as the DuPont (celestine) deposits at Sleeping Beauty (Durrell, 1953). Strontium carbonate (strontianite) has also been documented at the Columbus/Gem mine (Housley, p. c. to RER). East of Calico Ghost Town, fibrous celestine seams cut across bedding planes of siltstone, and may be replacing gypsum, var. satin spar. Crystalline varieties of gypsum (hydrous calcium sulfate) occur in association with sulfate and borate minerals.

Deposition of the strontium/borate deposits was apparently not synchronous, and began prior to 18.7Ma where the SrB sits below the PST at Columbus/Gem, and DuPont claims. Mineral deposition occurred at approximately 16.2 Ma in the Mud Hills (Figure 2, 3) a span of 2.3 Ma.

Deposition of each marker facies is dependent on a different set of conditions (MSL—lake transgression across vegetated fanglomerates; BPL—shallow water with biohermal mats; SrB—hot spring minerals), and repetition of the sequence (MSL–BPL–SrB) is consistent from place to place. Since each of the marker bed units was deposited by individually different lacustrine conditions, it is difficult to propose a model that would repeat this specific sequence of depositional conditions at different times in multiple isolated basins.

Depositional setting

Dibblee (1968) and Steinen (1966) mapped and described the Barstow Formation, focused on outcrops in the Mud Hills and adjacent areas. As summarized in Woodburne and others (1990), the formation is composed largely of about 1,000 m of lacustrine and fine-clastic sediments that crop out in this district, with numerous

interbedded air-fall tuffs, and associated fossil mammals. On at least the southern, western, and northern margins of the Barstow Basin, the Barstow Formation includes thick successions of interbedded marginal alluvial conglomerate beds that are especially well developed and exposed in the lower part of the formation; in the Mud Hills, this unit is designated as the Owl Conglomerate Member. Whereas the above-cited papers focused mainly on the Mud Hills, we present below information designed to provide additional details of this marginal facies in areas not previously discussed in detail, and which contain information directly pertinent to the age and depositional setting of this member.

The marker units discussed above lie upon, or are sometimes interbedded with, beds of conglomeratic alluvium, which tend to lie at the base of the section. Dokka and Glazner (1982) refer to Miocene strata in Daggett Ridge as Barstow Formation. As noted above, these outcrops are physically well removed from the districts that typically contain this formation (see Fig. 1) and we hesitate to refer to conglomeratic units in this region as the Owl Conglomerate Member, pending further analysis. Still, the Columbus/Gem mine on eastern Daggett Ridge displays a sequence of in part, overturned to the north, but otherwise north-dipping conglomeratic beds that is, in part, overlain by the marker package discussed above, and in part interbedded with the MSL. This unit of granitic conglomerate and mudflow deposits is about 330 m thick, and is overlain as well by a 150 m succession of silt, sandstone, BPL and SrB that is capped by the 18.7 Ma Peach Spring Tuff (PST). The entire 0.8 m.y.-long marker package is thus constrained in age by the overlying PST (18.7 Ma). Therefore, the lower limits of the conglomerate must be greater than 19.5 Ma old.

In the southern part of the Mud Hills in the vicinity of Red Division Quarry, in Trident Canyon (TC, Fig. 1A), the marker package is within and above the Owl Conglomerate (Woodburne et al., 1990, Fig. 4; Steep-side Quarry area). A magnetostratigraphic section that includes the MSL within the Owl Conglomerate suggests that the MSL is about 16.8 Ma old (Figure 2). The Owl Conglomerate also includes the Rak Tuff (16.3 Ma) and the Red Tuff (19.3 ± 0.2 Ma, Woodburne, 1996). Since the 19.3 Ma date for the Red Tuff is at the base of the Owl Conglomerate, and a projected date of >19.5 Ma is proposed for the base of the Columbus/Gem conglomerate, the latter is considered to pre-date the Owl Conglomerate Member exposed in the southern part of the Mud Hills. Coarse breccias of similar age at Harvard Hill are recognized below a 19.1 Ma tuff (Leslie and others, 2010). These early conglomerates may, in part, represent debris from the basin margin during uplift responsible for the

initial closing of the Barstow Basin. It appears likely that the deposition of the conglomeratic units in the Barstow Basin, whether or not included within the Barstow Formation, began at least by about 19.5 Ma and was partially coeval with the evolution of the Mojave Extensional Belt (Dokka, 1989; Dokka and Travis, 1990; Fillmore and Walker, 1996).

Conglomerate with Rounded Clasts (CRC)

This distinctive facies is not in a consistent position with respect to the marker bed sequence. Sub-rounded to well-rounded, large pebble to cobble clasts in well-defined, thin, clast-packed beds are interbedded within siltstone and sandstone in sections of the Barstow Formation. These CRC horizons occur at Lenwood, Wall Street Canyon in the Calico Mountains, Toomey/Yermo Hills, Agate Hill and Lime Hill (Fig. 1). Clast rounding suggests an origin as lag gravels from a mature pre-Miocene erosional surface reworked into the Miocene depocenter. The degree of rounding distinguishes the CRC clasts from the angular to sub-angular clasts of conglomerates of eastern Daggett Ridge (Columbus/Gem) and from the Owl Conglomerate and Gravel Hills conglomerate. These conglomerates with rounded clasts all appear to come from sources on the north side of the developing basin, and may represent two different depositional environments. Those at Wall Street Canyon and at the Yermo/Toomey Hills are thin, cross stratified and clast-supported, and the cobbles were perhaps collected by sheet floods and deposited in broad streams. Conglomerates at Lenwood, Agate Hill and Lime Hill have large, matrix supported clasts, and are thick bedded to massive, and may represent debris flows (facies descriptions by Fillmore and Walker, 1996). At the Yermo Hills, these CRC beds are between 16.5 Ma and 15 Ma based on an upper dated tuff (Miller and others, this volume) and biostratigraphic information (see below). Since these conglomerates are stratigraphically higher than the coarse Owl Conglomerate and Daggett Ridge Conglomerate, their presence in these lacustrine siltstones may represent continued uplift on the north side of the basin.

Wall Street Canyon

Well-rounded cobbles include felsic granite, aplite, dark quartzite, marble, and green-red calcsilicate rock in the Barstow Formation at Calico Ghost Town, north of the Calico Fault. These lithologies suggest a northerly source near Goldstone or the Paradise Range. The cobbles occur in a unit of red and green calcareous conglomeratic sandstones between brown platy limestones and the borate interval (P.M. Sadler, p. c. 2010).

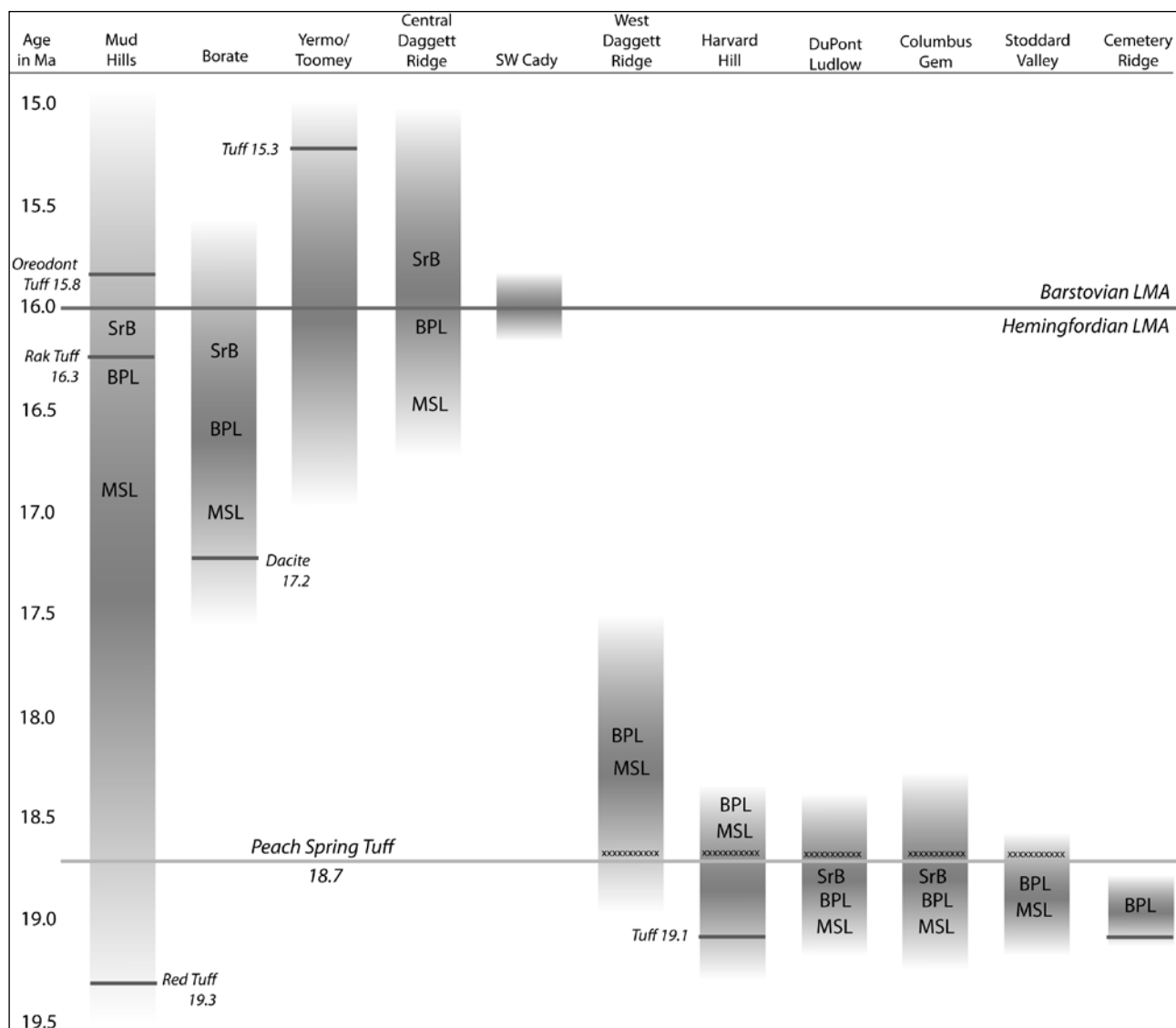


Figure 3. Stratigraphic sections related to the He/Ba transition and the Peach Spring Tuff.

Toomey Hills–Yermo Hills (Emerald Basin)

Arkosic sandstone in siltstones of the Barstow Formation in the Yermo/Toomey Hills contains well-rounded pebbles of mafic granodiorite, gabbro granitic gneiss, granite, andesite, rhyolite, and quartzite. Less common are lineated granite mylonite, red-green calcsilicate rock, and phyllite. This resembles a suite of cobbles similar to that described (Byers, 1960) from the Paradise and Goldstone ranges to the north. Two green-colored (celadonite alteration?) pebble beds are present, separated by 1–5 m of yellow arkosic sand. Cross-bedding is prominent, inconsistently oriented, and indicates a fluvial origin. Columnar tufa around casts of fibrous vegetation is upright in the lower conglomerate. The CRC in the Yermo Hills is below a 15.3 Ma white tuff (Miller and others, this vol.) and, in the Toomey Hills, is associated with Hemingfordian NALMA (>16 Ma) vertebrate fossils. Although no other

marker units are present in the Yermo/Toomey section, a 3 m thick biogenic limestone that overlies the white tuff can be mapped westward to a point where it overlies layers of BPL in siltstone of the Barstow Formation in the eastern Calico Mountains (P.M. Sadler, p. c. 2010).

Grandview at Lenwood

CRC clasts in Barstow Formation outcrops at Grandview consist of Proterozoic granitic rock, silicified recrystallized Paleozoic limestone, Jurassic granitic rock (some albitized), garnet schist from the Paradise Range, and rhyolite. This CRC is 210 m below a sequence of marker units that consists of MSL, BPL and an as yet undated pink, porphyritic andesite.

Agate Hill

Rounded pebbles in Barstow Formation beds of arkose include calcsilicate rock, phyllite (argillite of Byers, 1960), quartzite, Jurassic- and Cretaceous-looking gran-

ite, felsic and gabbroic dike rocks, andesite, rhyolite, basalt clasts, and shale. Silicified red conglomerate beds appear to be sedimentary breccias consisting of rounded clasts in an arkosic matrix. The rock unit is faulted against beds considered to be the Barstow Formation that contain colemanite (Byers, 1960), and perhaps of similar age to colemanite deposits in the Calico Mountains. The assemblage suggests derivation from a source similar to the modern Goldstone area.

Lime Hill South (Martha's Cafe)

Medium to dark gray silicified beds of arkose potentially pertaining to the Clews Fonglomerate (Byers, 1960) of about 22 Ma (Woodburne and Reynolds, 2010) with distinct thin parallel bedded pebble beds show long sweeping crossbeds that indicate south-directed transport. Well-rounded pebble-size clasts are common and cobbles are uncommon. Clast types are generally granitic, with minor chert, siltstone, and altered fine-grained volcanic rocks. Outcrops are overlain by Yermo gravels and separated from the Barstow Formation exposed in a hand-dug well to the east. Possible source is from the Clews Fonglomerate (Byers, 1960).

Main Base Fort Irwin

Rounded cobbles of local diorite occur low in a section containing the MSL, BPL and borates (Byers, 1960).

Peach Spring Tuff

The Peach Spring Tuff (PST; Wells and Hillhouse, 1989) crops out in the Mojave Desert region in the Cady Mountains, Daggett Ridge, and other sites. Pink tuffaceous units that crop out in the Daggett Ridge region have since been recognized as the Peach Spring Tuff, as well. The age of the PST is had been generally accepted as circa 18.5 Ma (Neilson and others, 1985). Hillhouse and others (this volume) examined remanent magnetism for tuffs at West Gem, Stoddard Cutoff, Stoddard Valley and other localities in this region, and showed them to be consistent with that of a 17.9 Ma welded tuff in the central Cady Mountains (Miller, 1980) and also with the PST known through a wide area (Glazner and others, 1986; Wells and Hillhouse, 1989). The informally-named "Shamrock Tuff" (Leslie and others, 2010, this Vol) is a thick, green tuff in the lacustrine section below the MSL at Harvard Hill. Zircon crystals in this tuff have been dated at ~18.7 Ma (Leslie and others, this vol); the Shamrock Tuff is thought to represent a lacustrine facies of the PST.

The PST is an 18.7 (Hillhouse, and others this volume) Ma near-instantaneous depositional event providing temporal constraint where it occurs within the marker sequence in the Barstow Formation. On eastern Daggett Ridge at Columbus /Gem the marker section

(MSL, BPL, SrB) is below the PST. Ten km to the west on Daggett Ridge near the Stoddard Cutoff, the MSL and BPL are above the PST. At Harvard Hill, the MSL occurs above the Shamrock Tuff (= lacustrine PST). The variable stratigraphic position of the marker sequence in relation to the PST (Fig. 3) implies time-transgressive deposition for that sequence.

Summary of the marker sequence

The MSL and the BPL developed at lake margins under transgressive conditions. Massive limestone precipitates due to the advent of saline conditions in shallow waters of the central basin; tubular stromatolites are deposited under conditions of rising shallow lakes inundating plants. BPL is deposited as biohermal mats that coarsen toward the shore and thin toward basin center. All of these units, including the SrB are time transgressive in the sense that they are older than 18.7 Ma on eastern Daggett Ridge and younger than 16 Ma in the Mud Hills. It is not yet clear whether the marker succession was physically continuous across space and time (Fig. 3) although the succession (MSL, BPL, SrB) is remarkably consistent. In the Calico Mountains and on Daggett Ridge, the SrB is related to hot spring mineral exhalites (Schuiling, 1999), where the elements and ions are variable away from the source, e.g. the Calico volcanic center. Elsewhere (Lead Mountain, Mud Hills, DuPont near Ludlow), the strontium and borate concentrations may have precipitated from concentrated solutions in saline lake water. Therefore, the deposition of carbonate marker facies depends upon a sequence of depositional conditions relating to lake expansion, salinity, elevation of basin perimeters, and rising and falling lake waters across various substrates. In contrast, the venting of mineral-rich water from subaqueous springs (spanning 2 Ma) probably is related to the Calico Mountains volcanism over ~17 to 16 Ma (Singleton and Gans, 2008), which may be related to extensional tectonics.

Paleontology

Plant, invertebrate and vertebrate fossils occur in the Calico Mountains. The Barstow Formation in the Calico Mountains rests stratigraphically on the Pickhandle Formation which is dated at 19.0 Ma at Little Borate Canyon (Singleton and Gans, 2008). Most important to this discussion are sediments containing horse fossils which provide biostratigraphic data. Early Miocene horse biostratigraphy has been well documented in the Mud Hills (Woodburne, 1991, this vol.; Pagnac, 2009) where *Parapliohippus carrizoensis* is found in beds that range in age from about 17 to 16 Ma (Woodburne, 1998; Tedford and others, 2004), *Acritohippus stylodontus* ranges from 16–14.8 Ma, and *Scaphohippus intermontanus* is known from 14.9–

14.0 Ma (Pagnac, 2009). The Miocene horse biostratigraphy from the Barstow Formation in the Mud Hills helps clarify the stratigraphy in the Calico Mountains and that in the Yermo and Toomey Hills. The Toomey Hills contains the transition from *P. carrizoensis* to *A. styliodontus* at 16 Ma (Woodburne, 1991). The Calico Mountains contain both *P. carrizoensis* (RER, pers. obs.) in the Tbc₃ of Singleton (2004; light red-brown sandstone below the SrB Horizon) and *S. ("Merychippus") intermontanus* (Dibblee, 1970) in a fault-bounded block of Tbc_{sl} (Singleton, 2004). Thus, the Calico Mountain section of the Barstow Formation ranges in age from at least 16.0 to 14.9 Ma (Woodburne, this vol.; Pagnac, 2009). The Calico section contains the MSL, BPL and SrB marker units with inter-fingering dacite avalanche breccias that have older dates (17.1–16.9 Ma, Singleton, 2004).

The Daggett Ridge Powerline locality (SBCM 1-109-2) contains the small *P. carrizoensis* and a large horse (possibly *A. styliodontus*) suggesting the Hemingfordian/Barstovian transition at 16 Ma (Reynolds, 1991b); Woodburne, 1991). The sequence at this locality contains the MSL, BPL and SrB, and fossils are located within the upper of three brown platy limestones (Fig. 3). This suggests that BPL #3 is about 16.0–16.2 Ma at central Daggett Ridge, comparable to its position in the Mud Hills but younger than at eastern Daggett Ridge, where BPL #3 sits below the 18.7 Ma PST.

The National Lead locality in the southwestern Cady Mountains (Dibblee and Bassett, 1966a) contains *P. carrizoensis* at the base of a composite BPL section above MSL. This suggests an age of 17 to 16 Ma (Tedford and others, 2004) for the marker sequence in the southwestern Cady Mountains. The marker sequence in the southeastern Cady Mountains contains the MSL, BPL and SrB which underlie the 18.7 PST.

There is great potential to extract environmental and depositional data from plants, invertebrates and the minerals in the Barstow Formation at various outcrops. The Barstow Formation in the Mud Hills contains well-preserved wood and leaves of juniper, poison oak, two species of oak tree (*Quercus convexa*, *Quercus dayana*), palm, and water reeds (Alf, 1970) that typify a chaparral or parkland community. Palm has been located from a horizon between the MSL and BPL at Little Borate.

Many species of very well preserved silicified insects and other invertebrates have been described in the Barstow Formation of the Calico Mountains, Mud Hills, and Black Canyon. Fossil categories include ostracodes, foraminifera, pelecypods, gastropods (Taylor, 1954), crustaceans, arachnids (3 orders), and insects (14 orders, 30 species; Jenkins, 1986; See Additional Reading in references section). These fossils are preserved in three

layers of petroliferous, calcareous stromatolites (Jenkins, 1986; Spencer, 2006). Field observations (Spencer, 2006; RER pers. obs. 2008) indicate that layers of the fossiliferous stromatolites in the Black Canyon anticline occur between the BPL and a stromatolitic SrB. Fossiliferous stromatolite nodules occur in the Mud Hills and the Calico Mountains between the BPL and SrB (Reynolds, 2000, 2001; Leggitt, 2000, 2002).

Although the taxa vary between each locality, the method of preservation through replacement by colloidal silica and other minerals appears to be consistent (Park, 1995; Spencer, 2006). The similarity of the depositional environment and modes of preservation reinforce lateral continuity of a Miocene lake from Black Canyon through the Mud Hills and into the Calico Mountains (Spencer, 2006). Spencer (2006) also notes that such a correlation using fossiliferous concretions and the marker beds (Reynolds and Woodburne, 2001a) contradicts the report that lacustrine sediments of the Barstow Formation in the Calico Mountains were deposited between 19 and 17 Ma (Singleton and Gans, 2008).

Regional relationships

The south-central Cady Mountains (Woodburne, 1998) and the western, central, and eastern Cady Mountains contain early Miocene sediments and late Arikarean and Hemingfordian fossil faunas (Moseley, 1978; Miller, 1980; Williamson, 1980). Geologic mapping (Dibblee and Bassett, 1966b; Moseley, 1978; Miller, 1980; Williamson, 1980) has located the Peach Spring Tuff in these deposits, but the marker sequence reported from the Barstow Formation appears to be absent (Reynolds, pers. obs.). The Barstow Formation in the Alvord Mountains overlies late Hemingfordian and early Barstovian fauna (Byers, 1960; Woodburne, 1991, this vol.) and lies above the Spanish Canyon Formation, which contains the Peach Spring Tuff (Byers, 1960; Hillhouse and others, this vol.). The marker sequence has not been located in the Spanish Canyon Formation or the Barstow Formation of the Alvord Mountains, and therefore, the use of "Barstow Formation" is questioned. The Alvord and Cady mountains contain sediments and faunas that overlap the age of deposition of the Barstow Formation, but apparently do not contain the sequence of marker units that are characteristic of the early depositional history of the Barstow Formation. This suggests that the Alvord and Cady sedimentary sections were deposited under different conditions, or in different depositional basins than the Barstow Basin. Similar age sediments south and east of Ludlow (e.g., Miller, 1994) and west of Barstow remain to be examined for the marker beds.

Basin formation

The time-transgressive character of the marker units in the early basin, older than the PST (18.7 Ma) to the southeast, and younger than the PST in the central and western portions, suggests a basinal depocenter that expanded from the east-southeast to the west-northwest. Shifting depocenters might be expected during the tilting of listric-normal fault blocks in the upper plate of an extending terrane. This suggests that the Barstow Formation was deposited (in part) during a period that included active extensional tectonics (Singleton and Gans, 2008; compare Fillmore and Walker, 1996). Deposition ended at different times in different lobes of the basin. In Stoddard Basin, there is little deposition evident after the PST, so deposition may have ended before 18 Ma. On western Daggett Ridge, deposition after the PST event continued well past deposition of the MSL and BPL (perhaps a 0.5 Ma interval) and thickness of silcrete paleosols indicate cessation of deposition in the realm of 17.5 Ma. Deposition of the Barstow Formation in the Mud Hills portion of the basin continued past the deposition of the Lapilli Tuff (13.4 ± 0.2 Ma, Woodburne, 1996) and perhaps as late as 13 Ma.

Summary

Exploration of outcrops of Miocene sediments in the Barstow Basin away from the type section (Durrell, 1953; Woodburne, 1991; MacFadden and others, 1990) of the Barstow Formation suggests that initial deposition of the lacustrine facies in the expanded Barstow Basin can be recognized by a sequence of marker horizons. These horizons are stratigraphically below and older than the 18.7 PST in Stoddard Valley, Columbus/Gem on southeastern Daggett Ridge, and at DuPont near Ludlow; are stratigraphically above and younger than the 18.7 PST at Harvard Hill and northwestern Daggett Ridge; and began deposition between 17 and 16 Ma in the Mud Hills and Calico Mountains.

Deposition of each marker facies is dependent on a different set of conditions (MSL—lake transgression across fanglomerates; BPL—biothermal mats in shallow water; SrB—hot spring mineral exhalites), and repetition of the sequence consistently in different isolated basins seems improbable.

The marker sequence offers evidence of deposition in a basin with a depocenter that moved northwestward as the basin was filled, tilted, or blocked to the southeast. Although it can be argued that an early southeastern lake fed into a younger northwestern lake, no evidence of basin spill-over has been found. That the marker beds were deposited by multiple different lacustrine events is

improbable given the specific sequence of depositional conditions needed to produce such a marker series.

References

- Alf, R. M., 1966, Mammal Trackways from the Barstow Formation, California. *Southern Calif. Acad. Sci.* Vol. 65, No. 4, 258-264.
- Alf, R. M., 1970, A preliminary report on a Miocene flora from the Barstow Formation, Barstow, California. *Southern Calif. Acad. Sci.* Vol. 69(3, 4), p. 183-189.
- Awramik, S. M., H. P. Buchheim, Leroy Leggitt, and Kyung Sik Woo, 2000. *Oncoids of the Late Pleistocene Manix Formation, Mojave Desert region, California*. Redlands: San Bernardino County Museum Association Quarterly 47(2):25-31.
- Baker, C. L., 1911. Notes on the later Cenozoic history of the Mohave Desert region of southeastern California. *University of California Publications in Geological Sciences*, 6:333-383.
- Burbank, D.W., and Barnosky, A.D. 1990. The magnetochronology of Barstovian mammals in southwestern Montana and implications for the initiation of Neogene crustal extension in the northern Rocky Mountains. *Geological Society of America Bulletin* 102: 1093-1104.
- Byers, F. M., Jr. 1960. geology of the Alvord Mountain Quadrangle, San Bernardino County, California. *U. S. Geological Survey Bulletin*, 1089-A:71P.
- Cole, J. M., E. T. Rasbury, G. N. Hanson, I. P. Montañez and V. A. Pedone, 2005. Using U-Pb ages of Miocene tufa for correlation in a terrestrial succession, Barstow Formation, California. *Geological Society of America Bulletin* 2005;117:276-287.
- Cooper, J. F., Jr., G. E. Dunning, T. A. Hadley, W. P. Moller, and R. E. Reynolds. 2002. The Sulfur Hole, Calico District, San Bernardino County, California. *Calif. State Univ. Fullerton, Desert Symposium 2002 Proceedings Volume*. p.29-36.
- Dibblee, T. W., Jr. 1968. Geology of the Opal Mountain and Fremont Peak quadrangles, California. *California Division of Mines and Geology, Bulletin* 188:64p.
- Dibblee, T. W., Jr. 1970. Geologic map of the Daggett Quadrangle, California. San Bernardino County, California. *U. S. Geological Survey Miscellaneous Geologic Investigations Map I-592*, 1:62,500.
- Dibblee, T. W., Jr. and A. M. Bassett. 1966a. Geologic map of the Newberry Quadrangle, California. San Bernardino County, California. *U. S. Geological Survey Miscellaneous Geologic Investigations Map I-461*, 1:62,500.
- Dibblee, T. W., Jr. and A. M. Bassett. 1966b. Geologic map of the Cady Mountains Quadrangle, California. San Bernardino County, California. *U. S. Geological Survey Miscellaneous Geologic Investigations Map I-467*, 1:62,500.
- Durrell, Cordell. 1953. *Geological Investigations of strontium deposits in southern California*. California Division of Mines and Geology Special Report 32:23-36.
- Dokka, R.K., 1989, The Mojave extensional belt of southern California: Tectonics, v. 8, p. 363-390.
- Dokka, R.K. and A.F. Glazner. 1982. Aspects of early Miocene extension of the central Mojave Desert, pp. 31-45, in J.D. Cooper (compiler), Volume and guidebook, "Geologic Excursions in the California Desert," 78th Annual Meeting, Cordilleran Section, Geological Society of America.
- Dokka, R.K., McCurry, M.M., Woodburne, M.O., Frost, E.G., and Okaya, D.A., 1988, A field guide to the Cenozoic crustal structure of the Mojave Desert, in, Weide, D.L., and Faber, M.L., eds. This extended land: Geological journeys in the southern Basin and Range: Geological Society of America, Cordilleran Meeting, field trip guidebook, p. 21-44.

- Dokka, R.K. and C.J. Travis. 1990. Late Cenozoic strike-slip faulting in the Mojave Desert, California. *Tectonics*, 9:311-340.
- Dokka, R.K., Henry, D.J., Ross, T.M., Baksi, A.K., Lambert, J., Travis, C.J., Jones, S.M., Jacobson, C., McCurry, M.M., Woodburne, M.O., and Ford, J.P., 1991, Aspects of the Mesozoic and Cenozoic geologic evolution of the Mojave Desert: Geological Society of America, San Diego, Annual Meeting, field trip guidebook.
- Fillmore, R.P., and Walker, J. D., 1996, Evolution of a supradetachment extensional basin: The lower Miocene Pickhandle Basin, central Mojave Desert, California: Geological Society of America Special Paper 303, p. 107-126.
- Glazner, A.F., Nielson J.E., Howard K.A., and Miller D.M., 1986, Correlation of the Peach Spring Tuff, a large-volume Miocene ignimbrite sheet in California and Arizona: *Geology*, v. 14, p. 840-843.
- Glazner, A.F., Walker, J.D. Bartley, J.M., and Fletcher, J.M., 2002, Cenozoic evolution of the Mojave block of southern California: Geological Society of America Memoir 195, p. 19-41.
- Hillhouse, J.W., D.M. Miller, and B.D. Turrin. 2010. Correlation of the Miocene Peach Spring Tuff with the geomagnetic polarity time scale and new constraints on tectonic rotations in the Mojave Desert, California, *in* Overboard in the Mojave: 20 million years of lakes and wetlands, R.E. Reynolds and D.M. Miller (eds). California State University Fullerton, Desert Studies Consortium, this volume.
- Leggitt, V. L., 2000, Three-Dimensional microfossils from Rainbow Basin: Barstow Formation, Mojave Desert, California: *Geol. Soc. Amer., Abstracts*, 32(7) p. A-15.
- Leggitt, V. Leroy, 2002. Preliminary Report on the Stratigraphic Setting and Microstructure of Avian Eggshell Fragments from the Calico Mountains: Barstow Formation, Mojave Desert, California. *in* Reynolds, R.E., (ed), 2002. Between the basins: exploring the western Mojave and southern Basin and Range Province. California State University, Fullerton, Desert Studies Symposium: 51-57.
- Leslie, S.R., D.M. Miller, J.L. Wooden, and J.A. Vazquez. 2010. Stratigraphy, age, and depositional setting of the Miocene Barstow Formation at Harvard Hill, central Mojave Desert, California, *in* Overboard in the Mojave: 20 million years of lakes and wetlands, R.E. Reynolds and D.M. Miller (eds). California State University Fullerton, Desert Studies Consortium, this volume.
- Lindsay, E. H., 1972. Small mammal fossils from the Barstow Formation, California. University of California Publications in Geological Sciences, 93:104p.
- Lindsay, E. H. 1991. Small mammals near the Hemingfordian/Barstovian boundary in the Barstow syncline. *Redlands: San Bernardino County Museum Association Quarterly* 38(3,4): 78-79.
- Lindsay, E.H., 1995. Copemys and the Barstovian/Hemingfordian boundary. *Journal of Vertebrate Paleontology* 15(2):357-365.
- Lindsay, E.H., and R.E. Reynolds. 2008. Heteromyid rodents from Miocene faunas of the Mojave Desert. pp.213-235, *in* X. Wang and L. G. Barnes (eds.), *Geology and vertebrate paleontology of western and southern North America, Contributions in honor of David P. Whistler*, Natural History Museum of Los Angeles County Science Series 41:i-viii, 1-388.
- Link, M.H., 1980, Sedimentary facies and mineral deposits of the Miocene Barstow Formation, California, *In* Fife, D.L. and Brown, A.R., eds., *Geology and Mineral Wealth of the California Desert: South Coast Geological Society*, p. 191-203.
- MacFadden, B.J., Swisher, C.C. III, Opdyke, N.D., and Woodburne, M.O., 1990, Paleomagnetism, geochronology, and possible tectonic rotation of the middle Miocene Barstow Formation, Mojave desert, southern California: Geological Society of America Bulletin, v. 102, p. 478-493.
- Merriam, J. C., 1919. Tertiary mammalian faunas of the Mohave Desert. University of California Publications in Geological Sciences, 86:437-585.
- Miller, D.M., 1994, Cenozoic deposits in the Lava Hills and southern Bristol Mountains, southeastern California, *in* Sherrod, D.R., and Nielson, J.E., eds., *Tertiary Stratigraphy of highly extended terranes, California, Arizona, and Nevada: U.S. Geological Survey Bulletin* 2053, p. 99-107.
- Miller, L., 1952. The avifauna of the Barstow Miocene of California. *Condor*, 54: 296-301.
- Miller, L., 1966. An addition to the bird fauna of the Barstow Miocene. *Condor*, 68(4): 397.
- Miller, S. T., 1980. Geology and mammalian biostratigraphy of a part of the northern Cady Mountains, California. U. S. Geol. Survey OFR. 80-978:121p.
- Miller, D.M., S.R. Leslie, J.W. Hillhouse, J.L. Wooden, J.A. Vazquez, and R.E. Reynolds. 2010. Reconnaissance geochronology of tuffs in the Miocene Barstow Formation: implications for basin evolution and tectonics in the central Mojave Desert *in* Overboard in the Mojave: 20 million years of lakes and wetlands, R.E. Reynolds and D.M. Miller (eds). California State University Fullerton, Desert Studies Consortium, this volume.
- Moseley, C. G., 1978. Geology of a portion of the northern Cady Mountains, Mojave Desert, California. Univ. Calif. Riverside, Dept. of Earth Sciences (Master's Thesis).
- Pagnac, Darrin, 2009. Revised large mammal biostratigraphy and biochronology of the Barstow Formation (Middle Miocene), California. Univ. Calif. Museum of Paleontology, *PaleoBios* 29(2):48-59.
- Reheis, M. C., D. M. Miller, and J. L. Redwine. 2007. Quaternary Stratigraphy, Drainage-Basin Development, and Geomorphology of the Lake Manix Basin, Mojave Desert, Guidebook, Friends of the Pleistocene, Pacific Cell, October 4–7, 2007. U. S. Geological Survey Open-File Report 2007–1281.
- Reheis, M.C. and D.M. Miller. 2010. Environments of nearshore lacustrine deposition in the Pleistocene Lake Manix basin, south-central California *in* Overboard in the Mojave: 20 million years of lakes and wetlands, R.E. Reynolds and D.M. Miller (eds). California State University Fullerton, Desert Studies Consortium, this volume.
- Reynolds, R.E., 1991a. Biostratigraphic relationships of Tertiary small vertebrates from Cajon Valley, San Bernardino County, California, *in* Inland southern California: the last 70 million years, M.O. Woodburne, R.E. Reynolds, and D.P. Whistler, ed. *Redlands, San Bernardino County Museum Association Quarterly* 38(3,4):54-59.
- Reynolds, R. E., 1991b. Hemingfordian/Barstovian Land Mammal Age Faunas in the Central Mojave Desert, Exclusive of the Barstow Fossil Beds. *In* Inland Southern California: The Last 70 Million Years, ed. M.O. Woodburne, R.E. Reynolds, and D.P. Whistler. *Redlands: San Bernardino County Museum Association Quarterly* 38(3,4):88-90.
- Reynolds, R. E., 1998. Flamingo egg from the Miocene sediments of the Calico Mountains, San Bernardino County, California. *San Bernardino County Museum Association Quarterly*, 45(1, 2):106.
- Reynolds, R.E., 2000. Marker units suggest correlation between the Calico Mountains and the Mud Hills, central Mojave Desert, California, *in* Reynolds, R.E. and Reynolds, J. (eds), *Empty Basins, Vanished Lakes*. San Bernardino County Museum Association Quarterly, 47(2): 3-20.
- Reynolds, R.E. 2001. Marker bed correlations between the Mud Hills, Calico Mountains, and Daggett Ridge, central Mojave Desert, California. Geological Society of America Abstracts with Programs, Cordilleran section, GSA and Pacific Section 33(3): A-70.

- Reynolds, R.E., 2003a. Widespread early Miocene marker beds unite the Barstow Formation, central Mojave Desert. *Western Association of Vertebrate Paleontologists, Abstracts and Program*.
- Reynolds, R.E., 2003b. Reuniting the Barstow Basin. *Western Association of Vertebrate Paleontologists Annual Meeting Field Trip Guide*: 11 p.
- Reynolds, R.E., and Woodburne, M.O., 2001. Review of the Proboscidean datum within the Barstow Formation, Mojave Desert, California. *Journal of Vertebrate Paleontology, Abstracts of Papers*, 21(3): 93A.
- Reynolds, R.E., and Woodburne, M.O., 2002a. Marker bed correlations between the Mud Hills, Calico Mountains, and Daggett Ridge, central Mojave Desert, California [abs], in Reynolds, R.E., (ed), 2002. *Between the basins: exploring the western Mojave and southern Basin and Range Province*. California State University, Desert Studies Consortium: 82.
- Reynolds, R.E., and Woodburne, M.O., 2002b. Review of the Proboscidean datum within the Barstow Formation, Mojave Desert, California [abs], in Reynolds, R.E., (ed), 2002. *Between the basins: exploring the western Mojave and southern Basin and Range Province*. California State University, Desert Studies Consortium: 82-83.
- Reynolds, R.E., 2004. Widespread early Miocene marker beds unite the Barstow Formation of the central Mojave Desert: *Inland Geological Society*, p. 1-10.
- Reynolds, R. E., R. L. Reynolds, and E. H. Lindsay. 2008. Biostratigraphy of the Miocene Crowder Formation, Cajon Pass, southwestern Mojave Desert, California; pp.237-253, in X. Wang and L. G. Barnes (eds.), *Geology and vertebrate paleontology of western and southern North America, Contributions in honor of David P. Whistler*, Natural History Museum of Los Angeles County Science Series 41:i-viii, 1-388.
- Sarjeant, W.A.S., and Reynolds, R.E., 1999. Camelid and horse footprints from the Miocene of California and Nevada in Reynolds, R.E., 1999. *Fossil Footprints*. San Bernardino County Museum Association Quarterly, 46(2): 3-20.
- Schuling, W. T., 1999. A Miocene hot spring exhalite in the southern Calico Mountains, Redlands: *San Bernardino County Museum Association Quarterly* 46(3):89-94.
- Singleton, 2004. *Geologic evolution of the southeastern Calico Mountains, central Mojave Desert, California*. Unpublished M. S. thesis, Univ. Calif. at Santa Barbara., p. 94.
- Singleton, J.S., and Gans, P.B., 2008, *Structural and stratigraphic evolution of the Calico Mountains: Implications for early Miocene extensions and Neogene transpression in the central Mojave Desert, California: Geosphere*, v. 4, p. 459-479.
- Spencer, R. S., 2006, *Arthropod-bearing lacustrine deposits, Black Canyon, Barstow Formation*. M. S. Geology, Loma Linda Univ., Ca., p. 100.
- Steinen, R. P., 1966. *Stratigraphy of the middle and upper Miocene Barstow Formation, San Bernardino County, California*. Univ. Calif. Riverside, Dept. of Geol. Sci., MS Thesis..
- Taylor, D. W., 1954. *Nonmarine mollusks from the Barstow Formation of southern California*. U. S. Geol. Survey Prof. Paper 254-C, p. 67-80.
- Tedford, R. H. L. B. Albright III, A. D. Barnosky, I. Ferrusquia-Villafraña, R. M. Hunt Jr., J. E. Storer, C. C. Swisher III, M. R. Voorhies, S. D. Webb and D. P. Whistler. 2004 *Mammalian Biochronology of the Arikarean through Hemphillian Interval. In Late Cretaceous and Cenozoic Mammals of North America: Biostratigraphy and Geochronology*, ed. M.O.Woodburne. New York: Columbia University Press: 169-231.
- Weber, H. F., Jr., 1966. *Silver Mining in Old Calico*. Mineral Information Service, Calif. Div. Mines & Geol., Vol. 19 (5), p. 71-80.
- Weber, H. F., Jr., 1967. *Silver Deposits of the Calico District*. Mineral Information Service, Calif. Div. Mines & Geol., Vol. 20 (1), p. 3-8.
- Wells, R.E., and Hillhouse, J.W., 1989, *Paleomagnetism and tectonic rotation of the lower Miocene Peach Spring Tuff: Colorado Plateau, Arizona, to Barstow, California: Geological Society of America Bulletin*, v. 101, p. 846-863.
- Williamson, D. A., 1980. *The geology of a portion of the eastern Cady Mountains, Mojave Desert, California*. Univ. Calif. Riverside, Dept. of Earth Sciences (Master's Thesis).
- Wood, H. E., II, R. W. Chaney, J. Clark, E. H. Colbert, G. L. Jepson, J. B. Reedide, Jr., and C. Stock. 1941. *Nomenclature and correlation of the North American continental Tertiary*. Geological Society of America., 52:1-48.
- Woodburne, M.O. 1991. *The Mojave Desert Province*. In *Inland Southern California: The Last 70 Million Years*, ed. M.O. Woodburne, R.E. Reynolds, and D.P. Whistler. *Redlands: San Bernardino County Museum Association Quarterly* 38(3,4):60-77.
- Woodburne, M. O., 1996. *Precision and resolution in mammalian chronostratigraphy: principals, practices, examples*. *Jour. Vert. Paleo.*, 16(3). 531-555.
- Woodburne, M. O., 1998. *Arikarean and Hemingfordian faunas of the Cady Mountains, Mojave Desert Province, California*. *Geol. Soc. Amer. Special Paper* 325, p. 197-210.
- Woodburne, M. O. and C. C. Swisher, III. 1995. *Land Mammal high-resolution geochronology, intercontinental overland dispersals, sea level change, climate, and vicariance*. In *Geochronology, Time Scales and Global Stratigraphic Correlations*, ed. W. A. Berggren, D. V. Kent, M-P. Aubry, and J. Hardenbol. *SEPM Special Publication* 54:335-364.
- Woodburne, M. O., R. H. Tedford, and C. C. Swisher, III. 1990. *Lithostratigraphy, Biostratigraphy and Geochronology of the Barstow Formation, Mojave Desert, southern California*. *Geological Society of America*, 102:459-477.
- Woodburne and Reynolds, 2010 (This vol) ; *2010 Desert Symposium Volume*, California State University, Desert Studies Consortium. p. X-X.

Additional Reading: Environmental Data

- Belk, D. and Schram, F. R., 2001, *A new species of anostracans from the Miocene of California: Jour. of Crustacean Biology*, v. 21(1), p. 49-55.
- Degens, T. W., Pierce, W. D., and Chillingar, G.V., 1962, *Origin of petroleum-bearing freshwater concretions of Miocene age: Amer. Assoc. Petr. Geologists Bulletin*, v. 46, no. 8, p. 1522-1525.
- Jenkins, J. E., 1986. *Miocene invertebrates from the Calico Mountains, San Bernardino County, California*. *San Bernardino County Museum Association Quarterly*, Vol. 33., 42p.
- Kirkby, R. A., 1959, *Pseudomorphs after Life*, in *Gems and Minerals*, Nov., p.22-23.
- Leggitt, V. L., 2001, *Three-Dimensional arthropods from microlaminated sediments of the Barstow Formation, Mojave Desert, California: Geol. Soc. Amer. Bulletin*, v. 333, no. 3, p. A54.
- Leggitt, V. L., 2002, *Preliminary Report on the Stratigraphic Setting and Microstructure of Avian Eggshell Fragments from the Calico Mountains, Barstow Formation, Mojave Desert, California*, CSU Fullerton, *Desert Studies Symposium Volume*, p. 51-58.
- Leggitt, V. L., 2003, *Silicified microfossils from Black Canyon: Miocene Barstow Formation, San Bernardino County, California*, *Geol. Soc. Amer. Abstracts with programs*, v. 34, no. 6, p. 495.

- Lister, K. H., 1970, Paleocology of insect-bearing Miocene beds in the Calico Mountains, California: M. A. Thesis, Univ. Calif., Los Angeles, 108pp.
- Palmer, A. R., 1957. Miocene arthropods from the Mojave Desert: U. S. Geological Survey Professional Paper 294-G, p. 237-280.
- Palmer, A. R., and A. M. Bassett, 1954, Nonmarine Miocene arthropods from California. *Science* 120: 228-229.
- Palmer, A. R., 1960. Miocene copepods from the Mojave Desert: *Journal of Paleontology*, v. 34, p. 447-452.
- Park, L. E., 1995, Geochemical and paleoenvironmental analysis of lacustrine arthropod bearing concretions of the Barstow Formation, Southern California: *Palaios*, V. 10, p.44-57.
- Park, L. E., and K. F. Downing, 2001, Paleocology of an exceptionally preserved arthropod fauna from lake deposits of the Miocene Barstow Formation, Southern California, U. S. A.: *Palaios*, V. 16, p.175-184.
- Pierce, W. D., 1958, Fossil arthropods of California, No. 21, Termites from Calico Mountains nodules: *Bull., Southern Calif. Acad. Sci.* Vol. 57, Part 1, p.14.
- Pierce, W. D., 1959, Fossil arthropods of California, No. 22, A progress report on the nodule studies: *Bull., Southern Calif. Acad. Sci.* Vol. 58, p.72-78.
- Pierce, W. D., 1960a, Fossil arthropods of California, No. 23, Silicified insects in Miocene nodules from the Calico Mountains: *Bull., Southern Calif. Acad. Sci.* Vol. 59, p.138-143.
- Pierce, W. D., 1960b, Silicified Turbellaria from Calico Mountains nodules: *Bull., Southern Calif. Acad. Sci.* Vol. 59, p.138-143.
- Pierce, W. D., 1962. The significance of petroliferous nodules of our desert mountains: *Bull., Southern Calif. Acad. Sci.* Vol. 61, Part 1, p.7-14.
- Pierce, W. D., and J. Gibron, Sr., 1962. Fossil arthropods of California, No. 24, Some unusual arthropods from the Calico Mountains nodules: *Southern Calif. Acad. Sci.* Vol. 61, Part 3, p.143.
- Seiple, E., 1983, Miocene insects and arthropods in California: *California Geology*, November, Calif. Div. Mines and Geology. p. 246-248.

Miocene and Late Pleistocene stickleback spines from the Mojave Desert, California

Michael A. Bell¹ and Robert E. Reynolds²

¹Department of Ecology and Evolution, Stony Brook University, Stony Brook, NY 11794-5245, US

mabell@life.bio.sunysb.edu

²Redlands, California, 92373, rreynolds220@verizon.net

Abstract

Sticklebacks are small, primitively marine fishes with robust dorsal and pelvic spines. Threespine and ninespine sticklebacks both occur in the Pacific basin and colonize fresh water, but the ninespine is restricted to high latitudes in the eastern Pacific. Sticklebacks are weak swimmers, and freshwater sticklebacks rarely ascend high-gradient streams. Thus, they are usually restricted to coastal lowlands below 200 m elevation. Stickleback spines are distinctive and preserve well in fluvial sedimentary deposits. The Argonaut site is east of Yermo, California in sediments of the Miocene Barstow Formation, and it is isolated from the Pacific Ocean and Colorado River drainage by passes at elevations of 700 meters or greater. Occurrence of fossil sticklebacks prior to 16 Ma at this site indicates regional colonization by sticklebacks while the Miocene Barstow Basin was part of a lowland coastal drainage.

Introduction

Sticklebacks of the family Gasterosteidae are small, primitively marine, teleost fishes from the Atlantic and Pacific basins in the northern hemisphere (Wootton 1976; Nelson 2006; Kawahara et al. 2009). The family comprises five genera, three of which are monospecific. Bowne (1994) provided a detailed description of stickleback osteology, Wootton (1976, 1984) summarized their geographic distribution, and Kawahara et al. (2009) recently published a robust cladogram based on extensive DNA sequence data. Sticklebacks are named for the presence of two to 15 free spines anterior to the dorsal fin, which itself begins with a weak spine. They also have robust pelvic spines that articulate with a pelvis that is highly derived in size and shape in all stickleback genera except *Spinachia* (Bowne 1994). The dorsal and pelvic spines are distinctive and preserve well as isolated elements in fluvial deposits (Bell 1994).

Only two of the five stickleback genera, *Gasterosteus* (threespine) and *Pungitius* (ninespine), occur in the north Pacific (Wootton 1976, 1982), where they have also been reported from Neogene deposits (Bell 1994, 2009; Bell et al. 2009). The earliest articulated stickleback specimen that can be assigned with confidence to species is a 13 Ma *Gasterosteus aculeatus* from a marine deposit in Palos Verdes, California (Bell et al. 2009). Several marine and freshwater deposits in western North America dated at about

10 Ma contain articulated specimens of *G. aculeatus* (Bell 1994). Only one *Pungitius* specimen from a 7 Ma deposit near Homer, Alaska has been reported (Rawlinson and Bell 1982). The nominal species *Pungitius haynesi* (David 1945) from the Ridge Formation of southern California is actually a threespine stickleback (Bell 1973a). The threespine stickleback has been in the southern California region for at least 13 Ma, but the paleozoogeography of the ninespine is too limited to interpret.

Historically, *G. aculeatus* has occurred as far south as northern Baja California (Miller and Hubbs 1969; Riuz-Campos et al. 2000; Sánchez-González et al. 2001) at about 32° north latitude. The southern range limit of *Pungitius* in the eastern Pacific is in the Aleutian Islands, Kodiak Island, and Cook Inlet, Alaska between 50° and 60° north latitude (Wootton 1976, 1984; Mecklenberg et al. 2002), but *Pungitius* occurs much farther south in the western Pacific to about 32° north latitude at Shanghai, China. (Multiple *Pungitius* taxa have been recognized in Asia, but their systematics is uncertain [Mattern 2007].) There is no evidence that *Pungitius* ever occurred as far south in the eastern Pacific as the Mojave Desert, at about 34 degrees north.

The zoogeography and paleozoogeography of freshwater fishes provide excellent insights into paleohydrography (Smith 1978, 1981). The distribution of sticklebacks

results in part from movement through the ocean. All but one derived genus of the Gasterosteidae, *Culaea* (brook stickleback), includes marine populations, and the Aulorhynchidae, the sister group to the Gasterosteidae, is strictly marine (Wootton 1976, 1984; Kawahara et al. 2009). The geographic distribution of threespine stickleback (Bell 1976; 1995; Bell and Foster 1994) and their phylogeography (Taylor and McPhail 1999; 2000; Colosimo et al. 2004) both require frequent colonization of fresh water by marine and anadromous (sea run) stickleback. The distribution of ninespine stickleback appears to demand the same dispersal routes. However, sticklebacks are weak swimmers, and freshwater populations rarely ascend high-gradient streams or occur above 200 m elevation (Hardy 1978; Nelson 2006). Their presence in a habitat indicates that they colonized it while there was low-elevation or low-gradient access from the ocean.

In this report, we describe threespine stickleback spines to facilitate their identification as isolated elements from fluvial deposits, redescribe stickleback spines from the Argonaut site in the Miocene Barstow Basin, and review the historical and Pleistocene distribution of *G. aculeatus* in southern California. These records indicate that stickleback occurred in coastal streams that once drained the Miocene Barstow Basin to the Pacific, and may have survived in that interior basin for the last 16 Ma. Alternatively, stickleback may have gone extinct in this basin and recolonized Mojave Desert lakes during the Late Pleistocene from high-elevation headwater streams in the Transverse Ranges.

Threespine stickleback spines

The dorsal and pelvic spines of threespine stickleback consist of dense bone that preserves well, even in high-energy depositional environments where other fish bones tend to fragment. They are distinctive and can easily be assigned to the Gasterosteidae (Fig. 1). They vary in size, depending both on the size of the specimen and on the population from which it came. Relative spine lengths tend to be greater in stickleback from populations that are

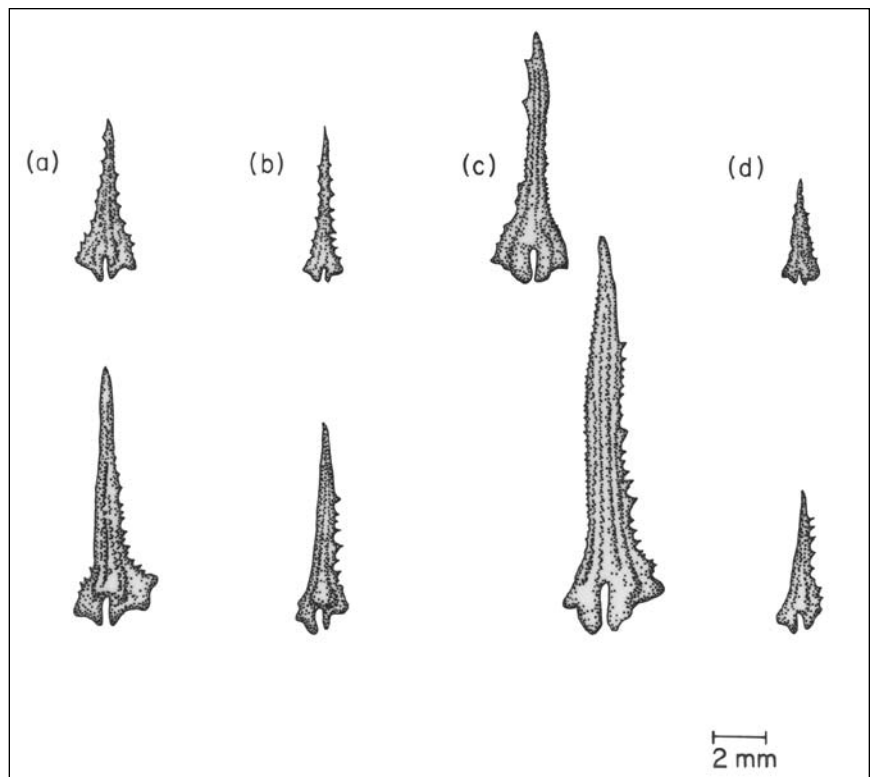


Figure 1. Anterior aspect of dorsal (upper) and right pelvic spines (lower) from modern *Gasterosteus aculeatus* showing their diagnostic structures, including the broad basal flanges, proximal processes, and marginal serrations. Source populations: (a) Birch Cove, Cook Inlet Alaska (71.3 mm, anadromous), (b) Big Lake, Cook Inlet, Alaska (55.3 mm), (c) Bakewell Lake, Portland Canal, Alaska (73.3 mm), and (d) Santa Clara River, California (43.8 mm, *G. aculeatus williamsoni*). Body sizes in parentheses are standard length, the distance from tip of upper jaw to end of vertebral column. From Bell (1994).

exposed to predation by fishes (Hagen and Gilbertson 1972; Reimchen 1994). The spines of large specimens from populations subject to intense fish predation range up to about 15 mm long (Fig. 1 c) but are usually smaller. The smaller anal fin spine and third dorsal spines are similar but shorter and may be almost as wide at the base as they are long.

The proximal ends of the spines are divided into a pair of processes that grasp a condyle on either a dorsal pterygiophore (dorsal spine) or the pelvic girdle (Hoogland 1951). Bilaterally symmetrical flanges are located on the lateral edges of the dorsal spine just distal to the proximal processes. The corresponding flange along the dorsal edge of the pelvic spine is larger and more distal than its smaller counterpart on the ventral edge. (There is a distinctive cusp on the dorsal edge of the pelvic spine of the blackspotted stickleback, *Gasterosteus wheatlandi*, which is endemic to the north Atlantic Ocean [Hubbs 1929].) Asymmetry of the basal flanges on opposite edges of the pelvic spines distinguish them from dorsal spines and can be used to determine whether fossil pelvic spines are from the right or left side (Fig. 1, 2).

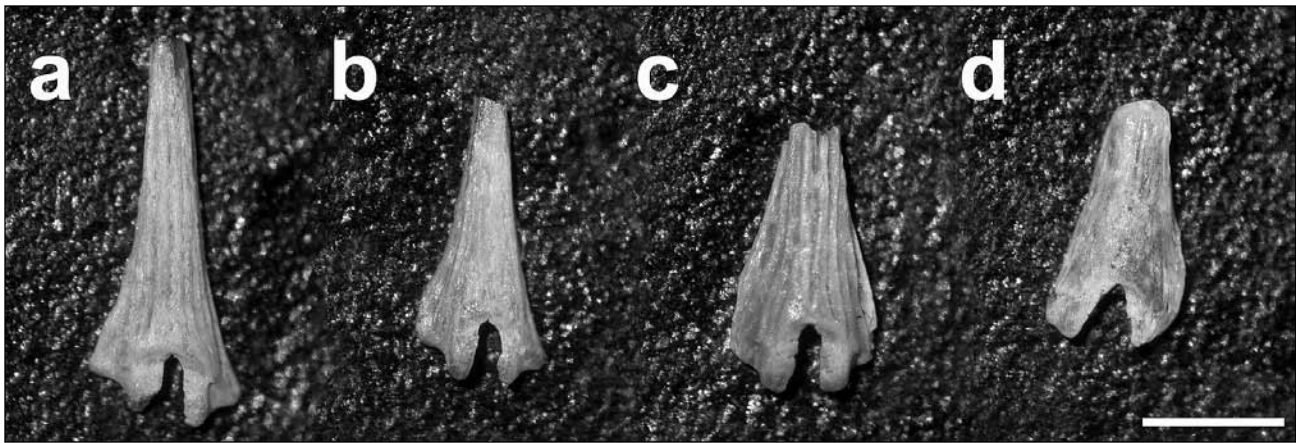


Figure 2. Anterior aspect of fossil stickleback spines from the Argonaut site (SBCM RPLI 1.76.43). The degree of damage and wear increases from a to d. The scale bar is 1 mm. (a) left pelvic spine with broken tip (SBCM 1814-1014), (b) left pelvic spine with broken tip (SBCM 1814-1015), (c) slightly worn and broken spine (SBCM 1814-1013), and (d) severely worn spine that is probably a stickleback spine (SBCM 1814-1017).

The shaft of the spines usually bears a series of large denticulations or serrations along the lateral edges of the dorsal spines and the dorsal and ventral margins of the pelvic spines. These denticulations become smaller distally. In addition, there are usually rows or scattered denticulations along the anterior surface of the spine, but they may be represented by relatively smooth ridges (Fig. 2). Spine denticulation in *G. aculeatus* tends to be greater in European populations subject to fish predation (Gross 1978), and this relationship appears to apply to western North American populations, as well (Bell, pers. obs.). The edges of the spines of the fossil threespine stickleback, *Gasterosteus doryssus*, from west-central Nevada (Bell 1994; 2009) almost always lack denticulations and are completely smooth. The posterior surfaces of both spines are slightly concave and smooth.

Taphonomic processes may reduce denticulation of stickleback spines. They may be absent or reduced to a series of gentle waves on the edges of spines from deposits in which most of the spines are denticulated. This variation may result either from wear during transport or acid etching in the gastrointestinal tract of a predatory birds that regurgitate fish bones (Wilson 1987; Bell 1994, 2009).

Fossil stickleback spines from the Argonaut site

Methods and Materials

Six stickleback spines reported previously by Reynolds (1991) from the Argonaut site (SBCM 1.76.43) are in the San Bernardino County Museum (SBCM) Department of Geological Sciences collection (Appendix 1). They were examined under a Leica S6D dissecting microscope, and digital images of four spines showing the range of

preservation, were captured using a Cannon PowerShot S45 Camera mounted on this microscope.

Results

The general morphology of these specimens, including the long proximal processes, expanded basal flanges, and longitudinal ridges generally resemble those of modern threespine stickleback spines (Fig. 1). The two best-preserved specimens (Fig. 2a, b) are left pelvic spines with smooth edges and anterior surfaces with longitudinal grooves and ridges instead of rows of denticles. A third specimen (Fig. 2c) also has longitudinal grooves and ridges, but only the proximal half of the spine is present, and the basal flanges have been worn off. It also appears to be a left pelvic spine, but that is uncertain. The final specimen is severely worn and rounded. Its identity would be questionable unless there were other more well-preserved stickleback spines in the sample. Its symmetry suggests that it is a right pelvic spine. These spines could represent *G. aculeatus*, but their morphology does not preclude other sticklebacks.

Distribution of extant threespine stickleback within and near the Mojave Desert

G. aculeatus has historically been widespread in southern California but almost exclusively at low elevations in coastal drainages (Miller and Hubbs 1969; Ross 1973). Important exceptions to this generalization include an extinct population in the Mojave River (an endorheic drainage; Miller and Hubbs 1969) and phenotypically divergent, extant populations in Holcomb Creek, a tributary to the Mojave River at 1960 m elevation (Bell 1982), and in the endorheic Shay Creek drainage at more than 2050 m elevation (Haglund and Buth 1988; Moyle 2002). *G. aculeatus williamsoni* occurs in the headwaters of

the Santa Clara River (a coastal drainage) at an elevation of 760 m, just down slope from the Palmdale saddle point 55 km (33 miles) west of Mojave River drainage tributaries (Miller 1960; Bell 1978). *G. aculeatus* also occurs at an elevation of 1140 m in the headwaters of Sespe Creek, a western tributary of the Santa Clara River (Bell 1978; unpubl. data.). Threespine stickleback populations also occurred in the Los Angeles and Santa Ana rivers, which are adjacent coastal drainages to the south of the Mojave Desert and the San Bernardino Mountains (Miller and Hubbs 1969; Ross 1973). Historical occurrence of threespine stickleback populations in drainages around the Mojave Desert coupled with their usual absence at high elevations or inland of high-elevation barriers throughout their global distribution suggests that they did not disperse across high elevation barriers to enter the Mojave Desert region. Instead, their presence in the Mojave Desert and in adjacent high-elevation streams suggests that threespine stickleback were present in drainages that transgressed these areas before they were uplifted and either persisted at high elevations or became isolated inland of the mountains.

This inference is supported by Vaughan's (1922; see also Hinds 1952; Norris and Bebb 1976) interpretation of the physiography of Holcomb Valley near San Geronio Mountain on Holcomb Creek in the headwaters of the Mojave River. He concluded that Holcomb Valley was originally part of a coastal drainage that was uplifted without significant tipping and erosion. A more recent analysis by Binnie et al. (2007) was consistent with this interpretation. Although the San Bernardino Mountains were glaciated during the Pleistocene, glaciers did not occur below 2650 m elevation (Sharp et al. 1959). Thus, the Holcomb Valley drainage originally connected to a coastal plain drainage, where it is likely to have been colonized from the Pacific Ocean by threespine stickleback. Threespine stickleback could have persisted in Holcomb Valley (Bell 1982) during rapid uplift of the Transverse Ranges after 6 Ma (Meisling and Weldon 1989; Reynolds et al. 2008) and seeded lower-elevation reaches of the Mojave River during the late Pleistocene glacial maximum (Reynolds 2005). If stickleback populations represented in the Argonaut site became extinct in the Miocene due to desiccation, populations from Holcomb Creek or other streams, which had been in coastal plain drainages before uplift of the Transverse Ranges, could have recolonized the Mojave Desert after they were captured by Mojave River tributaries that flowed to Pleistocene lakes. The Mojave River did not become a cohesive drainage that reached Lake Manix until after 500 Ka (Jefferson, 1991).

Fossil stickleback from southern California

Southern California has the most extensive fossil record for threespine stickleback in the world. Two complete, articulated specimens from marine deposits show that they were present in the adjacent Pacific Ocean at least 13 Ma (Bell 1977, Bell et al. 2009). Complete articulated remains from freshwater deposits west of the Mojave Desert have been reported from the Late Miocene Hemphillian North American Land Mammal Age (NALMA) deposits of the Ridge Formation (Crowell 2003) of the southeastern Coast Range Province (Jahns, 1954; Bell 1973b). *G. aculeatus* also occurs in the 2.2 Ma Bautista Formation to the south in the northern Peninsular Range Province (Jahns, 1954; Bell 1973a). Other fossil stickleback deposits have been reported elsewhere in southern California and Nevada (reviewed by Bell 1994; 2009).

There have been several reports of fossil stickleback spines from within the Mojave Desert. Although these spines are consistent with the morphology of *G. aculeatus*, it is the only fossil stickleback species reported from southern California, and it is the only extant native California stickleback (Moyle 2002), it would be premature to assign Miocene fossil stickleback spines from the Mojave Desert to *G. aculeatus*. While stickleback spines from more recent deposits almost certainly represent *G. aculeatus*, earlier records could represent ninespine stickleback, which occur at comparable latitudes to the Mojave Desert elsewhere, or an extinct stickleback that is unknown as articulated remains from the fossil record.

Reynolds (1991) reported the earliest occurrence of fossil stickleback spines in the Mojave Desert from the Argonaut locality in the Barstow Formation in the Toomey (Yermo) Hills. We re-examined six fish spines from this locality (Fig. 2; Appendix 1) to verify their identification, and they clearly exhibit the morphology of stickleback spines (Fig. 1). Fossil mammals from this deposit indicate that it is at the transition between the Hemingfordian and Barstovian NALMA (Woodburne, 1991; Tedford et al. 2004; Woodburne this volume). Thus, available dates indicate that they are at least 16 Ma.

There have also been several reports of stickleback spines from Late Pleistocene fluvial and lacustrine deposits along the Mojave River. Jefferson (1991) reported diverse fossil vertebrates, including a single stickleback element from the Camp Cady Local Fauna, which ranges in age between 300 and 20 Ka. The composition of fossil animal (Jefferson 1991) and plant (Jefferson 1987) assemblages indicate relatively cool conditions that would favor occurrence of *G. aculeatus*. Late Pleistocene occurrences of stickleback from fluvial deposits along the Mojave River at Daggett (Reynolds and Reynolds, 1985;

1991a; Roeder, 1985; Reynolds, 2005) date to 12,210 ybp. Fossil sticklebacks are reported from north of the San Gabriel Mountains from Late Pleistocene sediments of Lake Thompson near Lancaster, (Reynolds and Reynolds, 1991b).

Discussion

Dates of fossil stickleback spines in Miocene and Pleistocene deposits from the Mojave Desert are separated by a 16 Ma hiatus. Stickleback spines from the Argonaut site in the Miocene Barstow Formation in the Toomey Hills are at least 16 Ma, and there are several reports of Pleistocene stickleback from Coyote Lake (RER pers. obs.) and Daggett (Roeder, 1985). One scenario for this long hiatus is that threespine stickleback entered the region of the Miocene Barstow Formation from the Pacific Ocean at or before 16 Ma and persisted in low-elevation, aquatic habitats there until Holocene desiccation or human activities caused their extinction (see Miller and Hubbs 1969). Persistence of stickleback in the Mojave Desert after early Barstovian NALMA time is difficult to document, since a province-wide depositional hiatus is recorded (Woodburne, 1991, this volume). Where Clarendonian and Hemphillian sediments are present (Dove Spring Fm., Whistler 1991; Avawatz Mountains, Reynolds and Whistler 1990), stickleback are absent, although the taxonomic records from the Dove Spring Formation include fish, amphibians and pond turtle (Whistler, 1991).

Alternatively, stickleback populations represented at the Miocene Toomey Hills site may or may not have belonged to *G. aculeatus* and went extinct with desiccation of the Miocene Barstow basin. Then threespine stickleback colonized the Mojave River drainage between 20 and 12 Ka from high-elevation headwater streams in the San Bernardino Mountains where threespine stickleback had persisted for the last 6 Ma during uplift of the Transverse Ranges.

The fossil stickleback spines from the early Miocene Argonaut site in the Barstow Formation are the earliest occurrence of stickleback (*Gasterosteidae*) remains reported and probably represent *G. aculeatus*. But *G. aculeatus* is one of two stickleback species from the Pacific Basin, and we cannot yet rule out the possibility that this small sample of spines represents a ninespine stickleback species (*Pungitius*) or an extinct stickleback taxon. Regardless of which stickleback occurs at the Argonaut site, these spines can be used to date the basal node that separates *Gasterosteus* from all other stickleback species (Kawahara et al. 2009).

It is interesting that the stickleback spines from the Argonaut site lack denticulations. The areas on the anterior

surfaces of the spines and along their edges form smooth ridges. This condition is consistent with absence of predatory fishes. Except for Pacific salmon of the genus *Oncorhynchus*, large, native, predatory fishes are absent from the extant fish fauna of southern California (Moyle 2002) and apparently were absent throughout the Neogene (Smith 1982). Furthermore, there are no large fish bones from the Argonaut site, so absence of denticles from the stickleback spines is expected.

Although fossil stickleback spines are easy to identify, few have been reported from the fossil record (Bell 1994). Fossil assemblages from fluvial deposits have probably not generally been searched carefully for stickleback spines, and such a search could produce earlier records than the Argonaut site. Small fossil vertebrates obtained by sediment screening are generally studied by paleomammalogists who focus on the rodent teeth and might simply identify stickleback spines as fish bones. Special attention should be given to deposits that produce the more conspicuous remains of the Pacific (or western) pond turtle, *Clemmys marmorata* (e.g., Jefferson 1991), which commonly occurs today in low-gradient coastal California streams (Germano and Bury 2001) in sympatry with extant stickleback (Bell, unpubl. observations). Careful examination of screened samples from Neogene deposits from western North America is likely to produce many additional records of fossil sticklebacks.

Acknowledgments

We thank E. Scott of the San Bernardino County Museum for making the fossil stickleback spines from the Argonaut site available for examination in 2008 and for their loan to MAB in 2010. Use of *C. marmorata* remains to identify promising deposits for fossil stickleback spines was suggested to MAB by a graduate student at the University of California, Berkeley two decades ago. Although his name has been forgotten, his as yet unexploited insight is acknowledged. We also thank G. R. Smith for critically reading the manuscript, and P. J. Park for producing and editing the digital images used in the figures. This is contribution 1195 from Ecology and Evolution at Stony Brook University.

Literature cited

- Bell, M.A. 1973a. Pleistocene threespine sticklebacks, *Gasterosteus aculeatus*, (Pisces) from southern California. *J. Paleontol.* 47:479-483.
 Bell, M.A. 1973b. The Pliocene stickleback, *Pungitius haynesi*, a junior synonym of *Gasterosteus aculeatus*. *Copeia* 1973: 588-590.
 Bell, M.A. 1976. Evolution of phenotypic diversity in the *Gasterosteus aculeatus* superspecies on the Pacific coast of North America. *Syst. Zool.* 25: 211-227.

- Bell, M.A. 1977. A Late Miocene marine threespine stickleback, *Gasterosteus aculeatus aculeatus*, and its zoogeographic and evolutionary significance. *Copeia* 1977:277-282.
- Bell, M.A. 1978. Fishes of the Santa Clara River system, southern California. *Nat. Hist. Museum Los Angeles County Contrib. Sci.* 295:1-20.
- Bell, M.A. 1982. Melanism in a high elevation stream population of *Gasterosteus aculeatus*. *Copeia* 1982: 829-835.
- Bell, M.A. 1994. Paleobiology and evolution of threespine stickleback. In *The Evolutionary Biology of the Threespine Stickleback* (M. A. Bell and S. A. Foster, eds.), pp. 438-471. Oxford University Press, Oxford.
- Bell, M.A. 1995. Intraspecific systematics of *Gasterosteus aculeatus* populations: implications for behavioral ecology. *Behaviour* 132: 1131-1152.
- Bell, M.A. 2009. Implications of fossil threespine stickleback for Darwinian gradualism. *J. Fish Biol.* 75: 1997-1999.
- Bell, M.A. & Foster, S.A. 1994. Introduction to the evolutionary biology of the threespine stickleback. In *The Evolutionary Biology of the Threespine Stickleback* (M. A. Bell and S. A. Foster, eds.), pp. 1-27. Oxford University Press, Oxford.
- Bell, M.A., J.D. Stewart and P.J. Park. 2009. The world's oldest fossil threespine stickleback. *Copeia* 2009:256-265.
- Binnie, S.A., W.M. Philips, M.A. Summerfield, and L.K. Fifield. 2007. Tectonic uplift threshold hillslopes, and denudation rates in a developing mountain range. *Geology* 35:743-746.
- Bowne, P.S. 1994. Systematics and morphology of the *Gasterosteiformes*. In *The Evolutionary Biology of the Threespine Stickleback* (M. A. Bell and S. A. Foster, eds), pp. 28-60. Oxford University Press, Oxford.
- Colosimo, P.F., C.L. Peichel, K. Nereng, B.K. Blackman, M.D. Shapiro, D. Schluter, and D.M. Kingsley. 2004. The genetic architecture of parallel armor plate reduction in threespine sticklebacks. *Public Libr. Sci., Biol.* 2: 0635-0641.
- Crowell, J.C. 2003. Introduction to geology of the Ridge Basin, southern California. In *Evolution of Ridge Basin, southern California* (J. C. Crowell, ed.), pp. 1-16. *Geological Soc. Amer. Spec. Pap.* 367, Boulder.
- David, L.R. 1945. A Neogene stickleback from the Ridge Formation of California. *J. Paleontol.* 19: 315-318.
- Germano, D.J., and R.B. Bury. 2001. Western Pond Turtles (*Clemmys marmorata*) in the Central Valley of California: status and population structure. *Trans. West. Sect. Wildl. Soc.* 37: 22-36.
- Gross, H.P. 1978. Natural selection by predators on the defensive apparatus of the three-spined stickleback, *Gasterosteus aculeatus* L. *Can. J. Zool.* 56:398-413.
- Hagen, D.W., and L.G. Gilbertson. 1972. Geographic variation and environmental selection in *Gasterosteus aculeatus* L. in the Pacific northwest, America. *Evolution* 26:32-51.
- Haglund, T.R. and D.G. Buth. 1988. Allozymes of the unarmored threespine stickleback (*Gasterosteus aculeatus williamsoni*) and identification of the Shay Creek population. *Isozyme Bull.* 21:196.
- Hardy, J.E., Jr. *Development of Fishes of the Mid-Atlantic Bight, and Atlas of Egg, Larval, and Juvenile Stages. Vol. II. Anguillidae through Syngnathidae.* Biol. Serv. Progr., U. S. Fish Wildlife Serv., Washington, D. C.
- Hinds, N.E.A. 1952. Evolution of the California landscape. *Calif. Div. Mines Bull.* 158:1-240.
- Hoogland, R.D. 1951. On the fixing-mechanism in the spines of *Gasterosteus aculeatus* L. *P. K. Ned. Akad. C Biol.* 54:171-180.
- Hubbs, C.L. 1929. The Atlantic American species of the fish genus *Gasterosteus*. *Occ. P. Museum Zool. U. Mich.* 300: 1-9.
- Kawahara, R., M. Miya, K. Mabuchi, T.J. Near, and M. Nishida. 2009. Stickleback phylogenies resolved: Evidence from mitochondrial genomes and 11 nuclear genes. *Mol. Phylogenet. Evol.* 50: 401-404.
- Jahns, R. H., 1954. Investigations and problems of southern California geology. In *Geology of Southern California, Chapter I, General Features*, Calif. Div. Division of Mines, *Bull.* 170: 5-30.
- Jefferson, G.T. 1987. The Camp Cady local fauna: paleoenvironment of the Lake Manix Basin. San Bernardino County Museum Association *Quarterly* 34:3-35.
- Jefferson, G.T. 1991. The Camp Cady Local Fauna: Stratigraphy and paleontology of the Lake Manix basin. San Bernardino County Museum Association *Quarterly* 38:93-99.
- Mecklenberg, C.W., T.A. Mecklenberg, and L.K. Thorsteinson. 2002. *Fishes of Alaska.* American Fishries Society, Bethesda.
- Meisling, K.E., and R.J. Weldon. 1989. Late Cenozoic tectonics of the northwestern San Bernardino Mountains, southern California. *Geol. Soc. Amer. Bull.* 101:106-128.
- Miller, R.R. 1960. The type locality of *Gasterosteus aculeatus williamsoni* and its significance in the taxonomy of Californian sticklebacks. *Copeia* 1960:348-350.
- Miller, R.R. and C.L. Hubbs. 1969. Systematics of *Gasterosteus aculeatus*, with particular reference to intergradation and introgression along the Pacific coast of North America: a commentary on a recent contribution. *Copeia* 1969:52-69.
- Mattern, M.Y. 2007. Phylogeny, systematic, and taxonomy of sticklebacks. In: *The Biology of the Three-Spined Stickleback* (S. Östlund-Nilsson, I. Mayer, and F. A. Huntingford, eds.), pp. 1-40. CRC Press, Boca Raton.
- Moyle, P.B. 2002. *Inland Fishes of California. Revised and Expanded.* University of California Press, Berkeley.
- Nelson, J.S. 2006. *Fishes of the World*, 4th ed. Wiley, Inc., New York.
- Norris, R.M., and R.W. Bebb. 1976. *Geology of California.* Wiley, New York.
- Rawlinson, S.E. and M.A. Bell. 1982. A stickleback fish (*Pungitius*) from the Neogene Sterling Formation, Kenai Peninsula, Alaska. *J. Paleontol.* 56:583-588.
- Reimchen, T.E. 1994. Predator and morphological evolution in threespine stickleback. In *The Evolutionary Biology of the Threespine Stickleback* (M. A. Bell and S. A. Foster, eds.), pp. 240-276. Oxford University Press, Oxford, UK.
- Reynolds, R.E. 1991. Hemingfordian/Barstovian land mammal age faunas in the central Mojave Desert, exclusive of the Barstow fossil beds. San Bernardino County Museum Association *Quarterly* 38:88-90.
- Reynolds, R.E. 2005. Late Pleistocene fluvial fish flushing introduces stickleback (*Gasterosteus* sp.) to Mojave drainages. In *Geological and Biotic Perspectives on Late Cenozoic Drainage History of the Southwestern Great Basin and Lower Colorado River Region: Conference Abstracts* (M.C. Reheis, ed.), pp. 16. U. S. Geological Survey Open-File Report 2005-1404.
- Reynolds, R.E., and R.L. Reynolds. 1985. Late Pleistocene faunas from Daggett and Yermo, San Bernardino County, California. In *Geological investigations along Interstate 15, Cajon Pass to Manix Lake: Redlands*, pp. 175-191. San Bernardino County Museum.
- Reynolds, R.E., and R. L. Reynolds. 1991a. Structural implications of late Pleistocene faunas from the Mojave River Valley, California. In *Inland southern California: the last 70 million years* (M.O. Woodburne, R.E. Reynolds, and D.P. Whistler, eds.) San Bernardino County Museum Association *Quarterly* 38: 100-105.
- Reynolds, R.E. and R.L. Reynolds. 1991b. Late Pleistocene Faunas of Lake Thompson. San Bernardino County Museum Association *Quarterly* 38: 114-115.
- Reynolds, R. E., and D. P. Whistler. 1990. Early Clarendonian Faunas of the Eastern Mojave Desert, San Bernardino County, California. San Bernardino County Museum Association *Special Publication* 90:105-108.

- Reynolds, R.E., R.L. Reynolds, and E.H. Lindsay. 2008. Biostratigraphy of the Miocene Crowder Formation, Cajon Pass, southwestern Mojave Desert, California. In *Geology and vertebrate paleontology of western and southern North America, Contributions in honor of David P. Whistler* (X. Wang and L.G. Barnes, eds.), pp. 237-253. *Nat. Hist. Museum Los Angeles County Sci. Ser.* 41, Los Angeles.
- Roeder, M.A., 1985. Late Wisconsin records of *Gasterosteus aculeatus* (threespine stickleback) and *Gila bicolor mojavensis* (Mojave tui chub) from unnamed Mojave River sediments near Daggett, San Bernardino County, California. In *Geological investigations along Interstate 15, Cajon Pass to Manix Lake* (R. E. Reynolds, ed.), pp. 171-174. San Bernardino County Museum, Redlands.
- Ross, S. T. 1973. The systematics of *Gasterosteus aculeatus* (Pisces: Gasterosteidae) in central and southern California. *Nat. Hist. Museum Los Angeles County Contrib. Sci.* 243:1-20.
- Ruiz-Campos, G., S. Contreras-Balderas, M. de L. Lozano-Vilano, S. Gonzalez-Guzman, and J. Alaniz-Garcia. 2000. Ecological and distributional status of the continental fishes of northwestern Baja California, México. *Bull. S. Cal. Acad. Sci.* 99: 59-90.
- Smith, G.R. 1978. Biogeography of intermountain fishes. In *Intermountain Biogeography: A Symposium* (K. T. Harper and J. L. Reveal, eds.), pp. 17-42. *Great Basin Naturalist Mem.* 2: 163-193.
- Smith, G.R. 1981. Late Cenozoic freshwater fishes of North America. *Ann. Rev. Ecol. Syst.* 12:163-193.
- Sánchez-González, S., G. Ruiz-Campos, and S. Contreras-Balderas. 2001. Feeding ecology and habitat of the threespine stickleback, *Gasterosteus aculeatus microcephalus*, in a remnant population of northwestern Baja California, México. *Ecol. Freshw. Fish.* 10:191-197.
- Sharp, R.P., C.R. Allen, and M.F. Meier. 1959. Pleistocene glaciers on southern California mountains. *Amer. J. Sci.* 257:81-94.
- Taylor, E.B. and J.D. McPhail. 1999. Evolutionary history of an adaptive radiation in species pairs of threespine sticklebacks (*Gasterosteus*): insights from mitochondrial DNA. *Biol. J. Linn. Soc.* 66: 271-291.
- Taylor, E.B. and J.D. McPhail. 2000. Historical contingency and ecological determinism interact to prime speciation in sticklebacks, *Gasterosteus*. *Proc. R. Soc. Lond. B* 267:2375-2384.
- Tedford, R.H., L.B. Albright, III, A.D. Barnosky, I. Ferrusquia-Villafraña, R.M. Hunt, Jr., J.E. Storer, C.C. Swisher, III, M.R. Voorhies, S.D. Webb, and D.P. Whistler. 2004. Mammalian biochronology of the Arikareean through Hemphillian interval. In *Late Cretaceous and Cenozoic Mammals of North America: Biostratigraphy and Geochronology* (M.O. Woodburne, ed.), pp. 169-231. Columbia University Press, New York.
- Vaughan, F.E. 1922. Geology of San Bernardino Mountains north of San Gorgonio Pass. *Bull. Dept. Geol. Sci. Univ. Calif.* 13:319-411.
- Whistler, D. P., 1991. Geologic History of the El Paso Mountains region. In *Inland Southern California: the last 70 million years* (M.O. Woodburne, R.E. Reynolds, and D.P. Whistler, eds.) San Bernardino County Museum Association *Quarterly* 38: 108-114.
- Wilson, M.V.H. 1987. Predation as a source of fish fossils in Eocene lake sediments. *Palaios* 2:497-504.
- Woodburne, M. O., 1991. The Mojave Desert Province. In *Inland Southern California: the last 70 million years* (M.O. Woodburne, R.E. Reynolds, and D.P. Whistler, eds.) San Bernardino County Museum Association *Quarterly* 38: 60-77.
- Wootton, R.J. 1976. *The Biology of the Sticklebacks*. New York, NY: Academic Press.
- Wootton R.J. 1984. *A Functional Biology of Sticklebacks*. Berkeley: University of California Press.

Appendix 1.

The following specimens from the Argonaut site (1.76.43) and deposited in the San Bernardino County Museum (SBCM), Department of Geological Sciences collection were assigned by the curatorial staff to the Gasterosteidae, and identifications were confirmed by the authors.

L1814-1012, one badly worn, probable stickleback spine with broken tip.

L1814-1013, one pelvic or median spine base.

L1814-1014, one left pelvic spine with broken tip.

L1814-1015, one left pelvic spine with broken tip.

L1814-1016, one median spine base.

L1814-1017, one badly worn spine base.

75 years of fieldwork in the Barstow Formation by the Raymond Alf Museum of Paleontology

Don Lofgren¹ and Rajsavi Anand²

¹Raymond Alf Museum of Paleontology, 1175 West Baseline Road, Claremont, CA 91711

²The Webb Schools, 1175 West Baseline Road, Claremont, CA 91711

Introduction

About 40 miles east of Los Angeles and nestled up against the San Gabriel Mountains in Claremont, California, is the Raymond M. Alf Museum of Paleontology. The Alf Museum is located on the campus of The Webb Schools, a private secondary school. The Alf Museum of Paleontology, established in 1936, is the only nationally accredited museum on a high school campus in the United States and the only large paleontology museum in the world that involves secondary school students in all aspects of its operations. The museum was founded by its namesake, Raymond Alf, a math–science teacher who became passionately devoted to the study of paleontology. Alf's first experience with fossils occurred in 1926 when he was a young man of 20 toiling in the fields of Kansas harvesting wheat. On a break from this hot, dusty work, Alf hiked to a nearby hill where outcrops of limestone were exposed and collected fossils of Cretaceous cephalopods and pelecypods. Because of their beauty, Alf kept them with him for the rest of his life. This early simmering interest in paleontology eventually led Alf to collect his first fossils at Barstow with Webb students in the 1930s. Thus began the now 75 year old legacy of Alf Museum crews collecting Miocene (15–20 million years old) fossils from the Barstow Formation.

Early years

If one had to pinpoint where Alf's interest in collecting at Barstow began, it was in 1935 when he was running errands and visited a photo shop in Claremont Village. There on a glass countertop was the display of a beautiful fossilized horse jaw. Alf was enthralled with the specimen and its row of shiny, jet-black teeth and inquired where he might find such a magnificent specimen. He learned that the jaw and other vertebrate fossils could be found in multi-colored sedimentary rocks exposed in an area called Rainbow Basin, about 5 miles northwest of Barstow. So Alf gathered together some Webb students and took his first fossil collecting trip to Barstow a few weeks later.



Figure 1. Raymond Alf (farthest right) and students pose before the trip to Barstow where Bill Webb (3rd from the left) found the new species of peccary. Also pictured is Art Clokey (2nd from the left) who was part of the first summer peccary trip in 1937.

Only a few teeth and bones were found, but the trip marked the beginning of what was to lead to hundreds of fossil collecting trips to Barstow by Alf.

Alf returned with seven students to Barstow in November 1936 (Figure 1) and an important discovery was made that changed his life and led to the founding of the Raymond M. Alf Museum of Paleontology. On that momentous trip, sophomore Bill Webb became bored with looking for fossils after an hour or so and started sliding down the gravelly slopes that occur throughout Rainbow Basin. On one such slope, he tore his jeans as he slid. When he looked uphill to see what caused the damage, Webb spotted a jaw fragment with teeth. He shouted to Alf who came over and they soon traced a trail of bone fragments uphill to their source, a skull that was partially exposed by erosion. Alf and Webb excavated the skull, and in the process, broke open the braincase. Inside, there was a cast of the brain composed of rock. Finding a “fossilized brain” got them very excited. They showed it to all at Webb and some disagreement ensued over whether it was actually a fossilized brain or not. So they decided to take it to Dr. Chester Stock, geology professor at Cal-Tech, a

Progress Bulletin
A NEWSPAPER DEDICATED TO THE INTERESTS OF THE GREATER POMONA VALLEY

Local News, Financial, Club Sports and Building

TWO POMONA, CAL., SATURDAY EVENING, JUNE 19, 1937. EIGHT PAGES PRICE 3 CENTS

Webb School Youth's Discovery Of Skull 20 Million Years Old Wins Recognition of Scientists

Believed to be one of the outstanding scientific discoveries of the year, the finding of a fossilized peccary skull by 11-year-old Billy Webb last November near Barstow has placed the Webb School for Boys north of Pomona, in the limelight of scientific attention, it was learned here today.

Discovery of this fossil supplies a definite link in the evolution of the peccary, the pig family that has inhabited the Western Hemisphere, authorities have revealed. Young Webb is the son of Headmaster and Mrs. Thompson Webb.

Approximately 20 million years old is the age of the fossil, Dr. Chester Stock of the California Institute of Technology, who is regarded as the leading mammalian paleontologist in America, has stated, after a comprehensive study of the object. He places it in the upper Miocene epoch of the tertiary period. The fossil is not only the connecting link in the evolution of that family, but it also gives a more detailed story of the geological and geographical conditions in the Great Basin during that period.

Hunted Fossils on Desert

Accompanied by Raymond M. Alf, biology and mathematics teacher at Webb school, blond Billy Webb was one of a party of school boys enjoying a week-end outing about Thanksgiving time on the desert. Such trips are popular. Alf has planned them to help the boys enjoy outdoor life, with the additional definite purpose of hunting for fossils to add to the Webb school museum.

Barstow was their destination, for it is considered the site of one of the finest fossil deposits in California, and has been worked by the American Museum of Natural History.

In previous excursions, Alf's boys had brought home remains of a three-toed horse and a prehistoric camel.

Fall Results in Discovery

It was a stumble down a hill that caused young Webb to be responsible for furthering scientific knowledge. As he fell, he noticed a portion of the fossil, and he returned to investigate it further. First he found a single tooth; then he uncovered the whole skull with the fossilized brain. They recognized it as a different kind of animal than any with which they were familiar, and they took it to Dr. Stock at Cal Tech.

Cleaning it, the scientist found the almost complete dentition of the animal. A young female it still has the milk dentition, and when some of the milk teeth were removed, the perfectly preserved permanent teeth were revealed. The scientific name given to the peccary by Dr. Stock is *Dysecolyus fricki*. It definitely connects the mammoth with the modern pig, he said.

Findings to be Published

A complete description of the fossil and Dr. Stock's findings will be published by the National Academy of Sciences publication, probably in its next issue. In his article, Dr. Stock credits Billy Webb with the discovery.

Announcement of the importance of the find was made last night at the Webb school "civic night" when the fossil received first prize in the hobby show. Paleontology has been Billy's hobby, and he hopes to continue his study of it when he goes to college in another year or two.




—Photos by Dean Granger

Sixteen-year-old Billy Webb, who found a skull authoritatively declared to be 20 million years old and hailed as one of the year's most important contributions to the science of paleontology, is shown holding the object which has attracted the interest of the scientific world. At left is Raymond M. Alf, Webb School instructor who was in charge of the party of boys on a desert excursion when Billy made his find; at right is Headmaster Thompson Webb, Billy's father. Lower photo is a close-up of the fossilized peccary skull, showing the upper and the detached lower jaw, and the prominent canine tooth or tusk at end of the jaw.

Figure 2. Discovery of the new peccary species generated a front page article in the newspaper; left-right, Raymond Alf, Bill Webb, and Thompson Webb (headmaster of Webb School).

widely recognized expert on fossil mammals, to settle the issue. Stock said they had found a fossilized brain cast, but more importantly, the skull represented a species of fossil peccary new to science. Alf and Webb were very excited and donated the skull and jaw fragments to Cal-Tech (CIT 2039; California Institute of Technology specimen number 2039; CIT 2039 is now part of the collections at the Natural History Museum of Los Angeles County) so Stock could study them. These specimens became

the type of the new genus and species, *Dysecolyus fricki* (Stock 1937). After publication of *D. fricki*, a story in the Pomona Progress Bulletin announced the discovery, complete with a picture of Ray Alf, Bill Webb, and Webb Schools headmaster Thompson Webb (father of Bill), with the peccary skull and jaw fragments (Figure 2). The next trip to Barstow resulted in a large collection of fossil vertebrates. This recent success and the lingering euphoria over the discovery of the new peccary species fired Alf's desire to expand the scope of his search for fossils.

In the summer of 1937, Alf and students Bill Webb and Art Clokey (who later became the filmmaker who created the claymation character Gumby) hatched a plan to go collect fossils at Agate Springs, Nebraska. Alf, Clokey, and Webb called themselves the "Peccary Society." Thereafter, students who went with Alf on collecting trips became members of this Peccary Society and the 1937 summer trip became known as the first Summer Peccary Trip (the tradition of annual summer trips with Webb students continues today). After collecting at Agate Springs, they prospected near Scenic, South Dakota (now Badlands National Park) and Alf ran into John Clark, a paleontologist and professor of geology at the University of

Colorado. Alf was greatly impressed by Clark and decided to become a paleontologist. On his return to Webb, Alf was granted a sabbatical and attended graduate school at the University of Colorado where he studied under Clark and completed a masters of science in geology.

When Alf returned to Webb in 1939 as an academic paleontologist, he was assigned to teach in a large classroom in the lower level of the new Thomas Jackson Library. This became Alf's classroom and museum for the



Figure 3. Raymond Alf's classroom in the basement of the school's library soon overflowed with fossils from the success of Peccary trips.



Figure 4. Student John Tuteur poses with his discovery, a partial skull still in the ground of *Megahippus mckennai* (RAM 910; Raymond Alf Museum of Paleontology specimen number 910).

next 29 years (Figure 3) and the student paleontology program at Webb became a permanent part of the school program. Also, fossil collecting trips were now known as "Peccary Trips." About this time, Alf made the Barstow Formation his main collecting area due to its proximity to Webb and because it was well known as a rich site for fossils. Collecting crews hired by Childs Frick had worked the Barstow area for many years and had collected thousands of fossils which are now housed at the American Museum of Natural History in New York. Alf had run into the Frick's crews on his early trips to Barstow and thus knew the great potential for fossils at Barstow. The discovery by Bill Webb of a new species of peccary in 1936 cemented Alf's interest and he took trips to Barstow on a regular basis until the onset of World War II.

Middle years

After the war ended, collecting trips resumed. By 1953, as reported in the News Bulletin of the Society of Vertebrate Paleontology, Alf was taking trips to Barstow nearly every weekend. The Peccary Society usually camped near the west end of the Rainbow Basin Loop Road and from there they hiked about a mile west to a dig site named Quarry 5 where jaws and teeth of *Merychippus* (grazing type of horse) were common.

The weekend collecting trips yielded many interesting specimens, but Alf got very excited in 1955 when an upper molar of a rare anchitherine or browsing type of horse was found. This single specimen indicated a new type of horse at Barstow was waiting to be discovered. Alf and students searched diligently for more specimens and two years later they hit the jackpot. On a trip to the west end of the Barstow outcrops in an area known as Fullers Earth Canyon, student John Tuteur found a partial skull of this rare horse in the Upper Member of the Barstow Formation that preserved almost the entire upper dentition (Figure 4). The partial skull became the type specimen of a new species, *Megahippus mckennai* (Tedford and Alf 1962), which was named after Alf's former student Malcolm McKenna, a curator at the American Museum of Natural History. This was the first record of *Megahippus* from the Pacific Coast and also the youngest known occurrence of the genus (Tedford and Alf 1962).

Prospecting in the Fullers Earth Canyon area continued for the next few years and in 1959 freshman student Al Korber found a proboscidean jaw



RARE FIND — Albert Korber, student at Webb School, examines the rare fossil he found on a field trip. The top of a 5¼-inch molar is reflected in the mirror. The trip leader, Raymond Alf, stands behind the student. (P-B photo)

Webb Student Finds Rare Fossil On 1st Paleontology Field Trip

Figure 5. Freshman student Al Korber, on his first peccary trip found a proboscidean jaw and tusk (RAM 907) which was featured in a local newspaper.

with tusk (proboscideans are elephant-like mammals with upper and lower tusks). Korber at first thought the tusk sticking from the outcrop was a piece of petrified wood, but Alf quickly identified it as a tusk. Korber had made a very rare find (Figure 5) and this remains the only proboscidean jaw ever found at Barstow.

By the late 1950s Alf expanded his search for fossils at Barstow to tracks and trackways of extinct Miocene vertebrates. This was fueled by successful trips to the Avawatz Mountains north of Baker, California, where the Peccary Society found Miocene tracks of camels and cats in the Avawatz Formation (Alf 1959). Alf and students searched areas of Rainbow Basin and Owl Canyon and in 1960 they found their first Barstow tracks. Later that year, Webb faculty member Lach McDonald spotted an elongate track that seemed to represent a large animal dragging its claws in mud within a sequence of steeply

dipping strata in Owl Canyon. Alf identified it as a large bear-dog, almost certainly that of *Amiphicyon*, a tiger-sized carnivore whose bones were known from the Barstow Formation, but were rarely found. Excavations revealed a set of multiple tracks of the bear-dog (Figure 6) and in 1964 Alf decided to collect the trackway for future display. The trackway was sawed into sections and taken back to the museum, to be reassembled at some later date when the proper display space was available. These bear-dog tracks and others representing cats and camels were the first track specimens described from the Barstow Formation (Alf 1966).

A growing problem with the collections made by the Peccary Society was the issue of proper storage and display space. There just wasn't enough available in Alf's classroom-museum in Jackson Library, so fossils were stored all over campus. For example, the bear-dog trackway was stored in a corner of a Quonset hut used by Webb maintenance staff. It was obvious that a new museum building was needed to house the collections and



Figure 6. The very rare bear-dog or *Amiphicyon* trackway (RAM 100) being measured before it was cut and removed. It was reassembled six years later in the museum's Hall of Footprints.



Figure 7. To collect the proboscidean trackway (RAM 187), the track slab had to be cut into pieces. Al Korber, who returned to Webb after college to teach, is manning the gas powered saw.

provide space for large exhibits like the bear-dog trackway. A fundraising drive was launched which proved very successful. The result was the opening of a new two story museum in 1967 with 8,000 square feet of exhibit space and two large rooms to house collections. At the dedication ceremony in 1968, the museum was formally named the Raymond M. Alf Museum after its founder. Above the front entrance, local artist Millard Sheets designed a mosaic of a peccary with flames projecting from its body, which signified “the flames of enthusiasm,” in reference to the passionate efforts of the Peccary Society in creating this museum. Alf was extremely proud of the new museum and referred to it as “a living memorial to what students and teachers can accomplish together.”

With the new museum came the task of deciding what to display and how to organize the exhibits. Alf’s vision was to have two themes, a Hall of Life, with exhibits showing the history of life on Earth, and a Hall of Footprints, with displays centered on the museum’s unique track collection. Displays for the Hall of Life were constructed in 1967–1968; those of the Hall of Footprints were built in 1970.

In 1968, about the time the new museum was dedicated, the Peccary Society found a set of four proboscidean tracks in a canyon a mile east of Owl Canyon. The trackway was very large and had to be sawed into many pieces before it could be safely removed (Figure 7). Once the trackway layer was cut, each piece had to be pried loose using metal shims (Figure 8) and then loaded on to a stretcher and carried to a nearby vehicle. It was a difficult process because the rock was not well cemented and would easily break into fragments, so it took a few trips to complete the removal of the 16 foot long trackway, a process that extended into 1969. Unlike the bear-dog trackway that was stored for years before it was reassem-



Figure 8. Metal shims are wedged under the track slab after it was cut into pieces, so each piece could be removed intact and transported to the museum.

bled in the Hall of Footprints, the spectacular proboscidean trackway was put on display only a few months after it



Figure 9. The camel trackway (RAM 166) on display in the Hall of Footprints shows two sets of parallel tracks.

was removed from the field. Another large trackway from Barstow that preserves tracks of two camels walking side by side is also displayed in the Hall of Footprints (Figure 9), as are single tracks of cats and camels.

After Alf's retirement in the 1970s, Alf Museum crews visited the Barstow area less frequently and few scientifically significant fossils were collected. One important specimen found in the 1970s was a partial proboscidean skull (RAM 908) from a now undocumented site east of Owl Canyon. This specimen represents the only proboscidean skull ever found in the Barstow Formation.

Recent years

Renewed collecting efforts at Barstow were initiated by Don Lofgren when he became museum director in 1991. Working under a state BLM permit, with field authorization from the Barstow BLM Office, Lofgren's goal was three-fold: 1) locate and document Alf's sites, especially the trackway localities; 2) surface collect all museum quality fossils found eroding from outcrop to preserve them for future generations; 3) provide Webb students with a high quality educational experience where they could collect and preserve fossils from the area in accordance with federal repository regulations.

In the early days of collecting at Barstow, maps were scarce and field notes were rarely taken so precise locality descriptions and map plots for many Barstow specimens did not exist. If fossils are undocumented, their scientific value is lessened. Fortunately, there were many excellent photographs of Barstow sites in the museum's archives that could be matched with specimens. Thus, Lofgren was able to relocate and document most of the Museum's important Barstow localities, including the bear-dog and proboscidean trackway sites (Lofgren et al., 2006). The museum's track collection was inventoried and cataloged and there were 322 specimens of tracks or trackways from the Barstow Formation (Lofgren et al. 2006). Of these, 295 are camel and 22 are felid; the remainder are canid, amphicyonid or proboscidean.

Surprisingly, the Alf Museum's large and important collection of Miocene tracks from Barstow was not studied until recently. This changed when William Sarjeant from the University of Saskatchewan and Robert Reynolds from the San Bernardino County Museum visited the museum and studied the Barstow track collection in the late 1990s. Two papers were published that described and named the Barstow camel and carnivore tracks (Sarjeant and Reynolds 1999; Sarjeant et. al 2002). Seven new ichnotaxa were recognized. These holotype, syntype, or paratypes specimens are:

RAM 182 (syntype): *Lamaichnum alfi* (camel track)

RAM 159 (syntype): *Lamaichnum alfi* (camel track)

RAM 166 (paratype): *Lamaichnum alfi* (camel trackway; Figure 9)

RAM 100 (holotype): *Hirpexipes alfi* (bear-dog trackway; Figure 6)

RAM 103 (holotype): *Felipeda bottjeri* (cat track)

RAM 104 (paratype): *Felipeda bottjeri* (cat track)

RAM 105 (paratype): *Felipeda scrivneri* (cat track)

Over the past two decades, Webb students and Alf Museum staff visited the Barstow Formation about seven times a year on average. These educational trips (often a required overnight field trip for a Webb paleontology course) are excellent learning opportunities for Webb students and also result in the collection of many important specimens. For example, a trip in 2004 located an



Figure 10, Freshman students on their first peccary trip plaster the camel skeleton (RAM 7648) found in 2004.



Figure 11. The camel skeleton after preparation, still in its plaster jacket to give it protection and support.

articulated camel skeleton from the Upper Member of the Barstow Formation. Erosion had removed the head and parts of the front legs, but otherwise the skeleton was completely intact, a very rare discovery. It took four weekend trips to encase the skeleton in a protective plaster jacket (Figure 10) and remove it. Then it took 200 hours to prepare the skeleton which was identified as the medium-sized camel, *Protolabis barstowensis*. The skeleton was left within its plaster jacket to protect and stabilize the specimen for exhibit in the Hall of Life (Figure 11). In 2008, Webb alumnus John Enders found a nearly complete lower molar of a proboscidean in Fullers Earth Canyon. This discovery shows that important specimens continue to be eroded from Barstow outcrops and that it is critical that federal repositories like the Alf Museum be permitted to collect these specimens before they are destroyed by erosion or removed by vandals. As of January 2009, the Alf Museum's Barstow collections numbered approximately 6,500 specimens.

Summary

For seventy-five years, Alf Museum crews have prospected the canyons northwest of Barstow in search of museum-quality fossils. Thousands of specimens, from the huge proboscidean trackway to small rodent teeth, have been collected from the Barstow Formation. These scientifically significant collections are the legacy of the work of Webb students and faculty and Alf Museum staff, especially museum founder Raymond Alf. Many of the Barstow specimens are on display in the museum's

Hall of Footprints and the Hall of Life so all can enjoy their beauty and appreciate their role in elucidating the history of life. The Alf Museum's Barstow collections are also available for study and can be accessed by qualified investigators for research purposes. The mission of the Alf Museum is to build collections that preserve the history of life and to promote knowledge of the scientific method using paleontology as the vehicle. The museum's long-term Barstow efforts are a prime example that the mission is being fulfilled.

Acknowledgements

We thank the hundreds of Webb faculty, staff, students, alumni and others who comprise the Peccary Society for their efforts in amassing the Alf Museum's large collection of Barstow Formation fossils. Also, many thanks to the many staff at both the California State BLM Office in Sacramento and the Barstow District BLM Field Office for issuing collecting permits to Alf Museum crews over the past few decades.

References cited

- Alf, R.M. 1959. Mammal footprints from the Avawatz Formation, California. *Bulletin of the Southern California Academy of Sciences* 58:1-7.
- Alf, R. M. 1966. Mammal trackways from the Barstow Formation, California. *Bulletin of the Southern California Academy of Sciences* 65:258-264.
- Lofgren, D. L., J.A. Greening, C.F. Johnson, S.J. Lewis, and M.A. Torres. 2006. Fossil tracks at the Raymond Alf Museum of Paleontology and management of tracks on public lands. *New Mexico Museum of Natural History and Science Bulletin* 34:109-118.

- Sarjeant, W. A.S. and R.E. Reynolds. 1999. Camel and horse footprints from the Miocene of California. Abstracts from the 1999 Desert Research Symposium, San Bernardino County Museum Association Quarterly 46:3-19.
- Sarjeant, W. A.S., R.E. Reynolds, and M. Kissell-Jones. 2002. Fossil creodont and carnivore footprints from California, Nevada, and Wyoming, p. 37-50 in Reynolds, R.E. (ed), Between the basins: exploring the western Mojave and southern Basin and Range Province. California State University, Desert Studies Consortium 2002.
- Stock, C. 1937. A peccary skull from the Barstow Miocene, California. Proceedings from the National Academy of Sciences 23:398-404.
- Tedford, R. H. and R. M. Alf. 1962. A new *Megahippus* from the Barstow Formation San Bernardino County, California. Southern California Academy of Sciences 61:113-122.

Paleomagnetism of Miocene volcanic rocks in the Newberry Mountains, California: vertical-axis rotation and a polarity transition

John W. Hillhouse¹, Ray E. Wells², and Brett F. Cox³

U.S. Geological Survey, 345 Middlefield Rd., MS 975, Menlo Park, CA 94025, United States

¹650 329 4932; jhillhouse@usgs.gov, ²650 329 4933; rwells@usgs.gov, ³650 329 4956; bcox@usgs.gov

Abstract

Paleomagnetism of volcanic rocks in the central Mojave Desert indicates that large fault-bounded blocks rotated clockwise in Early Miocene time (18.8–21 Ma), perhaps as a consequence of regional extension. Improved geologic mapping in the Newberry Mountains, California, has afforded tight stratigraphic control for our paleomagnetic study, allowing us to calculate a paleomagnetic pole that represents the axial dipole field. Our study of the volcanic rocks, which span at least two polarity zones (>22–21.3 Ma), reveals several large swings in paleodeclination and a transition to normal polarity. Paleomagnetic data from basalts define a polarity transition with a pole path that passes through eastern South America, where previous studies show a preferred clustering of transitional poles. After excluding these transitional directions from the 55 flows sampled, we used 21 independent readings of the ancient field to define the paleomagnetic pole (Lat.= -60.9°; Long.= 151.4°E, α_{95} = 10.0°). We infer a clockwise vertical-axis rotation of $42.7^\circ \pm 10.3^\circ$ by comparing the Newberry paleomagnetic pole with central North America data. The rotation is substantial, but 30° less than the previously reported rotation of 72.7°. The rotation affects rocks as young as ~21 Ma, but the unconformably overlying Peach Spring Tuff (18.8 Ma) shows little or no rotation. We conclude that the rotation occurred during an early Miocene extensional event and is not the result of Pliocene or younger strike-slip faulting in the region. Our study shows that knowledge of the stratigraphic ordering of paleomagnetic directions is essential to avoid bias in rotation estimates introduced by excursions, polarity transitions, and duplicate readings of the ancient field.

Introduction

We investigated Cenozoic tectonic rotations in the central Mojave Desert of California with a stratigraphically controlled sequence of paleomagnetic directions from lower Miocene volcanic rocks in the Newberry Mountains. Our data complement previous paleomagnetic studies designed to test tectonic models that relate regional strain to patterns of faulting in this part of the boundary zone between the Pacific and North American plates [e.g., Ross et al., 1989; Schermer et al., 1996 and references therein]. Although previous work suggests that the central Mojave region rotated approximately 50° about a vertical axis during the interval of 23–18.5 Ma [Ross et al., 1989; Wells and Hillhouse, 1989], apparent inconsistencies in the sense or amounts of rotation measured in nearby areas

have raised doubts about the regional significance and reliability of the paleomagnetic data. Part of the inconsistency might be due to insufficient sampling of natural geomagnetic variation; before mean paleodeclination can be judged to be a valid indicator of vertical-axis rotation, enough sampling must be done to ensure that the paleomagnetic pole is a true representation of the ancient geographic pole [Hillhouse, 1989]. Alternatively, some of the variation in paleomagnetic data between adjoining areas might reflect true variability of rotations in space and time. As more detailed geologic mapping and geochronologic studies become available from the Mojave Desert, a better integration of the geologic structure with paleomagnetic data will likely emerge, and apparent rota-

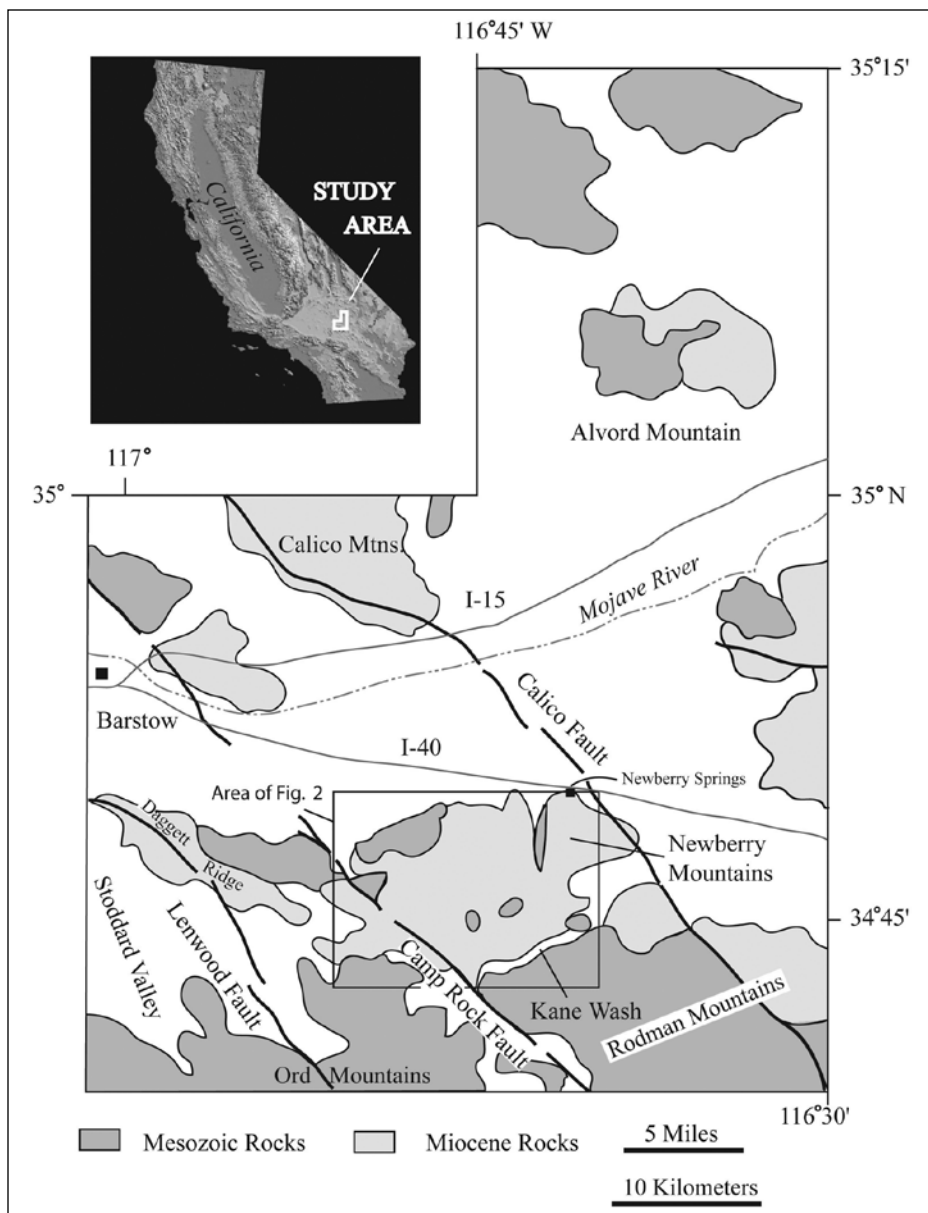


Figure 1. Location of Newberry Mountains study area in southern California. Generalized geologic map modified from Bortugno and Spittler [1986].

tions that are now quite puzzling quite possibly will be explained by newly discovered geologic structures.

Paleomagnetic studies in the Newberry Mountains and surrounding areas (Figure 1) provided a means to test models of large-scale deformation resulting from Quaternary faulting in the central Mojave Desert. By interpreting the offsets of all of the major dextral faults in the region, Garfunkel [1974] inferred that blocks, such as the Newberry Mountains, may have rotated counterclockwise as much as 30°. His hypothesis inspired a paleomagnetic test by Burke et al. [1982], which seemed to confirm a counterclockwise vertical-axis rotation. Two nearly concurrent paleomagnetic studies subsequently showed that tectonic rotation in this region is more likely to be related

to early Miocene extensional faulting, as opposed to the later and more conspicuous episode of strike-slip faulting. These studies also cast doubt on the counterclockwise rotation suggested by paleomagnetic data of Burke et al. [1982]. First, Ross et al. [1989] found evidence for large clockwise rotation (25° to 75° about a vertical axis) of the lower Miocene volcanic rocks in the Newberry and Rodman Mountains. Second, Wells and Hillhouse [1989] showed that the rotation occurred before deposition of the regionally extensive, 19-m.y.-old Peach Spring Tuff. This ash-flow unit is nearly flat-lying in the Newberry Mountains and it overlies the tilted and rotated Miocene volcanic strata. Paleomagnetic declination of the Peach Spring Tuff in the Newberry Mountains is nearly identical to that of equivalent deposits on the Colorado Plateau.

The observations of large-scale clockwise rotation in the central Mojave region have been ascribed to dextral shearing that either accompanied or immediately followed early Miocene crustal extension [Bartley and Glazner, 1991;

Dokka and Ross, 1995; Valentine et al., 1993; Ross, 1995]. Working in the Waterman Hills north of Barstow, Glazner et al. [1989] found evidence of crustal extension in the core-complex style known from the Colorado River corridor [also see: Dokka and Woodburne, 1986; Dokka et al., 1988; Dokka, 1989; Glazner et al., 1988; Walker et al., 1990]. They described the "Waterman Hills detachment fault," and proposed that it accommodated several tens of kilometers of extension during the early Miocene. Dokka [1986, 1989] had previously applied the core-complex paradigm to the Newberry Mountains south of Barstow, where he described the "Newberry Mountains detachment fault." Decoupling of the crust across a shallow low-angle fault during regional extension provides a

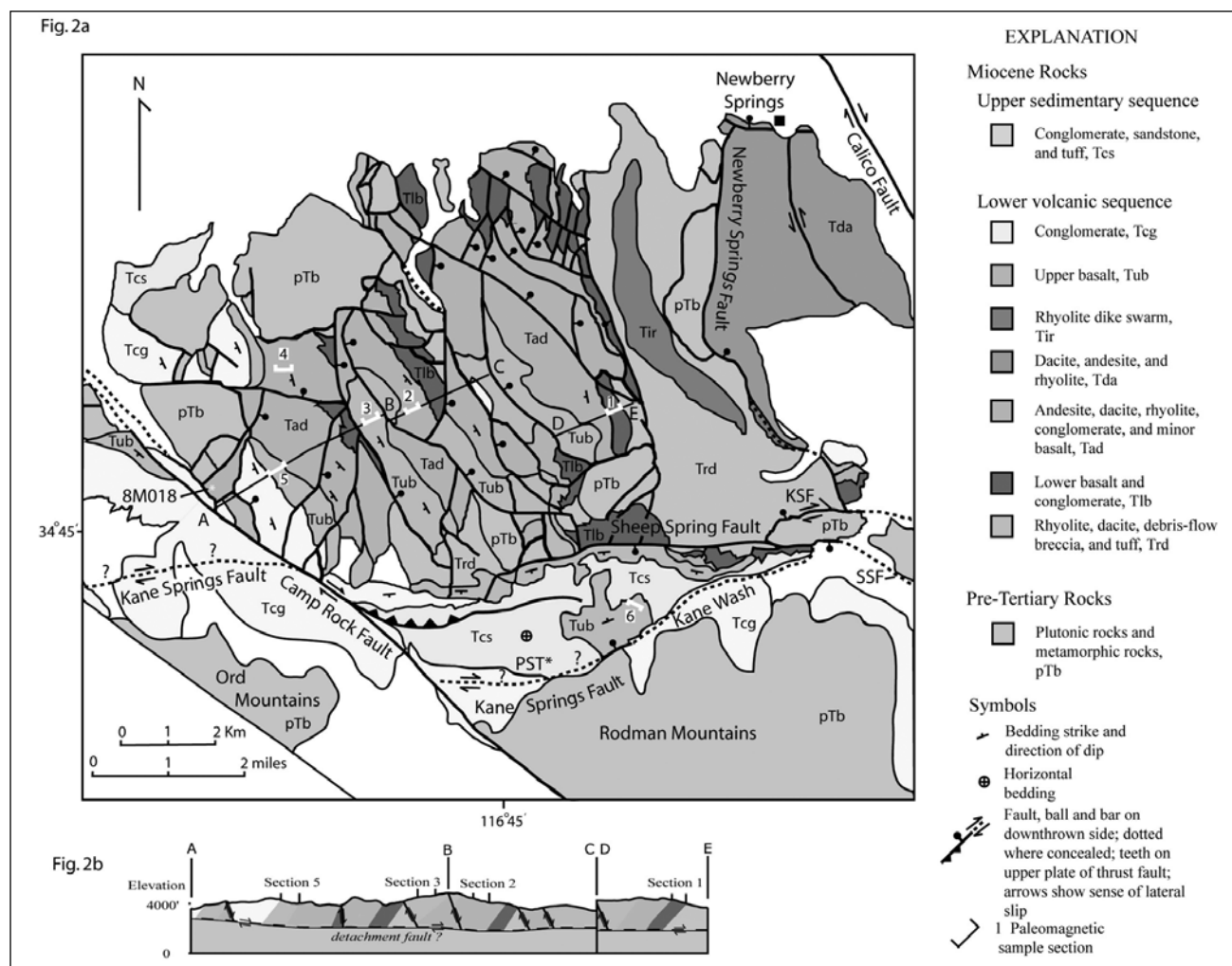


Figure 2. a) Geologic map (modified from Cox, in press) and locations of sections sampled for paleomagnetism in the Newberry Mountains. KSF, Kane Springs fault; SSF, Sheep Spring fault. 8M018, location of last site listed in Table 1. PST, paleomagnetic sample site in Peach Springs Tuff [Wells and Hillhouse, 1989]. b) Schematic cross section showing structure of Miocene strata; location of cross section and rock-unit explanations shown in part a. Vertical exaggeration approximately $\times 2.2$.

convenient way to accommodate the large vertical-axis rotations that Ross et al. [1989] and Ross [1995] inferred from the paleodeclination data. However, Glazner et al. [2002] disputed the existence of an exposed detachment fault in the Newberry Mountains, and proposed that highly extended terrane, as characterized by structures in the Waterman Hills, is not present there. Instead of detachment faulting, more deeply-seated dextral shearing or oroclinal bending along the southern boundary of the Waterman Hills extended terrane may have induced clockwise rotation of the Newberry Mountains and neighboring areas, either during early Miocene extension [Bartley and Glazner, 1991; Glazner et al., 2002], or perhaps during a post-extensional tectonic event [Dokka and Ross, 1995; Dokka et al., 1998].

The opportunity to further refine the paleomagnetic evidence for rotation arose when a new, more detailed geologic map of the Newberry Mountains became

available [Cox, in press]. With new stratigraphic and structural control, we designed a paleomagnetic study to cover much of the stratigraphic section and provide two different structural domains in the Newberry Mountains. Rocks spanning a long time interval must be sampled to fully average geomagnetic secular variation. Moreover, sampling sites in known stratigraphic order allows identification of extreme directional changes in the geomagnetic field, especially polarity transitions, and allows recognition of duplicate readings of the field. This sampling strategy produces a more reliable measure of tectonic rotation than is obtained by sampling scattered outcrops in unknown stratigraphic order. The two structural domains of interest are separated by a south-facing monocline and adjacent zone of east-striking faults on the south flank of the Newberry Mountains just north of Kane Wash (Figure 2). In the northern domain, the Tertiary volcanic beds generally dip to the southwest and

the stratigraphic section is repeated by many east-dipping normal faults. In the southern domain, the volcanic rocks alternately dip northward and southward owing to deformation by east-striking reverse faults and associated folds. We sampled both domains to look for evidence of differential rotation approaching the southern margin of the Tertiary outcrop belt.

An unexpected outcome of our study was the discovery of a polarity transition from a group of nine basalt flows that captured the field changing to normal magnetic polarity. Polarity transitions are rarely found, because the phenomenon occurs too rapidly to be preserved in volcanic rocks except during periods of frequent volcanic eruptions. While paleomagnetic data from transitions should be discounted in rotation calculations, new transition records provide important clues to the working of the geomagnetic dynamo. Recent debate about the polarity-transition process centers on whether or not the geomagnetic dip pole (as calculated from the observation site under the geocentric, axial dipole assumption) crosses the equator at preferred regions. If preferred crossings recur for a long period of time, then the lower mantle might be influencing the reversal process. This study adds information to the debate.

Geologic setting

The Newberry Mountains are located south of Interstate Highway 40 about 25 kilometers east-southeast of Barstow, California (Figure 1). The ridges rise about 750 m above an alluvial piedmont on the north flank of the mountains. Several large canyons give access to well-exposed strata in the center of the range. The local geology was mapped at 1:62,500 scale by Dibblee [1964a, b; 1970], Dibblee and Bassett [1966], and Dokka [1980]. The Cenozoic geologic history of the range was discussed in regional syntheses by Dokka [1980; 1983; 1986; 1989], Dokka and Glazner [1982], and Dokka and Woodburne [1986]. Cox et al. [1987] studied the mineral-resource potential of the Newberry Mountains and neighboring Rodman Mountains. The generalized stratigraphic succession of the Newberry Mountains revealed by these previous studies consists of: 1, granitic and metamorphic basement rocks of pre-Tertiary (mainly Mesozoic) age; 2, a lower volcanic sequence dominated by dacite, andesite, and basalt, all of early Miocene age; and 3, an upper sedimentary sequence mainly consisting of conglomerate, breccia, pebbly sandstone, and minor air-fall tuff, also of Miocene age.

Detailed stratigraphic and structural control for the present study is derived from a recently completed 1:24,000-scale geologic map of the Newberry Mountains [Cox, in press], a generalized version of which is shown

in Figure 2. The Miocene lower volcanic sequence constitutes the main body of the Newberry Mountains. The basal part of this sequence mainly consists of volcanic debris-flow breccia and flows of rhyolite and dacite (Figure 2, unit Trd). Our paleomagnetic study focuses on higher parts of the volcanic sequence, which are more heterogeneous, consisting of basalt, andesite, dacite, and rhyolite flows, interbedded with sedimentary breccia and conglomerate (Figure 2, units Tlb, Tad, Tub, and Tcg). The lithologic assemblage in the upper part of the volcanic sequence is most diverse in the central and western parts of the range. In contrast, an apparently correlative succession at the northeast end of the range (Figure 2, unit Tda) consists mostly of dacite and other silicic to intermediate volcanic rocks and contains very little basalt, conglomerate and sedimentary breccia.

Isotopic ages from volcanic strata and associated dikes suggest that the lower volcanic sequence ranges from about 23 Ma to 20 Ma. Volcanic debris-flow breccia and air-fall tuff in the basal part of the sequence yielded several $^{40}\text{Ar}/^{39}\text{Ar}$ ages ranging between 23.0 Ma and 22.7 Ma [Cox, in press]. Nason et al. [1979] reported a whole-rock potassium-argon age of 23.7 ± 2.4 Ma (age revised using modern decay constants) from the lowest mapped unit of basalt on the north flank of the range (Figure 2, unit Tlb). An andesite flow higher in the volcanic succession at the west end of the range (Figure 2, locality 8M018) was dated at 19.7 ± 0.6 Ma by potassium-argon on whole-rock material [Janet Morton, written communication, 1983]. Other evidence, cited below, suggests that this date may underestimate the true age of the andesite by at least 1 m.y. A wide, northwest trending swarm of rhyolite dikes in the eastern part of the range (Figure 2, unit Tir) is the likely source of rhyolite flows that locally overlie the dated andesite flow at the top of unit Tad. Potassium-argon and $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 21.4 to 20.5 Ma were obtained from the rhyolite dike swarm [Cox, in press]. An outlying rhyolite dike at the southeast end of the range, which intrudes the Newberry Springs fault and adjoining basalt flows (Figure 2, unit Tub?) yielded $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 21.17 ± 0.13 Ma and 21.33 ± 0.23 Ma on biotite and sanidine, respectively; this suggests that the upper basalt unit may be older than about 21.2 Ma [Cox, in press]. The uppermost part of the lower volcanic sequence consists of a unit of conglomerate and sedimentary breccia (Figure 2, unit Tcg). Ostensibly correlative deposits of conglomerate and breccia to the west on Daggett Ridge (Figure 1) envelop an andesite flow that was dated at 20.2 ± 0.3 Ma by the whole-rock K-Ar method [Dokka, 1989].

The upper sedimentary sequence occupies a narrow, east-trending depositional basin on the south flank of the Newberry Mountains. The basin fill consists of locally-derived conglomerate, breccia, and sandstone that overlie the lower volcanic sequence with pronounced angular unconformity. Intercalated within this clastic sequence is a distinctive layer of ash-flow tuff about 10 meters thick. The ash-flow tuff was originally dated at 21.0 ± 1.6 Ma by the fission-track method [Dokka, 1986]. However, the tuff was later determined to be a distal lobe of the Peach Springs Tuff of Young and Brennan [1974] on the basis of petrology [Glazner et al., 1986] and paleomagnetism [Wells and Hillhouse, 1989]. The Peach Spring Tuff, which forms prominent cliffs far to the east near Kingman, Arizona, yielded an average age of 18.5 ± 0.2 Ma from laser-fusion and incremental-heating $^{40}\text{Ar}/^{39}\text{Ar}$ methods [Nielson et al., 1990].

The Newberry Mountains consist of two structural domains separated by a south-dipping normal fault (Sheep Spring fault) and adjacent south-facing monocline (Figure 2). The northern domain is broken into many elongate blocks bounded by normal faults that generally dip moderately to steeply (40° to 80°) to the east. Many faults are arcuate, marked by concave-east traces on the map. Strata in the fault blocks on average dip west about 40° , but the dips shallow upward conspicuously in some blocks, ranging from as steep as 80° near the base of the succession to as shallow as 15° near the top. Owing to large dip separations on the order of 500–1500 m on the east-dipping normal faults, the central and upper parts of the Miocene volcanic succession are repeated several times as the range is traversed from east to west. One of the most prominent normal faults, here named the Newberry Springs fault, juxtaposes the uppermost parts of the lower Miocene succession against pre-Tertiary basement rocks and basal parts of the Miocene succession near the east end of the range. The normal faults typically also exhibit a component of left-lateral slip and consist of progressively older northwest-, north-, and northeast-striking sets; these relations suggest that normal faulting occurred simultaneously with clockwise rotation of the Newberry Mountains [Cox, in press]. Some faults deviate from the typical characteristics described above. A prominent north-northwest-striking fault with antithetic, down-to-west displacement cuts across the east-central part of the range. In a few places in the eastern and western parts of the range, steeply dipping strata in the lower part of the Miocene succession are cut by gently dipping normal faults that likely originated as high-angle normal faults and were later rotated to a low angle during tilting of the stratigraphic section [Dokka, 1989; Cox, in press; see Glazner et al., 2002, for a contrary view].

Although the steeply dipping normal faults clearly accommodated extension at shallow crustal levels near the land surface, the mechanism of extension at greater depths in the northern domain is uncertain. The steep normal faults may terminate downward against a gently dipping detachment fault [Dokka, 1986; 1989, 1993], as shown in Figure 2b. However, results of recent field studies suggest there are no detachment faults presently exposed at the land surface in the Newberry Mountains [Cox, 1995, 2005, in press; Glazner et al., 2002]. The schematic subsurface detachment fault shown in the cross section of Figure 2b is hypothetical; if such a feature exists beneath the Newberry Mountains, its true depth, dip, and geometric relations with hanging-wall blocks may differ considerably from what is shown.

The southern structural domain is a narrow, east-trending belt of faults and folds that crosses the southern Newberry Mountains directly north of Kane Wash (Figure 2). Recent studies suggest that this belt is a complex product of extensional and contractional deformation, but its precise role in the tectonic development of the Newberry Mountains and surrounding areas is not fully understood [Dokka, 1986; Bartley et al., 1990; Cox, 2005, in press]. Dokka [1980, 1986, 1989] proposed that the Newberry Mountains and adjacent areas to the east and west are bounded to the south by a major dextral fault that he named the Kane Springs fault. According to his interpretation, the Kane Springs fault is physically linked to a subhorizontal detachment fault underlying the Newberry Mountains and neighboring ranges, and the two faults jointly accommodated large differential extension of upper-crustal rocks in these mountains with respect to essentially unextended rocks in areas to the south. Glazner et al. [2000, 2002] argued that an ostensibly equivalent high-angle fault in neighboring mountains to the east (Silver Bell fault in Box Canyon area of northern Rodman Mountains) is a north-dipping normal fault with nil or minimal lateral slip; partly on these grounds, they concluded that northeast-directed early Miocene regional extension produced only very limited lateral displacement in the Newberry Mountains and adjacent areas to the east.

New geologic mapping [Cox, 2005, in press] clarifies the location and displacement of the Kane Springs fault in the Newberry Mountains area. Well-exposed north-dipping fault strands that are reasonably equated with the Kane Springs fault of Dokka [1986, 1989] were identified in two areas north of Kane Wash, in the southern and southeastern parts of the range. These two fault strands are now offset from one another owing to down-to-south displacement of the Sheep Spring fault, which obliquely intersects the Kane Springs fault (Figure 2). The Kane Springs fault apparently displaces pre-Tertiary rocks and

Miocene rocks by different amounts in the dextral sense, suggesting that the fault was active during regional extension [Cox, 1995, in press].

The locally pronounced shallowing of dips upward through the Miocene stratigraphic succession implies that regional extension and associated normal faulting were broadly contemporaneous with volcanism and sedimentation that began about 23 Ma in the Newberry Mountains [Dokka, 1986, 1989; Cox, 2005, in press]. The previously mentioned isotopically-dated rhyolite intruding the Newberry Springs fault near the southeast end of the range confirms that substantial amounts of normal faulting had occurred by 21 Ma [Cox, in press]. Relatively undisturbed, mostly flat-lying sedimentary strata at the top of the lower Miocene succession (the conglomerate, sandstone, and tuff unit, as depicted on Figure 2) record the end of extension and onset of a new paleogeographic and tectonic regime in the Newberry Mountains by about 18.5 Ma. The new tectonic regime included a pulse of north-south contraction and monoclinical arching that uplifted the ancestral Newberry Mountains and triggered deposition of locally derived coarse-grained volcanoclastic sediments in a new basin on the south flank of the range [Bartley et al., 1990; Cox, 2005, in press].

The Newberry Mountains are bounded by the prominent northwest-striking Camp Rock fault and Calico fault. These faults have cumulative right-lateral displacements of about 4 and 10 km, respectively [Garfunkel, 1974; Dokka, 1983; Cox et al., 1987; Jachens et al., 2002]. Their activity likely began during or following the latest Miocene [Dokka and Travis, 1990a] and continues to the present. The faults are part of a family of northwest-striking late Cenozoic dextral faults in the Mojave Desert that are related to the subparallel San Andreas fault [Dibblee, 1961]. Dokka and Travis [1990a] designated this family of faults as key features of the "Eastern California Shear Zone", which currently accommodates as much as 25% of the dextral shearing generated along the Pacific-North American plate boundary (Dokka and Travis, 1990b; Sauber et al., 1994).

Paleomagnetic methods

We collected one-inch diameter cores from sites in the lower volcanic sequence, starting with the lower basalt and conglomerate unit and ending near the top of the upper basalt unit (Figure 2). Each sampling site is inferred to be a single cooling unit, defined as a volcanic flow with distinct upper and lower contacts. The sites are grouped into six

stratigraphic sections (Figures 1 and 2): Section 1, units T1b and Tad, total thickness 220 m; Section 2, unit Tad, 210 m; Section 3, unit Tub, 75 m; Section 4, unit Tad, 100 m; Section 5, unit Tub, 335 m; Section 6, unit Tub, 210 m. The cumulative thickness of rock sampled, combining all sections and excluding overlapping intervals, is about 765 m. We usually collected 6 oriented cores, about one meter apart laterally, from each site. Due to a limited supply of drilling water, we collected only three cores per site in the lower basalt (Section 1). Whenever possible, we measured azimuths with a sun compass; in a few instances we used a magnetic compass only.

In the laboratory, magnetic remanence was measured in a superconducting magnetometer with three orthogonal pick-up loops. Partial demagnetization in alternating fields was used to remove viscous remanent magnetization, lightning effects, and other spurious magnetizations. The demagnetization treatments involved subjecting one specimen from each site to a 10-step sequence of alternating fields up to 100 mT. The apparatus was a shielded 400-Hz coil equipped with a reciprocating tumbler. From the results of the pilot specimen, we selected a range of 4 demagnetization steps to treat the remaining specimens from each site. Finally, the demagnetization data were examined on vector diagrams to ensure separation of the characteristic magnetization, which was calculated by the line-fitting method of Kirschvink [1980].

Bedding attitudes measured are consistent with the geologic mapping and are locally confirmed by the dip of sedimentary interbeds (Figure 3). After calculating the mean strike and dip of the observations, a bedding tilt correction was assigned to each stratigraphic section. We corrected the cleaned paleomagnetic data for tilt by simply rotating the magnetic vector about the strike by the



Figure 3. Flows and interbedded fluvial sediment in the upper basalt unit Tub, section 5. Length of rock hammer is 50 cm.

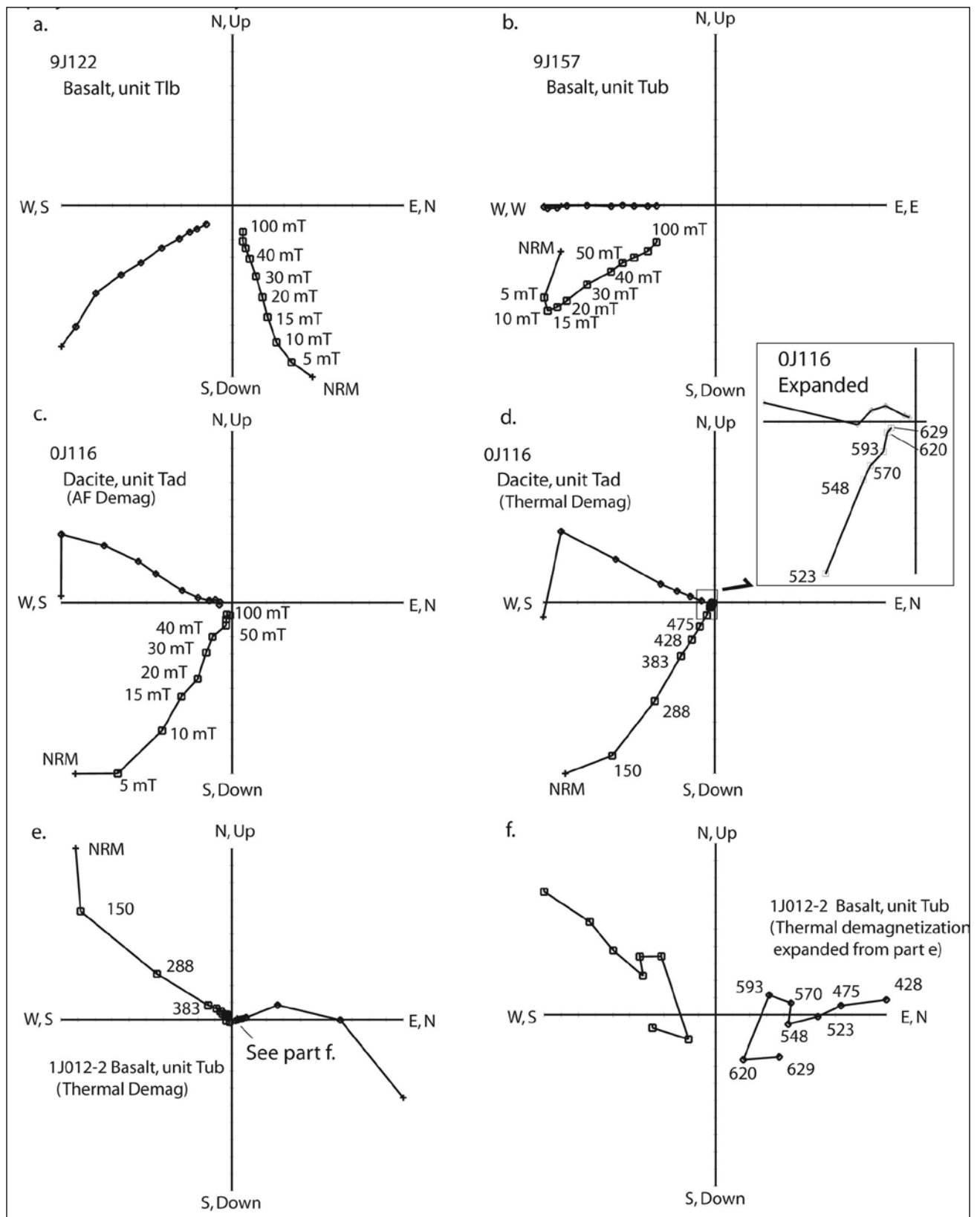


Figure 4. Examples of alternating-field and thermal demagnetization of specimens from units T1b (9J122: Section 1), Tub (9J157: Section 3; 1J012: Section 6) and Tad (OJ116: Section 2). The magnetization vector (not corrected for tilt) is depicted as projections into the vertical (diamonds) and horizontal (squares) planes. Temperatures of thermal demagnetization are given in degrees C. Tick marks are 10% of projected NRM intensity.

Table 1. Paleomagnetic data from Miocene volcanic rocks in the Newberry Mountains, California.

Section	Rock Unit	USGS #	Lat	Long	Str	Dip	D	I_c	D_c	NIN_c	K	α_{95}	λ	ϕ	
Section 1	Lower basalt	9.1121	34.772	243.277	140	38	163.8	11.3	177.1	3/3	107	12.0	-49.4	247.7	
		9.1124	"	"	159	"	167.5	1.1	170.9	3/3	196	8.8	-53.7	258.8	
		9.1127	34.772	243.276	"	"	166.2	13.5	179.9	3/3	6	56.6	-48.4	243.4	
		9.1130	"	"	"	"	152.4	21.1	169.1	3/3	126	11.0	-43.2	258.1	
		9.1133	"	"	"	"	153.5	17.9	167.5	3/3	126	11.0	-44.5	260.7	
Section 2	Dacite	9.1136	"	"	"	"	149.6	17.6	161.9	3/3	322	6.9	-44.0	268.0	
		9.1139	"	"	"	"	138.7	27.4	156.2	3/3	249	7.8	-35.8	272.0	
		9.1142	"	"	"	"	130.0	29.1	145.2	3/3	3662	2.0	-29.8	282.6	
		9.1145	34.771	243.275	"	"	17.5	34.7	160	3/3	195	8.8	75.1	353.4	
		9.1148	34.771	243.274	"	"	33.4	54.9	57.3	2/2	14	74.0	28.2	350.8	
		0.1110 ²	34.772	243.228	"	"	24.1	234.8	-24.1	233.9	5/6	52	10.7	-36.6	163.9
		0.1116 ²	34.773	243.229	"	"	-24.9	209.4	-50.5	183.6	6/6	119	6.2	-85.4	202.0
		0.1122	"	"	"	"	-68.0	311.3	-55.3	35.2	5/6	21	17.1	12.1	34.7
		0.1128 ²	34.774	243.230	"	"	-29.1	224.9	-62.3	198.8	6/6	1658	1.6	-73.0	116.4
		0.1134 ²	"	"	"	"	-16.9	227.1	-52.5	214.3	6/6	89	7.1	-62.6	147.1
Section 3	Upper basalt	9.1150	34.771	243.222	160	48	229.5	-36.7	224.3	6/6	610	2.7	-48.6	161.5	
		9.1156	"	"	"	"	2.3	234.5	-43.6	228.3	6/6	547	2.9	-47.6	152.3
		9.1162	34.770	243.221	"	"	3.9	233.8	-41.8	228.0	6/6	369	3.5	-47.3	154.4
		9.1150-167 ¹²	34.771	243.222	"	"	4.7	232.6	-40.7	226.8	18/18	286	2.1	-47.9	156.1
		9.1168	34.770	243.220	172	36	5.5	237.6	-27.4	234.4	6/6	451	3.2	-37.3	161.5
Section 4	Dacite	9.1174	"	"	"	"	234.9	-33.5	228.7	6/6	1437	1.8	-43.9	161.1	
		9.1180	"	"	"	"	1.4	238.0	-31.4	235.5	5/6	2397	1.6	-39.4	159.5
		9.1186	"	"	"	"	-6.5	235.2	-38.0	227.2	6/6	197	4.8	-46.7	158.4
		9.1192	34.771	243.219	"	"	-1.5	234.6	-33.1	228.5	6/6	1167	2.0	-44.0	162.5
		9.1198	"	"	"	"	-2.5	239.6	-35.6	230.0	6/6	64	8.5	-40.3	156.1
		9.1204	"	"	"	"	-11.0	236.0	-42.6	226.0	6/6	23	14.2	-49.2	154.7
		9.1168-209 ¹²	34.770	243.220	"	"	-2.4	236.5	-34.6	230.4	41/42	79	2.5	-42.9	159.2
		9.1210	34.772	243.219	"	"	-41.3	207.1	-51.7	167.9	6/6	396	3.4	-79.7	323.1
		9.1216	34.772	243.218	"	"	-42.6	207.0	-52.4	166.1	6/6	655	2.6	-78.3	328.5
		9.1222	"	"	"	"	-41.2	208.0	-52.2	168.7	6/6	457	3.1	-80.4	324.3
Section 5	Upper basalt	9.1210-227 ¹²	"	"	"	"	207.4	-52.1	167.6	18/18	520	1.5	-79.5	325.4	
		9.1228	34.779	243.205	168	35	-49.6	343.9	-40.7	193.3	6/6	254	4.2	-29.1	42.9
		9.1234	34.779	243.204	"	"	11.3	342.7	-35.8	113	6/6	160	5.3	50.3	167.9
		9.1240	34.779	243.201	"	"	60.6	310.0	33.1	285.4	6/6	16	17.5	22.6	160.1
		9.1246	34.761	243.200	160	38	19.7	252.0	-18.3	252.0	4/6	288	5.4	-20.1	155.7
		0.0027	34.761	243.201	"	"	21.4	258.4	-16.2	258.1	5/5	16	19.9	-14.4	153.1
		0.0032	"	"	"	"	23.9	250.8	-14.0	250.8	5/5	448	3.6	-19.8	158.3
		0.0038	28.0	253.5	"	"	-10.0	253.1	-10.0	253.1	4/4	613	3.7	-16.7	158.8
		0.0043	34.4	249.6	"	"	3.6	249.7	-3.6	249.7	3/3	107	12.0	-17.6	163.6
		0.0021-045 ¹²	"	"	"	"	24.9	253.2	-13.1	253.0	21/21	51	4.5	-17.8	157.5
Section 6	Upper basalt	0.0046	34.761	243.200	"	"	-26.2	222.4	-56.9	200.4	5/5	152	6.2	-73.3	137.9
		0.0052	"	"	"	"	-28.2	218.1	-54.8	194.7	6/6	158	5.3	-76.0	146.3
		0.0046-057 ¹²	"	"	"	"	-26.2	220.0	-55.8	197.2	11/11	156	3.7	-75.9	141.9
		0.0058 ²	3.5	233.8	"	"	32.8	230.6	-32.8	230.6	5/5	325	4.2	-42.1	160.4
		0.0063	43.6	198.4	"	"	-43.6	198.4	-55.1	152.9	6/6	416	3.3	-67.9	343.4
		0.0069	-43.2	202.1	"	"	-43.2	202.1	-57.2	156.1	6/6	208	4.7	-70.6	349.1
		0.0075	-44.0	193.4	"	"	-44.0	193.4	-52.3	149.0	5/5	241	4.9	-64.3	337.9
		0.0063-079 ¹²	-43.7	198.2	"	"	-43.7	198.2	-55.0	152.8	17/17	239	3.9	-64.1	343.2
		0.0082 ²	-23.1	228.5	"	"	-23.1	228.5	-41.2	214.1	5/6	20	17.3	-58.4	164.6
		0.0088 ²	34.760	243.198	"	"	-8.3	223.5	-41.2	214.1	5/6	20	17.3	-58.4	164.6
Morton ³	Upper basalt	0.0092 ²	34.759	243.198	"	"	-24.0	217.9	-52.8	196.6	6/6	142	5.6	-76.2	154.1
		0.0098 ²	"	"	"	"	-18.6	226.8	-51.9	212.7	6/6	207	4.7	-62.8	148.5
		0.1104 ²	"	"	"	"	-14.4	221.5	-46.1	208.2	6/6	771	2.4	-64.8	162.3
		1.0012	34.730	243.277	230	35	-36.1	281.1	-57.2	249.6	7/7	343	3.3	-35.1	128.2
		1.0008	34.730	243.276	"	"	-4.9	311.9	-33.4	309.4	5/5	12	22.8	15.6	111.3
1.0014	"	"	"	"	-6.1	320.3	-40.8	320.3	5/5	10	25.5	20.8	102.1		
1.0008-018 ¹	"	"	"	"	-5.5	316.1	-40.1	314.8	10/10	12	14.6	18.2	106.8		
1.0019 ²	"	"	"	"	-52.2	277.4	-64.8	219.4	5/6	43	11.8	-58.2	119.1		
1.0025	"	"	"	"	-31.7	246.8	-42.7	223.3	5/5	193	5.5	-48.6	164.1		
1.0030	34.732	243.276	"	"	-33.4	258.0	-34.3	230.8	5/5	38	12.6	-45.2	152.2		
1.0035	34.732	243.275	"	"	-30.5	246.9	-33.8	224.4	6/6	19	15.9	-47.6	164.0		
1.0025-040 ¹²	34.732	243.276	"	"	-31.9	250.3	-36.8	225.9	16/16	34	6.4	-47.4	160.4		
8M018 ²	34.76	243.19	83	23	-20.4	239.9	-27.7	249.8	7/7	150	4.9	-24.8	152.1		

Notes: Lat, Long: Coordinates of sample location. North latitude and East Longitude; Str, Dip: Strike and dip of flow bedding; I, D: Inclination and declination not corrected for tilt of bedding; I_c , D_c : Inclination and Declination corrected for tilt of bedding; NIN_c : Number of samples used in calculation/Number of samples collected; K : precision parameter [Fisher, 1953]; α_{95} : Ninety-five percent confidence limit of magnetic direction, in degrees; λ , ϕ : Latitude and East longitude of Virtual Geomagnetic Pole corrected for tilt of bedding.

¹ Range of sample numbers regrouped. ² Virtual Geomagnetic Poles from these samples were used in calculation of tectonic rotation. ³ J.L. Morton (written communication, 1983)

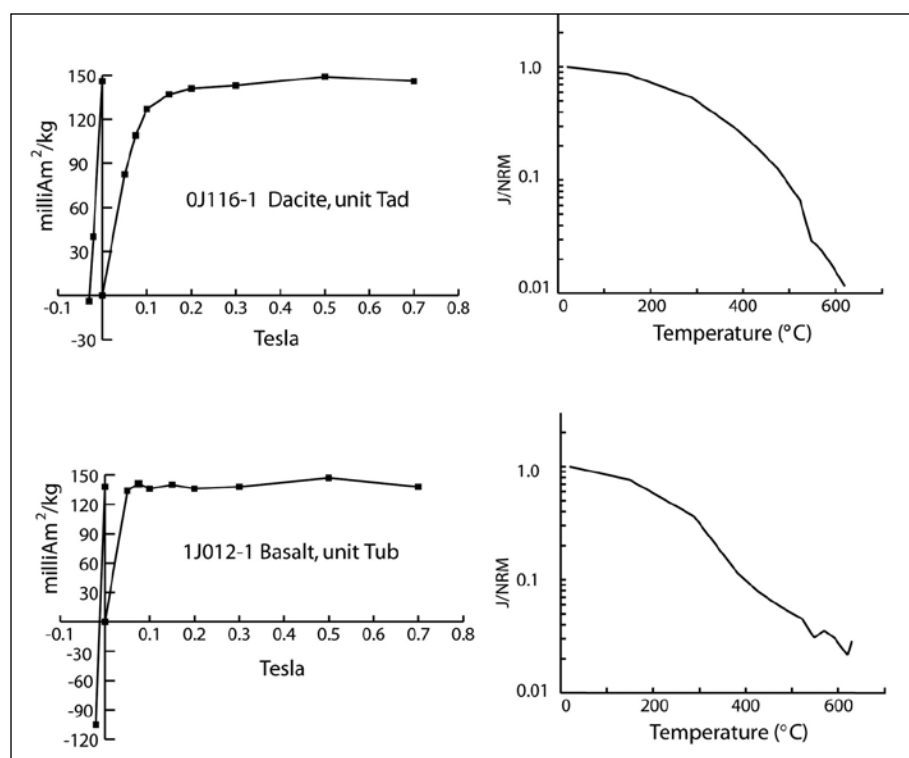


Figure 5. Isothermal remanent magnetization (IRM) and thermal demagnetization of NRM (right diagrams) from dacite (section 2, unit Tad) and basalt (section 6, unit Tub)

amount of dip. We did not expect that the tilt correction would introduce a significant apparent rotation, because the dip is moderate and mapping of the local structure did not suggest complex folding about non-horizontal axes.

Isothermal remanent magnetization (IRM) to saturation was measured from two specimens of the upper basalt unit (Section 6) and an underlying sequence of andesite and dacite flows (Section 2). Mini cores were weighed and then IRM was applied with an electromagnet to a peak field of 0.7 Tesla. Thermal demagnetization of the two specimens was performed in a low-field (<5 nanoTesla) furnace. These experiments were carried out in reconnaissance of the magnetic properties, and are not representative of the full range of properties in Miocene volcanic sequence.

Paleomagnetic results

Alternating-field demagnetization was very effective in removing secondary magnetization and reducing the scatter of magnetization directions within each site. Vector diagrams depicting the horizontal and vertical components of magnetization revealed minor secondary magnetizations that were usually removed by the 30 mT step. The diagrams consistently gave distributions of points aligned toward the origin, indicating unidirectional magnetization in each of the three superposed units T1b, Tad, and Tub (Figure 4). After calculating best-fit lines to the data, the

resultant direction for each specimen was compared to the others within a site by inspection of stereographic plots. A few specimen directions within a handful of sites lay far outside the general distribution of directions, and these data were rejected. We retained 97.5% of the original data through this data selection procedure. We then calculated a mean direction from the selected data from each site using the statistical model of Fisher [1953]. The resultant mean directions of magnetization, virtual geomagnetic poles, and statistical parameters are listed in Table 1 for the 54 cooling units collected in this study plus one site previously obtained by Janet L. Morton [written communication, 1983]. This additional sample is from an andesite flow below the upper basalt unit at the west end of the Newberry Mountains. It is from

the same outcrop that yielded a K-Ar minimum age of 19.7 Ma.

The 95-percent confidence limits ($\Rightarrow_{95\%}$; Table 1) are a measure of the quality of the site-mean direction, mainly being dependent on the amount of directional scatter among the specimens from each volcanic flow. With a few exceptions, the confidence limits for sites in this study are quite small, indicating that these potential problems are minor. About 50% of the site means have $\Rightarrow_{95\%}$ of 5° or less; 27% of the means have $\Rightarrow_{95\%}$ from 6° to 15° . Only two site means (Section 1, 9J127 and 9J148) were rejected for having unacceptably high $\Rightarrow_{95\%}$ of greater than 30° .

The principal magnetic mineral in the two test specimens is likely to be titanomagnetite, as supported by the IRM data (Figure 5). A specimen from the upper basalt (Tub) reaches IRM saturation beyond 0.1 Tesla, and the backfield curve yields a coercivity of remanence value of ~ 0.005 Tesla. The dacite specimen (Tad) nears saturation above 0.2 Tesla, again indicating titanomagnetite, although the gentle rise of the curve at higher applied fields suggests the presence of hematite in addition to the dominant titanomagnetite. Coercivity of remanence from the dacite curve is ~ 0.02 Tesla. As indicated by the thermal demagnetization examples (Figure 5), 98% of the natural remanent magnetization was unblocked by

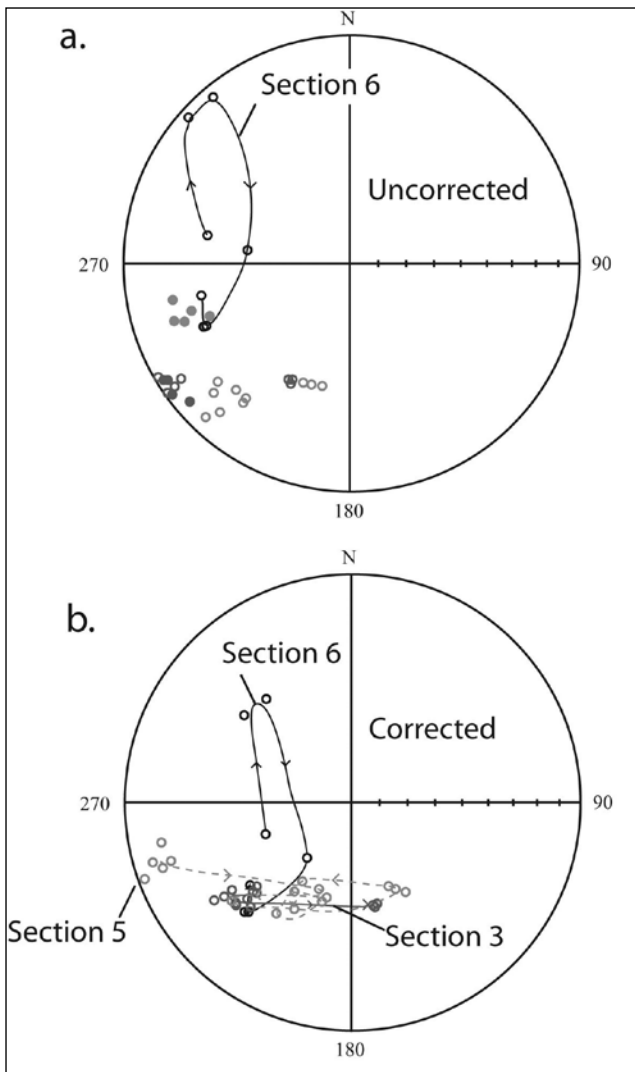


Figure 6. Equal-area projections of mean magnetic directions from individual cooling bodies in the upper basalt unit measured in sections 3 (red), 5 (green), and 6 before (a) and after (b) correction for tilt. Arrows and lines indicate stratigraphic order from bottom to top. Dots are in the lower hemisphere; circles are in the upper hemisphere.

the 550° temperature step, consistent with the presence of low-titanium titanomagnetite. A small fraction of the characteristic magnetization remains in the unblocking temperature range of 580° to 600°, indicating minor concentrations of titanohematite (Figure 4f).

The lack of variation in bedding attitudes across the area sampled prevented a successful fold test for the site means. If the tilt correction reduces the overall scatter significantly, then the magnetization is affirmed to have been acquired before the tilting occurred. However, in comparing directions from the several sections in the upper basalt (Figure 6), we find that the tilt corrections bring the means from the upper part of Section 6 into better correspondence with Sections 3 and 5. The tilt corrections also bring the inclinations of the upper basalt

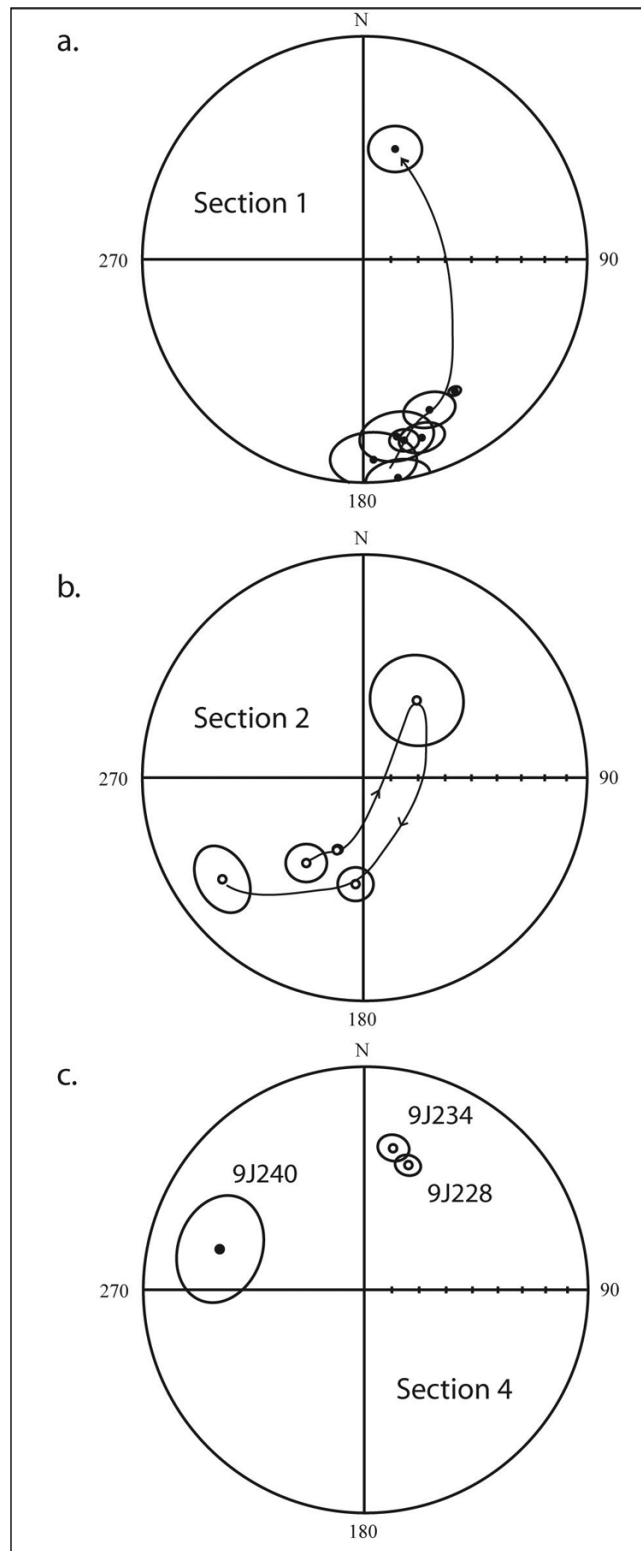


Figure 7. Equal area projections of mean magnetic directions corrected for tilt of bedding (with \Rightarrow_{95} confidence ovals) from individual cooling units from Section 1 (units T1b and Tad), Section 2 (unit Tad), and Section 4 (unit Tad). Lines and arrows indicate stratigraphic order from bottom to top. Dots are in the lower hemisphere; circles are in the upper hemisphere.

to steeper values that are closer to the expected axial dipole inclination. These observations are consistent with the magnetization being acquired before the beds were tilted, although the statistics of the fold test are not conclusive in this case. Section 6 was collected south the Sheep Spring fault zone to test for differential rotation between the southern structural domain and the more conspicuously extended northern domain. With the exception of two unusual directions from Section 6 (1J008 and 1J014), the means in Section 6 are similar to those in Sections 3 and 5 from the northern domain. Hence, we see no obvious evidence of differential rotation about a vertical axis across the domain boundary.

In the stereograms of Sections 3 and 5 (Figure 6), we see similar features in the sequence of directional changes from the southwest to the southeast quadrant from flows progressively higher in the sections. We ascribe the successive changes in declination within the stack of lava flows to paleosecular variation. Section 3 has a direction record that duplicates the middle part of Section 5, consistent with the geologic mapping that they are nearly identical in age. Matching of the fine detail is further evidence that the remanent magnetization is a primary component imparted during original solidification and cooling of the basalt.

Viewing our paleomagnetic results in stratigraphic order, we see the following: 1, a transition to normal polarity at the top of the lower basalt (Tlb; Figure 7a); 2, a return to reversed polarity plus a looping excursion into the northeast quadrant in the middle unit (Tad; Figure 7b); 3, continued reversed polarity with the direction of magnetization oscillating east-west in the upper basalt (Tub; Figure 6). In Section

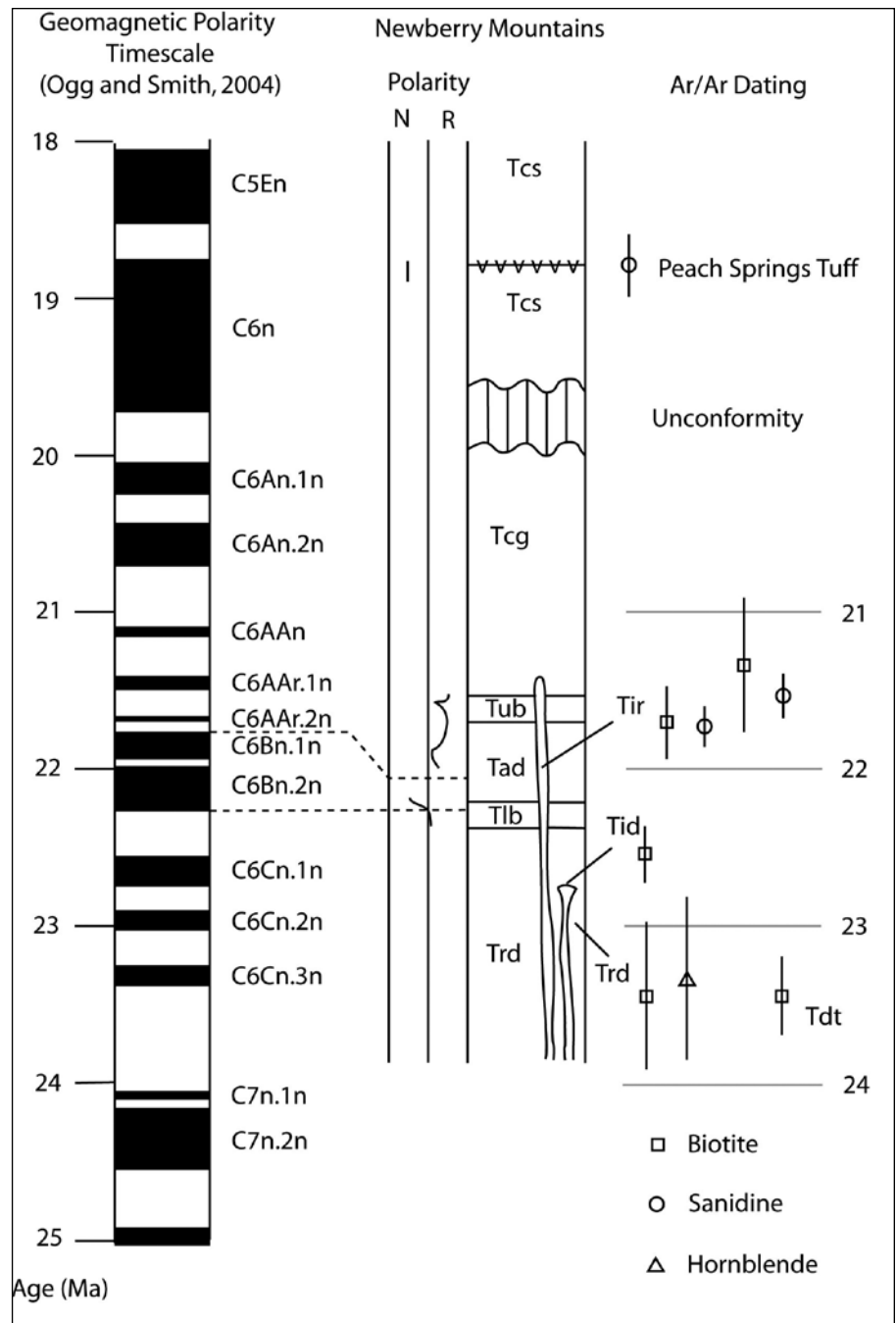


Figure 8. Comparison of $^{40}\text{Ar}/^{39}\text{Ar}$ dating and polarity zonation of volcanic and intrusive rocks in the Newberry Mountains with the Geomagnetic Polarity Timescale of Ogg and Smith [2004] as modified from Cande and Kent [1995]. Black bands indicate normal polarity; white bands denote reversed polarity. Polarity of Newberry Mountains volcanic succession is represented by schematic VGP latitude curve. Age-determinations from the Newberry Mountains were converted from the original source [Cox, in press], using the monitor age of 28.02 Ma (Fish Canyon Tuff sanidine equivalent age). Unit labels same as Figure 2; Tid, dacite dikes; Tdt, tuff near base of lower volcanic succession. Vertical bars indicate standard deviations of age analyses. Age of Peach Springs Tuff is from Hillhouse et al. [this volume].

6, we observe an excursion into the northwest quadrant, but we cannot be certain of this section's stratigraphic position within the upper basalt relative to Sections 3 and 5; most likely, Section 6 is stratigraphically above Sections

3 and 5. At a minimum, the sequence of paleomagnetic determinations samples a time interval that spans parts of two polarity zones and one polarity transition. This duration should be adequate to satisfy the geocentric axial dipole assumption when the paleomagnetic directions are averaged. Therefore, the data are suitable for drawing conclusions about tectonic rotations, provided the transitional directions are excluded.

Cox [in press] reported several $^{40}\text{Ar}/^{39}\text{Ar}$ dates that have enough precision to tie the Newberry Mountains polarity sequence to the marine magnetic anomalies of the Neogene Geomagnetic Polarity Timescale (GPT) of Lourens et al. [2004] and Ogg and Smith [2004]. To make the tie, we must convert the reported $^{40}\text{Ar}/^{39}\text{Ar}$ dates to account for revision to older ages of most magnetic polarity boundaries in the recent “astronomically tuned” GPT (Figure 8). Isotopic ages from the Newberry Mountains, as reported by Cox [in press], were obtained from the Menlo Park USGS laboratory, which used a monitor age equivalent to Fish Canyon Tuff (FCT) sanidine at 27.56 Ma. To make the proper comparison of the Newberry Mountains polarity zonation with the GPT, we increased the $^{40}\text{Ar}/^{39}\text{Ar}$ ages from the original source by 1.71% after assuming FCT = 28.02 Ma [Renne et al., 1994].

Our favored model for correlating the magnetozones of the Newberry Mountains volcanic sequence with the GPT is given in Figure 8. A rhyolite dike swarm (Figure 2, unit Tir), with converted ages ranging from 21.7 Ma to 21.3 Ma, is the probable source of rhyolite flows that directly underlie the upper basalt unit (Figure 8, uppermost part of unit Tad). Furthermore, at the southeast end of the range, an outlying rhyolite dike dated at 21.8 - 21.7 Ma intrudes basalt flows that are tentatively correlated with the upper basalt unit, implying that the latter may be older than about 21.7 Ma. The lower two units studied by this survey (Figures 2 and 6, units Tlb and Tad) are undated, but are constrained by isotopic dating of the rhyolite dikes and underlying flows to be older than 22 Ma and younger than 23.3 Ma. From the preponderance of reversed magnetic polarity, it appears likely that the lower basalt, upper basalt, and intervening strata all accumulated during the middle part of Chron 6, between about 22.5 and 21.7 Ma.

Tectonic rotation

Before making the rotation calculations we decimated the data to remove the transition and excursion directions, which reflect extraordinary features of the “normal” paleosecular variation. In the lower basalt unit (Section 1), the site mean from only the uppermost flow was retained for the tectonic analysis (9J145); the underlying flows, which yield south and shallow downward directions, are clearly part of a polarity transition to normal polarity (Figure 7). In the upper part of the Miocene sequence, six VGP's (0J122, 1J008, 1J014, 9J228, 9J234, and 9J240) are well removed from the main body of data. After calculating the mean VGP from the remaining data, we found that the six outlying VGP's were more than 55° from the mean.

Also, by inspection we reduced the number of VGP's by combining nearly identical results from adjacent cooling units to eliminate repeat sampling of brief time intervals. This procedure was undertaken to reduce possible bias in the rotation calculation arising from oversampled sequences of flows. Stereograms from the upper basalt (Sections 3, 5, and 6) show several clusters of nearly

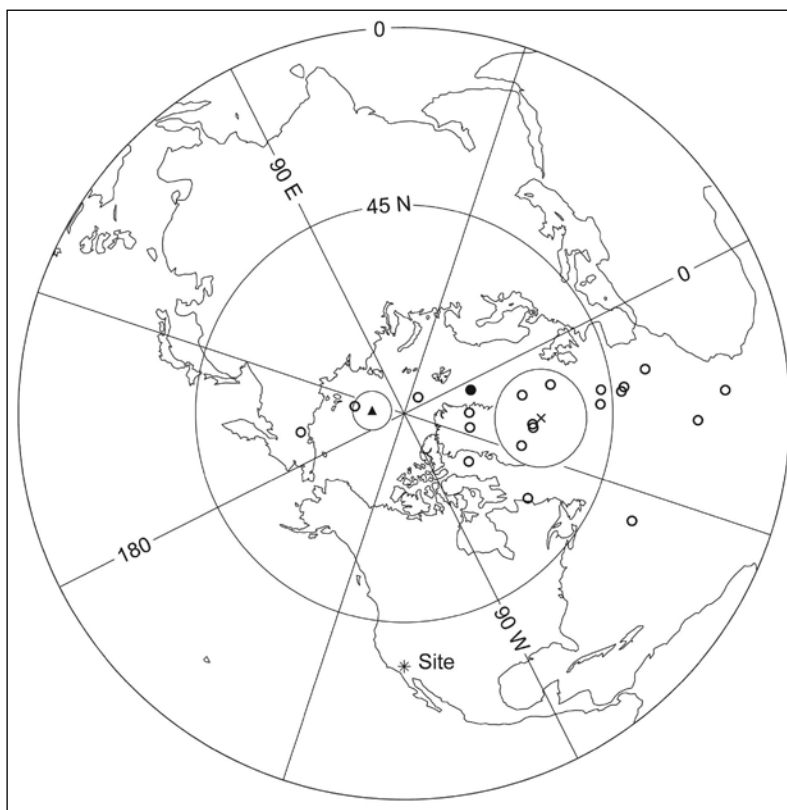


Figure 9. Virtual geomagnetic poles (dots: normal polarity, circles: reversed polarity) and mean paleomagnetic pole (X) from Miocene volcanic rocks of the Newberry Mountains. The Oligocene-Miocene paleomagnetic reference pole (triangle) for North America [22-38 Ma, Diehl et al., 1983] is shown for comparison. Large circles show \Rightarrow 95 confidence limits of paleomagnetic poles.

identical directions from superjacent flows (Figure 6). Each cluster consists of 3 to 5 site means with overlapping 95-percent confidence circles. We interpret these clusters to be caused by episodic volcanic eruptions occurring at a rapid pace relative to the rate of geomagnetic directional change. The clusters might also be simply the result of sampling overlapping streams of lava from a single eruption. One particular cluster in Section 5 represents a westerly swing of the magnetic direction to the outer fringe of the distribution of all site means, so the cluster is likely to span a very brief interval of time. We regrouped the data within each cluster to calculate a new site mean and statistics.

Table 1 lists the 20 site means retained from the original 54 sites that were collected in this study plus site 8M018. The resultant mean paleomagnetic pole (Figure 9) from the 21 selected sites in the Newberry volcanic rocks is 60.9° S, 151.4° E, $\alpha_{95} = 10.0^\circ$, $K=11.2$ (reversed polarity). The angular standard deviation (24.2°) of VGP's exceeds the global average (15°) for the site latitude, which implies sufficient temporal sampling of secular variation to reflect the axial dipole field (Merrill and McElhinny, 1983; p. 204). For a reference pole we use the southern-hemisphere equivalent to the Oligocene-early Miocene North American pole of Diehl et al. [1983]. We infer a tectonic rotation about the vertical axis centered on the Newberry Mountains of $42.7^\circ \pm 10.3^\circ$ clockwise. The anomaly in paleolatitude ($3.1^\circ \pm 8.6^\circ$) is not significant [Debiche and Watson, 1995].

Comparing the results of our study with the earlier work of Ross et al. [1989], our study confirms evidence for vertical-axis clockwise rotation of the early Miocene rocks of the Newberry Mountains. Ross et al. [1989] reported a clockwise rotation of $72.7^\circ \pm 20.0^\circ$, based on remanent magnetization measurements from 11 sites in andesite and basalt, including Janet Morton's data from site 8M018 (Table 1). The sites sampled by Ross et al. [1989] are mainly east of the sections sampled in our study, but some sites are in rocks equivalent to the beds studied here. Although the sense of rotation is the same for both studies, our estimate ($42.7^\circ \pm 10.3^\circ$) is significantly smaller than the value reported earlier. The difference may be due to local structural differences, different ages of the sampled volcanic units, or different samplings of secular variation. The flows and sills Ross et al. [1989] sampled in the eastern Newberry Mountains are generally older than the rocks in our study, so our result may reflect the later stages of the rotation episode.

Section 6 was collected south of the Sheep Spring fault to test for tectonic rotation in the southern structural domain of the Newberry Mountains, which generally appears to be less extended than the northern domain.

Declinations from Section 6 are generally consistent with data from Sections 3 and 5, suggesting that the Sheep Spring fault and adjacent monocline do not accommodate differential rotation relative to sites in the northern domain. A similar test of the structural relations across the Kane Springs fault would be difficult, because it does not appear that any volcanic beds of the appropriate age for a paleomagnetic test extend south of Kane Wash.

The age of the Newberry Mountains rotation is constrained by the Peach Spring Tuff (18.8 Ma, Hillhouse et al., this volume), which overlies Section 6 (Figure 2) and has paleomagnetic declinations indicating no significant vertical-axis rotation relative to the Colorado Plateau [Wells and Hillhouse, 1989]. The rotation mechanism, therefore, must be related to early Miocene tectonics. Recently active strike-slip faults, such as the Camp Rock fault, cannot be the principal cause of the rotation of the Newberry Mountains.

Valentine et al. [1993] reported paleomagnetic data from 2 sites in early Miocene volcanic rocks near Daggett Ridge (Figure 1), 10 km west of the Newberry Mountains. Their results (site 95: inclination = -30.3° , declination = 220.2° ; site 96: -45.1° , 228.7°) are consistent with the clockwise sense and magnitude of rotation that we infer from the Newberry paleomagnetic data. Therefore, we suggest that the Daggett Ridge area is the western extension of the rotated Newberry Mountains domain.

We note that the azimuths of Mesozoic porphyritic felsite dikes in the Rodman Mountains [Dibblee, 1964b] diverge in the clockwise sense ($\sim 55^\circ$) from the mean trend of presumably equivalent dikes in the Ord Mountains [Dibblee, 1964a] west of the Camp Rock fault (Figure 10). The dikes in question are considered part of the Jurassic Independence dike swarm, which has been proposed as a rotation indicator in the southern Sierra Nevada and Mojave Desert [James, 1989; Ron and Nur, 1996; Hopson et al., 2008]. Therefore, the inferred clockwise rotation of the dikes in the Rodman Mountains is consistent with the clockwise sense and general magnitude of rotation that is suggested by paleomagnetic data in the Newberry and Rodman Mountains. The Independence dike orientations also indicate that a major structure within the basement rocks has accommodated significant rotation between the Ord Mountains and Newberry Mountains.

As shown by the paleomagnetic data of Ross et al. [1989], the domain of early Miocene clockwise rotation extends eastward from the Newberry Mountains to the Bristol Mountains (Figure 10). Most, if not all, of the rotations occurred before 18.8 Ma, as indicated by relatively small paleodeclination anomalies measured from the Peach Spring Tuff [Wells and Hillhouse, 1989]. However,

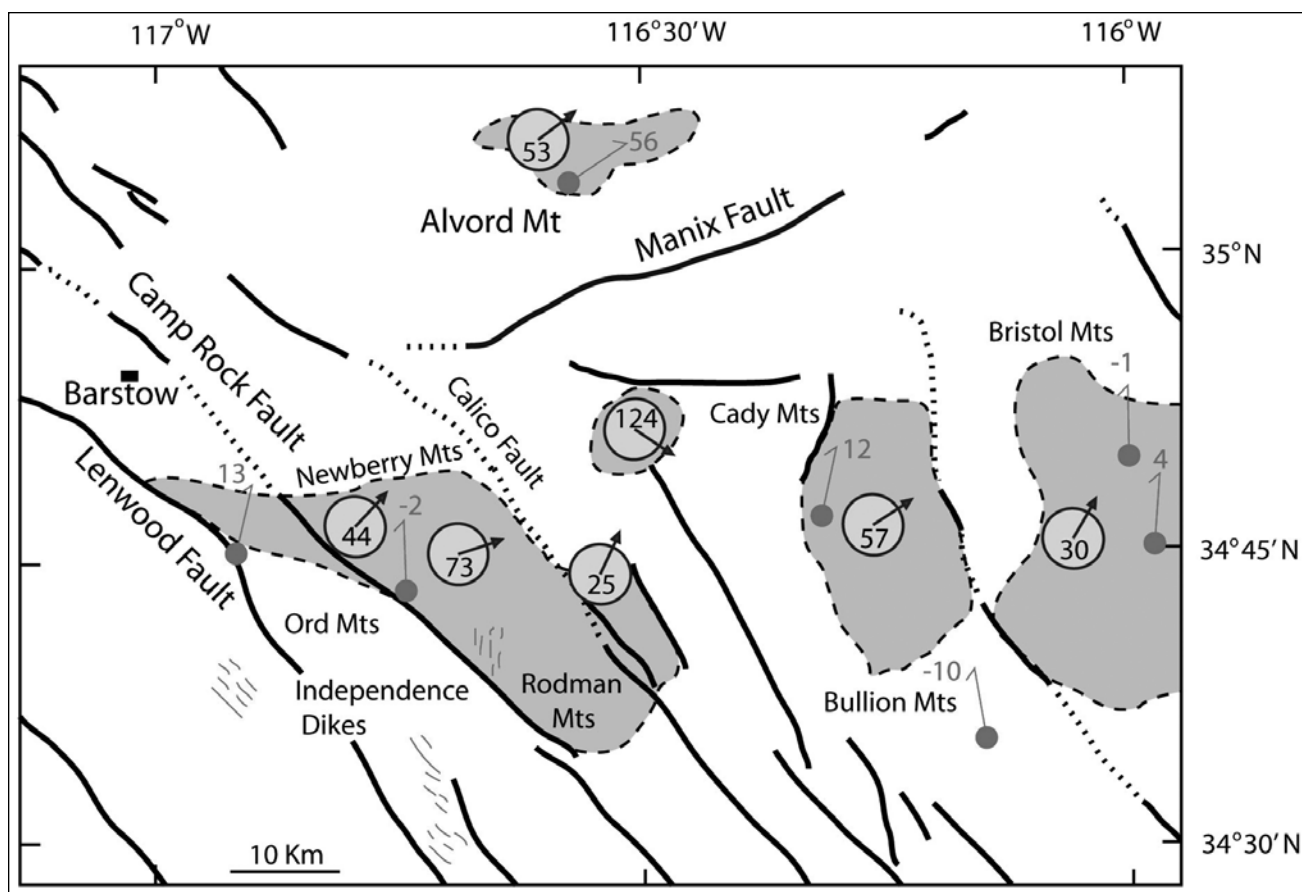


Figure 10. Map of the central Mojave Desert showing areas (tinted) that underwent clockwise rotation in Early Miocene time, as inferred from paleomagnetism of volcanic rocks. Larger circular symbols indicate amounts of vertical-axis rotation (in degrees) reported in this study (44°) and by Ross et al. [1989]. With the exception of the rotation of Alvord Mountain, the clockwise rotations mainly occurred before deposition of the Peach Springs Tuff (18.5 Ma). Red arrows denote rotations inferred from paleomagnetism of the Peach Springs Tuff [Wells and Hillhouse, 1989; this study].

rotations in rocks as young as 12.8 Ma were measured paleomagnetically at Alvord Mountain [Ross et al., 1989] and near Fort Irwin [Luyendyk et al., 1980; Schermer et al., 1996]. These younger rotations are presumed to be associated with east-striking sinistral fault systems confined to the northeastern part of the Mojave Desert. Geodetic measurements [Svarc et al., 2002; Savage et al., 2004] in the northeastern block give a clockwise rate of rotation during the last decade that is consistent with the Miocene net rotation inferred from paleomagnetic measurements in the same region.

We obtained paleomagnetic data from a welded tuff at Alvord Mountain that affirms the sense and magnitude of Miocene rotation reported by Ross et al. [1989]. In the eastern part of Alvord Mountain, the Spanish Canyon Formation [Byers, 1960] contains a sanidine- and sphene-bearing tuff that compositionally resembles the Peach Spring Tuff. We sampled this welded tuff for paleomagnetic analysis about 2 km southwest of Clews Ridge (SW $\frac{1}{4}$ sec. 4, T. 11 N., R. 4 E.; Lat. 35.069° N., Long. 243.432° E.). Using the demagnetization procedure and

line-fitting method described previously, we obtained a mean magnetization direction of 40.2° (inclination), 83.7° (declination), $\varphi_{95} = 4.5^\circ$, from 10 specimens. After being corrected for tilt of the bedding (strike 55°, dip 8° SE), the magnetization direction of the tuff was 36.1° (inclination), 89.1° (declination). This corrected inclination closely matches the unusually low inclination (36.4°) of the Peach Spring Tuff [Wells and Hillhouse, 1989]. We conclude that the Alvord Mountain tuff is correlative with the Peach Spring Tuff at its type locality in Arizona, given the compositional similarities and the distinctive magnetic inclination exhibited by both. After comparing the declinations obtained from the Alvord tuff locality with the reference declination from the Peach Spring Tuff on the Colorado Plateau, we infer a clockwise rotation of $56.1^\circ \pm 5.6^\circ$ at the eastern side of Alvord Mountain. This value is consistent with the 53.2° clockwise rotation measured by Ross et al. [1989] in basalt in the Alvord Mountain area. Miller and Walker [2002] questioned the validity of the rotation estimate, because apparent Independence dike alignments are undisturbed in the western part of Alvord

Mountain. In the eastern part of Alvord Mountain, especially east of Spanish Canyon, the dikes strike with a bimodal distribution of azimuths. Strikes of the eastern Alvord dikes are generally rotated 30°–90° clockwise from the dominant northwest trend of the Independence dike swarm [Miller and Walker, 2002]. Apparent rotation of the eastern Alvord dikes is consistent with the sense and magnitude of rotation that we measured from paleomagnetism of the Peach Spring Tuff east of Spanish Canyon.

In a synthesis of the regional tectonic history, Dokka and Ross [1995] and Dokka et al. [1998] attributed early Miocene rotation to forces generated in the upper crust of the Mojave region as the North America/Pacific plate boundary was changing from a subduction zone to a transform boundary. According to their interpretation, northwestward movement of the Pacific plate opened a crustal gap in the central Mojave region. Gravitational collapse then led to north-south extension above detachment faults. Atwater and Stock's (1998) revised Neogene plate motion history for the southwestern United States shows a northward-migrating slab window beneath the central Mojave Desert between 24 and 19 Ma, approximately the time of major volcanism and extension. As the extension waned about 19 Ma, the southern Sierra and central Mojave Desert underwent dextral shear, resulting in vertical-axis clockwise rotations of 40°–60°. In the area northwest of Barstow, Valentine et al. [1993] reported paleomagnetic evidence of substantial clockwise rotation in earliest Miocene time, followed by an episode of counterclockwise rotation (23°) prior to 18 Ma. They ascribed regional variations in the sense and amount of rotation to drag along transfer zones between extended domains. Glazner et al. [2002] argued against large-scale extension in the Newberry Mountains, citing evidence of transpressional structures that formed during the middle or late Miocene. They linked clockwise rotation in the Newberry Mountains to dextral shear in a zone that connects the highly extended

Table 2. Paleomagnetic directions and Virtual Geomagnetic Poles from the Newberry Mountains polarity transition

USGS #	I_c	D_{CR}	λ_{CR}	ϕ_{CR}
9J121	11.3	133.4	-30.3	300.2
9J124	1.1	127.2	-29.4	309.4
9J127	13.5	136.2	-31.4	296.9
9J130	21.1	125.4	-21.1	302.3
9J133	17.9	123.8	-21.1	304.8
9J136	17.6	118.2	-17.1	308.9
9J139	27.4	112.5	-9.3	308.3
9J142	29.1	101.5	-0.3	314.0
9J145	47.4	332.3	65.7	146.1

Notes: I_c , Inclination corrected for tilt of bedding; D_{CR} , Declination corrected for tilt and vertical-axis rotation; λ_{CR} , ϕ_{CR} , Latitude, longitude of Virtual Geomagnetic Pole, corrected for tilt and rotation. See Table 1 for locations and other details.

terranes of the Waterman Hills with the Colorado River core complexes (see also Bartley and Glazner, 1991).

Our study of the Newberry Mountains constrains the timing of volcanism, rotation and normal faulting to the interval from 23 Ma to 18.8 Ma, which coincides with the passage of the proposed slab window. Large amounts of rotation (exceeding 40°) that are measured in the central Mojave Desert (Figure 10) would be more easily accommodated by a shallow decollement than by bending of deep-seated blocks. Therefore, we favor a rotation

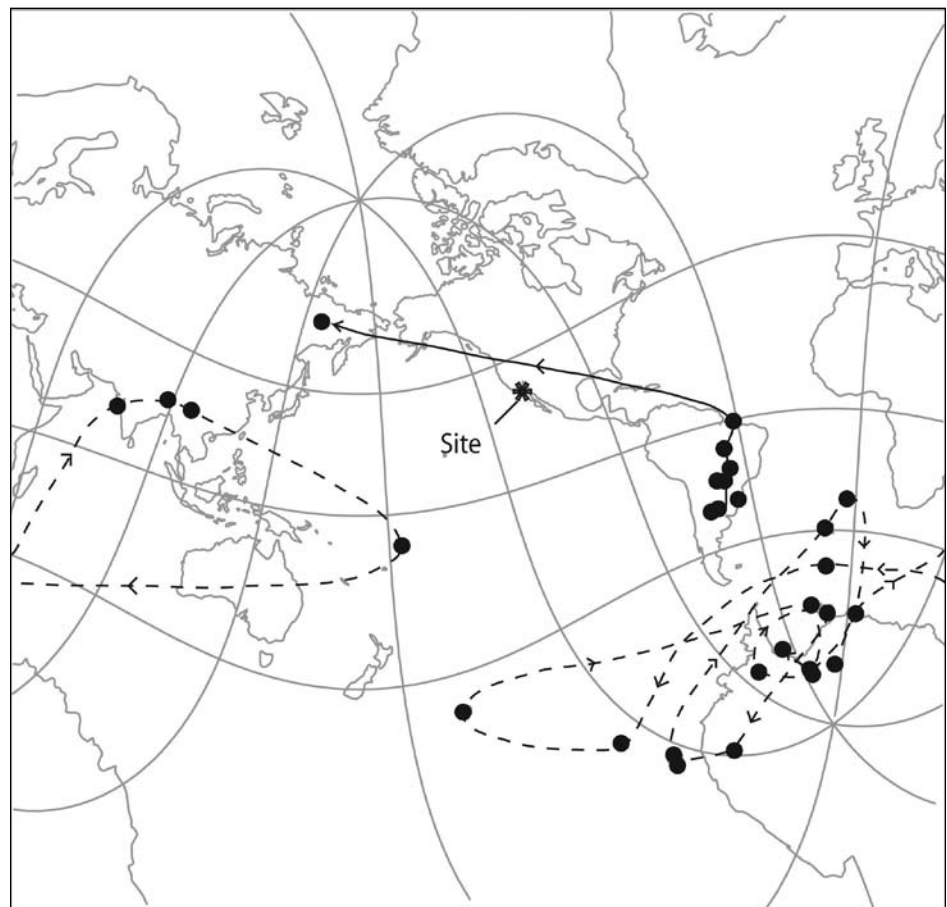


Figure 11. Virtual geomagnetic pole sequences from polarity transition (solid line) and paleosecular variation (dashed line) from volcanic units in the Newberry Mountains. All VGP's were corrected for clockwise rotation. Arrows depict stratigraphic sequence from older to younger ages.

mechanism involving large-scale extension to explain the Miocene tectonics of the Newberry Mountains–Bristol Mountains region.

Polarity transition

Paleomagnetic results from the lower basalts offer a rare glimpse of a polarity transition. The usual convention for presenting transitions is to convert directions of magnetization to virtual geomagnetic poles (VGP) to make worldwide comparisons of the reversal phenomenon. Although the VGP calculation is made with the dipole formula, this does not necessarily imply that the geomagnetic field was dipolar during the transition. In fact, the preponderant evidence from paleomagnetic studies of polarity transitions supports decay of the main dipole field and dominance of non-dipole features during the transition process. Before converting the Newberry transition to VGP's, we first correct the magnetization vectors for the vertical-axis clockwise rotation noted in the previous section (Table 2). The corrected VGP path (Figure 11) begins in South America at 31° S, crosses the Equator in northern Brazil, and then jumps to its endpoint in northeast Russia. The transition path contains 8 low-latitude VGP's. Because the oldest sample occurs within the low-latitude portion of the transition, we cannot be certain of the polarity state prior to onset of the transition. Using the terminology of Love [1998], we classify the Newberry Mountains transition as undetermined to normal polarity (I-N).

From a study of seven polarity transitions from volcanic rocks 8.5 Ma and younger, Hoffman [1992] observed preferential clustering of VGP's in two regions, one in the south Atlantic Ocean, including the southern part of South America, and the other near western Australia. Also, he noted the correspondence of the clusters with two sectors of longitude in which transitional VGP's from sediments were concentrated [Laj et al., 1992]. Hoffman [1992] reasoned that, if these preferred patterns of the transitional field could recur for millions of years, the lower mantle must be exerting control on the geomagnetic reversal process. This topic has stirred controversy, as the statistical significance of preferred VGP paths in transition compilations has been rejected by one study [Prevot and Camps, 1993] and supported by another [Love, 1998]. (For a review of the controversy, see Merrill and McFadden [1999]).

The Newberry Mountains transition, which has 8 VGP's falling within South America and partially overlapping the south Atlantic cluster observed by Hoffman [1992], adds support to the preferred-path hypothesis. At ~22 Ma, the Newberry transition is considerably older

than the bulk of polarity transitions studied in the above-mentioned compilations. The Steens Mountain transition, an extensive record from Miocene lavas (15.5 Ma) in Oregon, has a VGP path that transits South America before stabilizing in the normal polarity state [Mankinen et al., 1985; Mankinen, 1986]. Miocene basalts (~14.1 Ma) at Gran Canaria reveal a polarity transition (R-N) with VGP's clustering east of southern South America, in addition to clusters near India and the central Pacific Ocean [Leonardt et al., 2002]. Therefore, we see mounting evidence that transitional VGP's have occurred preferentially within the south Atlantic cluster since Early Miocene time.

Conclusions

Aided by detailed geologic mapping of the Newberry Mountains, we have measured remanent magnetization from a stratigraphically controlled sequence of basalt, andesite, and dacite of early Miocene age. After treatment by alternating fields to remove secondary magnetizations, the volcanic specimens give well-defined directions of magnetization that span a polarity transition and at least two polarity epochs. Details of paleosecular variation are repeated in several sections of basalt, consistent with the mapped stratigraphy, and affirm the primary nature of the magnetization. In comparison to the Geomagnetic Polarity Timescale [Cande and Kent, 1995; Ogg and Smith, 2004], the Newberry Mountains polarity sequence fits best with the middle part of Chron 6, and this correlation indicates that the lower and upper basalt units differ in age by approximately 0.5 million years. To calculate a mean paleomagnetic pole for tectonic analysis, we excluded the polarity transition and several unusual excursion directions. We also combined data from several volcanic flows to reduce bias that could arise from oversampling parts of the paleomagnetic record. After the magnetic directions are corrected for tilt of the bedding, the mean paleomagnetic pole for the Newberry Mountains volcanic rocks (~22-21 Ma) is: latitude 60.9° S, longitude 151.4°E, $\Rightarrow_{95} = 10.0^\circ$. We compared the Newberry pole to the Oligocene-early Miocene reference pole from North America [Diehl et al., 1983] to infer a vertical-axis rotation of the region. Finally, we corrected the magnetic directions of the polarity transition to remove the effects of tilt and vertical-axis rotation. The following conclusions are drawn from this study.

1. Clockwise rotation ($42.7^\circ \pm 10.3^\circ$) of the Newberry Mountains occurred 21-18.8 million years ago, at a time when the upper crust of the central Mojave Desert was being

extended. The sense of rotation is consistent with previous paleomagnetic work [Ross et al., 1989] in the same mountains, but the amount of rotation is significantly less in our study. Also, angular unconformities within the Miocene volcanic section indicate that tilting along normal faults occurred during volcanism.

2. We tested for differential rotation across a zone of east-striking faults and a parallel monocline north of Kane Wash, but found no significant difference in declination from sites on either side of the zone. The southern limit of early Miocene rotation is not constrained by our study. Rotation may have terminated at the Kane Springs fault, which forms the southern boundary of lower Miocene volcanic rocks in the Newberry Mountains region. Alternatively, the rotation may have extended southward into the Rodman Mountains, where anomalously oriented Jurassic dikes indicate that pre-Tertiary basement rocks may have been rotated clockwise ($\sim 55^\circ$) since 148 Ma. The recently active Camp Rock fault bounds the west side of the Newberry Mountains, but the fault is a young structure that postdates the rotation.

3. Virtual geomagnetic poles from the Newberry Mountains polarity transition pass through South America, in agreement with Hoffman's [1992] hypothesis that transitional poles fall preferentially within clusters in the South Atlantic or in western Australia. Persistence of the South America cluster since at least Early Miocene time suggests that the geomagnetic field is influenced by the lower mantle during the polarity-reversal process.

Acknowledgments

We thank Vicky Pease and Robert Bogar for assistance during this study, and acknowledge Janet Morton for sharing her paleomagnetic data. We have benefited from thoughtful reviews by David M. Miller and Mark Hudson.

References

- Atwater, T., and J. Stock, Pacific-North America plate tectonics of the Neogene southwestern United States; an update, in Ernst, W. G. and Nelson, C. A., eds., *Integrated earth and environmental evolution of the southwestern United States: The Clarence A. Hall, Jr. volume*: Columbia, Bellweather Publishing, p. 393-420, 1998.
- Bartley, J.M., A. F. Glazner, and E. Schermer, 1990, North-south contraction of the Mojave block and strike-slip tectonics in southern California: *Science*, v. 248, p. 1398-1401.
- Bartley, J. M., and A. F. Glazner, En echelon Miocene rifting in the southwestern United States and model for vertical-axis rotation in continental extension, *Geology*, 19, 1165-1168, 1991.
- Byers, F. M., *Geology of the Alvord Mountain Quadrangle, San Bernardino County, California*: U. S. Geol. Surv. Bull. 1089-A, 71 p., 1960.
- Bortugno, E. J., and T. E. Spittler, Geologic map of the San Bernardino quadrangle (California), Radiometric ages of rocks in the San Bernardino quadrangle, California, Calif. Div. Mines and Geol., Regional Geologic Map Series, No. 3A, Sheet 4, scale 1:250,000, 1-26 pp., 1986.
- Burke, D. B., J. W. Hillhouse, E. H. McKee, S. T. Miller, and J. L. Morton, Cenozoic rocks in the Barstow basin area of southern California – Stratigraphic relations, radiometric ages, and paleomagnetism, U. S. Geol. Surv. Bull. 1529-E, E1-E16 pp., 1982.
- Cande, S. C., and D. V. Kent, Revised calibration of the geomagnetic polarity timescale for the Late Cretaceous and Cenozoic, *J. Geophys. Res.*, 100, 6093-6095, 1995.
- Cox, B. F., Stratigraphy and structure of the Newberry Mountains, southern California, and implications for tectonic evolution of the Mojave Desert province, U. S. Geol. Surv. Sci. Invest. Rep., in press.
- Cox, B. F., Structure and tectonics at the south edge of an early Miocene rift zone, central Mojave Desert, California (abs.): San Bernardino County Museum Association Quarterly, v. 42, no. 2, p. 30-31, 1995.
- Cox, B. F., New geologic map of the Newberry Mountains, southern California, and implications for Miocene tectonic history of the Mojave Desert (abs.): Geological Society of America Abstracts with Programs, v. 37, no. 7., Paper no. 196-7, 2005.
- Cox, B. F., A. Griscom, J. E. Kilburn, G. L. Raines, and D. H. Knepper, Jr., Mineral resources of the Newberry Mountains and Rodman Mountains Wilderness study areas, San Bernardino County, California, U. S. Geol. Surv. Bull. 1712-A, A1-A28 pp., 1987.
- Debiche, M. G., and G. S. Watson, Confidence limits and bias correction for estimating angles between directions with applications to paleomagnetism, *J. Geophys. Res.*, 100, 24,405-24,429, 1995.
- Dibblee, T. W., Jr., Evidence of strike-slip faulting along northwest-trending faults in the Mojave Desert, U. S. Geol. Surv. Prof. Pap. 424-B, B197-B199, 1961.
- Dibblee, T. W., Jr., Geologic map of the Ord Mountains quadrangle, San Bernardino County, California, U. S. Geol. Surv. Misc. Geol. Inv. Map, I-427, scale 1:62,500, 1964a.
- Dibblee, T. W., Jr., Geologic map of the Rodman Mountains quadrangle, San Bernardino County, California, U. S. Geol. Surv. Misc. Geol. Inv. Map, I-430, scale 1:62,500, 1964b.
- Dibblee, T. W., Jr., Geologic map of the Daggett quadrangle, San Bernardino County, California, U. S. Geol. Surv. Misc. Geol. Inv. Map, I-592, scale 1:62,500, 1970.
- Dibblee, T. W., Jr., and A. M. Bassett, Geologic map of the Newberry quadrangle, San Bernardino County, California, U. S. Geol. Surv. Misc. Geol. Inv. Map, I-461, scale 1:62,500, 1966.
- Diehl, J. F., M. E. Beck, Jr., S. Beske-Diehl, D. Jacobson, and B. C. Hearn, Jr., Paleomagnetism of the Late Cretaceous-early Tertiary north-central Montana alkalic province, *J. Geophys. Res.*, 88, 10,593-10,609, 1983.
- Dokka, R. K., Late Cenozoic tectonics of the central Mojave Desert, California, Los Angeles, University of Southern California, Ph.D. dissertation, 193 pp., 1980.
- Dokka, R. K., Displacements on Late Cenozoic strike-slip faults of the central Mojave Desert, California, *Geology*, 11, 305-308, 1983.

- Dokka, R. K., Patterns and modes of early Miocene extension of the central Mojave Desert, California, in *Continental Extension Processes*, edited by L. Mayer, *Spec. Pap. Geo. Soc. Am.* 208, 75-95, 1986.
- Dokka, R. K., The Mojave Extensional Belt of southern California, *Tectonics*, 8, 363-390, 1989.
- Dokka, R.K., Original dip and subsequent modification of a Cordilleran detachment fault, Mojave extensional belt, California, *Geology*, 21, p. 711-714, 1993.
- Dokka, R. K., and A. F. Glazner, Aspects of early Miocene extension of the central Mojave Desert, in Cooper, J. D., compiler, *Geologic Excursions in the California Desert*, *Geol. Soc. Amer. Cordilleran Section Guidebook*, 31-45, 1982.
- Dokka, R. K., and T. M. Ross, Collapse of southwestern North America and the evolution of early Miocene detachment faults, metamorphic core complexes, the Sierra Nevada orocline, and the San Andreas fault system, *Geology*, 23, 1075-1078, 1995.
- Dokka, R. K., and C. J. Travis, Late Cenozoic strike-slip faulting in the Mojave Desert, California, *Tectonics*, 9, 311-340, 1990a.
- Dokka, R.K., and C. J. Travis, Role of the Eastern California Shear Zone in accommodating Pacific-North American plate motion, *Geophysical Research Letters*, 19, p. 1323-1326, 1990b.
- Dokka, R. K., and M. O. Woodburne, Mid-Tertiary extensional tectonics and sedimentation, central Mojave Desert, California, *Louisiana State Univ. Publ. Geol. Geophys., Tectonics and Sedimentation*, 1, 55 pp., 1986.
- Dokka, R. K., M. McCurry, M. O. Woodburne, E.G. Frost, and D. A. Okaya, A field guide to the Cenozoic crustal structure of the Mojave Desert, in Weide, D.L., and Faber, M.L., eds., *This extended land—Geological journeys in the southern Basin and Range: Las Vegas, University of Nevada, Department of Geoscience Special Publication no. 2*, p. 21-44, 1988.
- Dokka, R. K., T. M. Ross, and G. Lu, The Trans Mojave-Sierran shear zone and its role in early Miocene collapse of southwestern North America, in: Holdsworth, R. E., Strachan, R. A., and Dewey, J. F. (eds.), *Continental Transpressional and Transtensional Tectonics: Geological Society, London, Special Publications 135*, p. 183-202 1998.
- Fisher, R. A., Dispersion on a sphere, *Proc. Roy. Soc. London*, A217, 295-305, 1953.
- Garfunkel, Z., Model for the late Cenozoic tectonic history of the Mojave Desert, California, and for its relation to adjacent regions, *Geol. Soc. Am. Bull.*, 85, 1931-1944, 1974.
- Glazner, A. F., J. E. Nielson, K. A. Howard, and D. M. Miller, Correlation of the Peach Springs Tuff, a large-volume Miocene ignimbrite sheet in California and Arizona, *Geology*, 14, 840-843, 1986.
- Glazner, A. F., J. M. Bartley, and J. D. Walker, Geology of the Waterman Hills detachment fault, central Mojave Desert, California, in Weide, D. L., and Faber, M. L., eds., *This extended land-- Geological journeys in the southern Basin and Range: Las Vegas, University of Nevada, Department of Geosciences Special Publication no. 2*, 1988.
- Glazner, A. F., J. M. Bartley, and J. D. Walker, Magnitude and significance of Miocene crustal extension in the central Mojave Desert, California, *Geology*, 17, 50-53, 1989.
- Glazner, A. F., J. M. Bartley, and W. K. Sanner, Nature of the southwestern boundary of the central Mojave Tertiary province, Rodman Mountains, California, *Geol. Soc. Am. Bull.*, 112, 34-44, 2000.
- Glazner, A. F., J. D. Walker, J. M. Bartley, and J. M. Fletcher, Cenozoic evolution of the Mojave block of southern California, in Glazner, A. F., J. D. Walker, and J. M. Bartley, eds., *Geologic Evolution of the Mojave Desert and Southwestern Basin and Range: Boulder Colorado, Geol. Soc. Am. Mem.*, 195, 19-41, 2002.
- Hillhouse, J. W., Paleomagnetic methods, in *Geophysical Framework of the Continental United States*, edited by L. C. Pakiser and W. D. Mooney, *Geol. Soc. Am. Mem.* 172, 61-70, 1989.
- Hillhouse, J. W., D. M. Miller, and B. Turrin, Correlation of the Miocene Peach Springs Tuff with the Geomagnetic Polarity Time Scale and new constraints on tectonic rotations in the Mojave Desert, California: This volume.
- Hoffman, K. A., Dipolar reversal states of the geomagnetic field and core-mantle dynamics, *Nature*, 359, 789-794, 1992.
- Hopson, R. F., J. W. Hillhouse, and K. A. Howard, Dike orientations in the Late Jurassic Independence Dike Swarm and implications for vertical-axis tectonic rotations in eastern California: *Geol. Soc. Am. Spec. Pap.*, 438, 2008.
- Jachens, R. C., V. E. Langenheim, and J. C. Matti, Relationship of the 1999 Hector Mine and 1992 Landers fault ruptures to offsets on Neogene faults and distribution of late Cenozoic basins in the eastern California shear zone, *Bull. Seismo. Soc. Am.*, 92, 1592-1605, 2002.
- James, E. W., Southern extension of the Independence dike swarm of eastern California, *Geology*, 17, 587-590, 1989.
- Kirschvink, J. L., The least-squares line and plane and the analysis of palaeomagnetic data, *Geophys. J. R. Astron. Soc.*, 62, 699-718, 1980.
- Laj, C., A. Mazaud, R. Weeks, M. Fuller, and E. Herrero-Bervera, Statistical assessment of the preferred longitudinal bands for recent geomagnetic field reversal records, *Geophys. Res. Lett.*, 19, 2003-2006, 1992.
- Leonhardt, R., J. Matzka, F. Hufenbecher, and H. C. Soffel, A reversal of the earth's magnetic field recorded in mid-Miocene lava flows of Gran Canaria: Paleodirections, *J. Geophys. Res.*, 107(B1), 10.1029/2001JB000322, 2002.
- Lourens, L., F. Hilgen, N. J. Shackleton, J. Laskar, and D. Wilson, The Neogene Period, in *A Geologic Time Scale 2004*, eds. F. M. Gradstein, J. G. Ogg, and A. G. Smith, Cambridge University Press, 409-452, 2004.
- Love, J. J., Paleomagnetic volcanic data and geometric regularity of reversals and excursions, *J. Geophys. Res.*, 103, 12,435-12,452, 1998.
- Luyendyk, B. P., M. J. Kamerling, R. Terres, and J. S. Hornafus, Geometric model for Neogene crustal rotations in southern California, *Geol. Soc. Am. Bull.*, 91, 211-217, 1980.
- Mankinen, E. A., A reversal of geomagnetic polarity, *Phys. Bull.*, 37, no. 7, 287-288, 1986.
- Mankinen, E. A., M. Prevot, C. S. Gromme, and R. Coe, The Steens Mountain (Oregon) geomagnetic polarity transition, 1, Directional history, duration of episodes, and rock magnetism, *J. Geophys. Res.*, 90, 10,393-10,416, 1985.
- Merrill, R. T., and M. W. McElhinny, *The Earth's magnetic field*, Academic Press, New York, 401 p., 1983.
- Merrill, R. T. and P. L. McFadden, Geomagnetic polarity transitions, *Rev. Geophys.*, 37, 201-226, 1999.
- Miller, J. S., and J. D. Walker, Mesozoic geologic evolution of Alvord Mountain, central Mojave Desert, California, *Geol. Soc. Am. Mem.*, 195, 59-77, 2002.
- Nason, G. W., T. E. Davis, and R. J. Stull, Cenozoic volcanism in the Newberry Mountains, San Bernardino County, California, in *Cenozoic Paleogeography of the Western United States*, edited by J. M. Armentrout, M. R. Cole, and H. Terbest, Jr., *Soc. Econ. Paleontologists and Mineralogists, Pacific Coast Paleogeography Symposium 3*, 89-95, 1979.
- Nielson, J. E., D. R. Lux, G. B. Dalrymple, and A. F. Glazner, Age of the Peach Springs Tuff, southeastern California and western Arizona, *J. Geophys. Res.*, 95, 571-580, 1990.

- Ogg, J. G., and A. G. Smith, The geomagnetic polarity time scale, in A Geologic Time Scale 2004, edited by F. M. Gradstein, J. G. Ogg, and A. G. Smith, Cambridge University Press, 63-86, 2004.
- Prevot, M., and P. Camps, Absence of preferred longitudinal sectors for pole from volcanic records of geomagnetic reversals, *Nature*, 366, 53-57, 1993.
- Renne, P. R., A. L. Deino, R. C. Walter, et al., Intercalibration of astronomical and radioisotopic time, *Geology*, 22, 783-786, 1994.
- Ron, H., and A. Nur, Vertical axis rotations in the Mojave: Evidence from the Independence dike swarm, *Geology*, 24, 973-976, 1996.
- Ross, T. M., North-south-directed extension, timing of extension, and vertical-axis rotation of the southwest Cady Mountains, Mojave desert, California, *Geol. Soc. Am. Bull.*, 107, 793-811, 1995.
- Ross, T. M., B. P. Luyendyk, and R. B. Haston, Paleomagnetic evidence for Neogene clockwise tectonic rotations in the central Mojave Desert, California, *Geology*, 17, 470-473, 1989.
- Sauber, J., W. Thatcher, S.C. Solomon, and M. Lisowsky, Geodetic slip rate for the Eastern California Shear Zone and the recurrence time of Mojave Desert earthquakes, *Nature*, 367, p. 264-266, 1994.
- Savage, J. C., J. L. Svarc, and W. H. Prescott, Interseismic strain and rotation rates in the northeast Mojave domain, eastern California, *J. Geophys. Res.*, 109, B020406, doi:10.29/2003JB002705, 2004.
- Schermer, E. R., B. P. Luyendyk, and S. Cisowski, Late Cenozoic structure and tectonics of the northern Mojave Desert, *Tectonics*, 15, 905-932, 1996.
- Svarc, J. L., J. C. Savage and W. H. Prescott, Interseismic strain and rotation rates in the northeast Mojave domain, eastern California, American Geophysical Union, 2002 Fall Meeting, Program and Abstracts, 2002.
- Valentine, M. J., L. L. Brown, and M. P. Golombek, Cenozoic crustal rotations in the Mojave Desert from paleomagnetic studies around Barstow, California, *Tectonics*, 12, 666-677, 1993.
- Walker, J. D., J. M. Bartley, and A. F. Glazner, Large-magnitude Miocene extension in the central Mojave Desert: Implications for Paleozoic to Tertiary paleogeography and tectonics: *J. Geophys. Res.*, 95, 557-569, 1990.
- Wells, R. E., and J. W. Hillhouse, Paleomagnetism and tectonic rotation of the lower Miocene Peach Springs Tuff: Colorado Plateau, Arizona, to Barstow, California, *Geol. Soc. Am. Bull.*, 101, 846-863, 1989.
- Young, R. A., and W. J. Brennan, Peach Springs Tuff: Its bearing on the structural evolution of the Colorado Plateau and development of Cenozoic drainage in Mohave County, Arizona, *Geol. Soc. Am. Bull.*, 85, 83-90, 1974.

Geology and tectonic development of Alvord Mountain, Mojave Desert, CA

John H. Foster, Ph.D.

*Professor of Geology, Dept. of Geological Sciences, California State University, Fullerton, Fullerton, CA 92834,
jfoster@fullerton.edu*

Abstract

Alvord Mountain is a small cluster of hills capped by the basaltic pile now forming Alvord Peak in the central Mojave Desert. The Alvord Mountain basement rock is composed of a complex terrane of gneissic rocks as well as marbles and pelitic schists sharply folded and intruded by massive quartz monzodiorite and cut by a large swarm of quartz monzonite dikes. Miocene extension in the central Mojave Desert formed a series of basins within the granitic, metasedimentary, and gneissic terrane. Clews Basin was the easternmost of these formed as an intra-hanging wall basin on the upper plate of the east-dipping Central Mojave detachment fault. It became the depocenter for a variety of alluvial, fluvial and lacustrine sediments of Early Miocene age that coarsened upward into a tuffaceous and basaltic suite of rocks in the Middle Miocene. The youngest of the three units distinguished in Clews Basin is the thick sedimentary alluvial and fluvial fossiliferous and tuffaceous sediments of the Barstow Formation. The Barstow Formation is locally capped by a voluminous series of basalt flows. A cover of granitic coarse gravel caps the region. Pliocene contraction of the central Mojave Desert began along northwest-trending right-lateral strike-slip faults as well as a domain of clockwise-rotated blocks separated by east-striking left-lateral faults. The Alvord Mountain block formed by a bend in the block which caused Alvord Mountain to be uplifted and Coyote Lake to subside as intrablock nonrigid deformation domed up the small Alvord Mountain center and compressed and folded the sedimentary sequence into the Spanish Canyon anticline.

Introduction

Alvord Mountain is an isolated range of low hills north of the Mojave River between Barstow and Baker north of Interstate 15 in San Bernardino County, CA (see Fig. 1). The small range is northeast of Coyote Lake and north of Manix Lake beds. The central drainage of the hills is Spanish Canyon and it was down Spanish Canyon that emigrants from the Mormon settlements came to follow the Mojave River south into Cajon Valley and to the settlements in San Bernardino Valley in the mid to late 1800s. Indeed, the old wagon tracks are still preserved in the north rampart of Spanish Canyon where the trail descends from the area of Fort Irwin into Spanish Canyon wash. Rocks of igneous, metamorphic and sedimentary types are exposed beneath a thick gravel cover now eroded into a set of bluffs around the northern and eastern rim of the mountains. These rocks easily divide themselves into a southwestern igneous and metamorphic complex and a northeastern sedimentary suite with a central massif

of basaltic volcanic rock. This basaltic pile forms the highest peak in the mountain complex, called Alvord Peak, at 3,416 feet above sea level (see Fig. 2). The sedimentary complex is folded into an open anticline whose axis plunges steeply northwest in the area northeast of Alvord Peak. California State University Fullerton conducted its geologic field camp in the Alvord Mountain area during its winter break for several years in the early 1990s because of the many rock types, sedimentary environments, and structural complexity of the area. The area is shown on figure 3 and pertinent locations are noted there.

Basement rocks

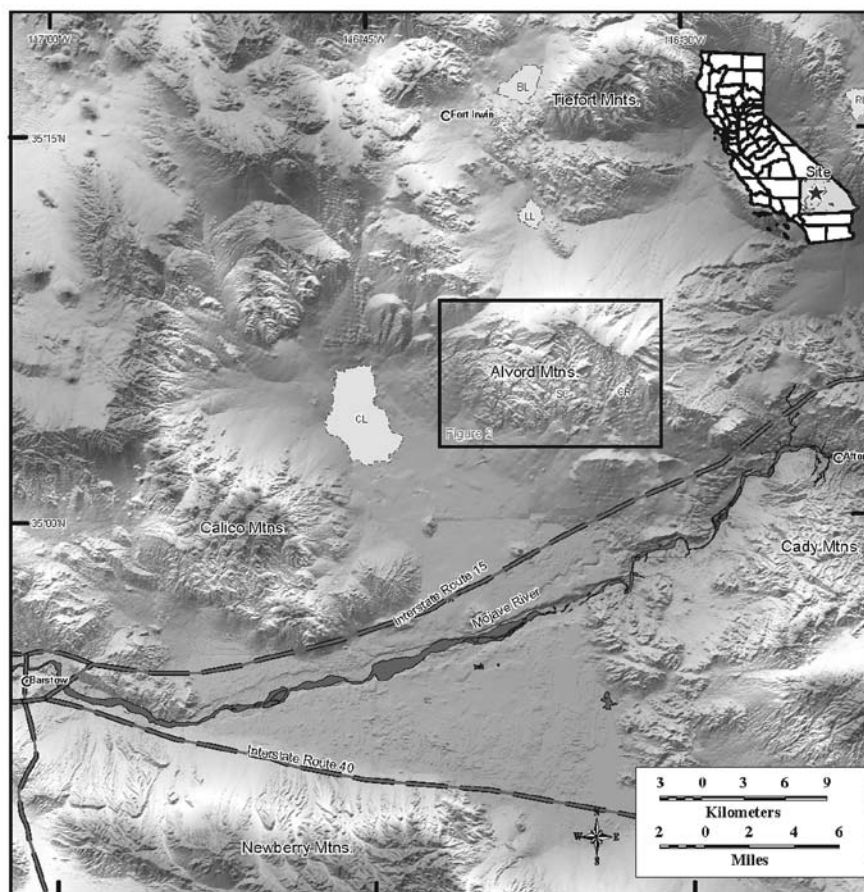
Igneous and metamorphic bedrock crops out extensively in the southwestern region of Alvord Mountain as a set of well-exposed linear belts (see Fig. 4A). Most of the igneous rock is composed of Jurassic quartz monzo-

nite and monzodiorite while the metamorphic rock is dominated by biotite schist (see Fig. 4B,C) and marble with Paleozoic-Mesozoic(?) sedimentary protoliths, (Miller and Walker, 2002). An excellent discussion of the evolution of the Mesozoic basement rocks of Alvord Mountain is contained in the article by Miller and Walker (2002). The summation below is primarily based on their work.

The above authors correlate the metasedimentary rocks to either the displaced Antler allochthon of Paleozoic age or the uppermost Paleozoic to lower Mesozoic parautochthonous borderland basin rocks. They conclude that the marble sequence that structurally overlies the schists is borderland-basin strata. Deformation of the metasedimentary rocks into an isoclinally folded belt is bracketed between the intrusion of a quartz monzodiorite of regional extent at around 179 Ma and undeformed gabbro and diorite dikes and small plutons dated at 149 Ma. Much later late Cretaceous intrusions created a thermal overprint on the originally upper greenschist and lower amphibolites facies metasediments. At this time, a large north-west trending set of granodiorite porphyry dikes formed which cut all pre-Tertiary rocks and that is part of the Late Cretaceous dike swarm of the central Mojave Desert, which is roughly parallel to the late Jurassic Independence dike swarm. The style and timing of deformation in the Alvord Mountain area is correlated to the Jurassic-Cretaceous East Sierran thrust system (Miller and Walker, 2002).

Sedimentary rock sequence

Five formation-level deposits have been identified from the Spanish Canyon sedimentary sequence including the Clews Formation, the "two tuffs", the Spanish Canyon Formation, the Barstow Formation, and the Alvord Peak Basalt (Byers, 1960). Above these formations is a thick cover of gravelly alluvium. A dissected terrace deposit caps ridges surrounding Spanish Canyon but occurs



A shaded relief map showing approximate location of the Alvord Mountains. Map also shows areas covered by Figure 1-2 and Figure 1-4. Abbreviated Localities: BL=Bicycle Lake, CL=Coyote Lake, CR=Clew's Ridge, LL=Langford Well Lake, SC=Spanish Canyon, RL=Red Pass Lake. Inset boxes show approximate locations of Figures 2.



Figure 1:
Location Map

below and is confined within the bluffs formed by the gravelly alluvium formation. Although Byers (1960) considered the Alvord Peak Basalt complex to be above the Clews Fonglomerate (his terminology) and below the Spanish Canyon Formation, it was later shown by Fillmore (1993) to be within or overlying the Barstow Formation, although this relationship is difficult to define in the field. In addition, Fillmore (1993) has removed the "two tuffs" from Byers (1960) original description of the Clews Fonglomerate, separating them into an overlying thin but distinctive unit that is conformable on the Clews Formation.

Clews Basin

Clews Basin is one of three basins named by Fillmore et al. (1994) and is described by them as an intra-hanging wall basin formed as the easternmost basin on the upper plate of the Central Mojave Detachment Fault System,

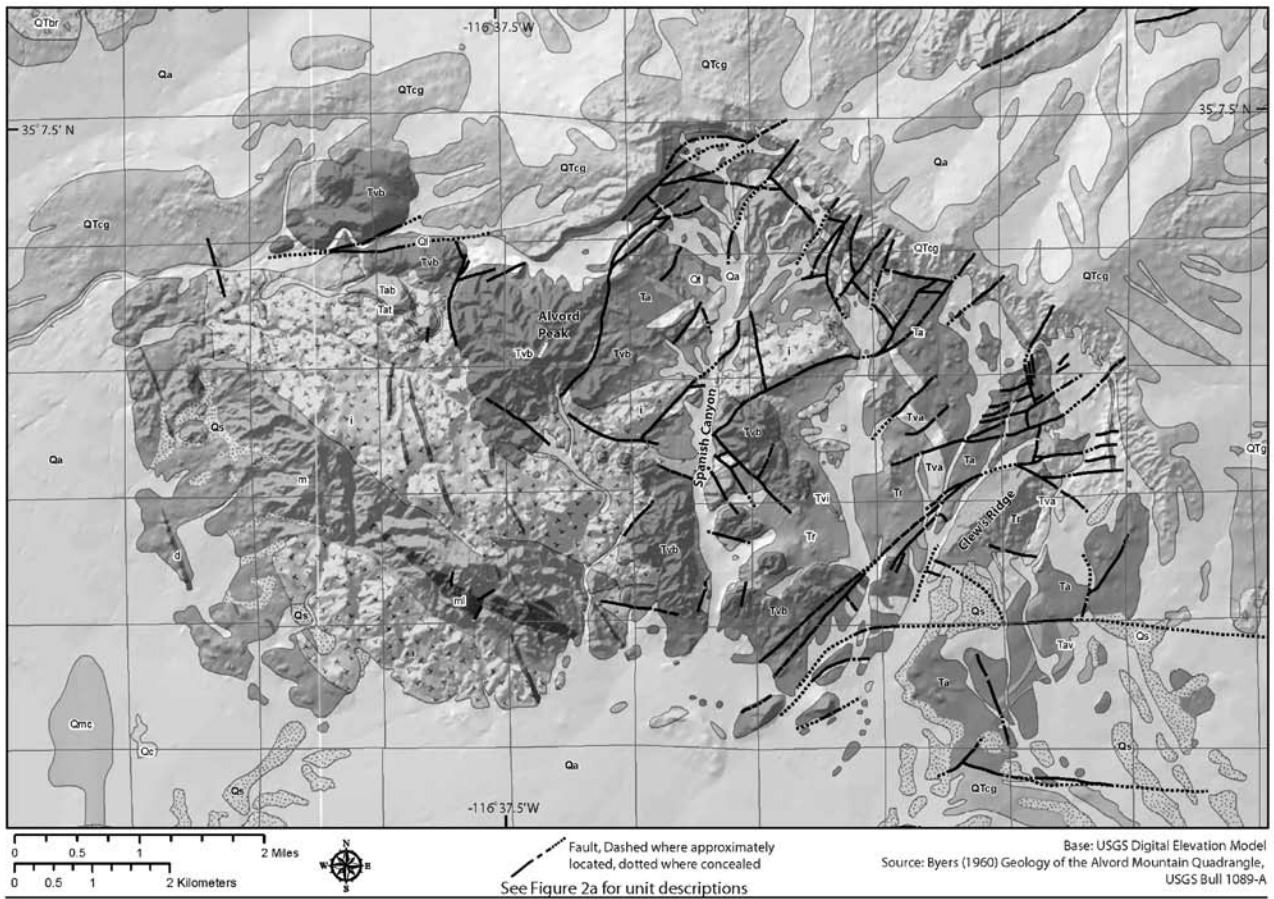


Figure 2:
Geologic Map of the Alvord Mountains

Figure 2a: Explanation for Geologic Map (Fig. 2) from Byers (1960). Units are shown as described by Byers (1960) with different interpretations explained in the text. In particular, the age of the Alvord Peak Basalt is more likely middle to late Miocene and the Clews Formation is separated in later publications from the uppermost sequence of tuffs including the Peach Spring Tuff and described as the "two tuffs" unit. Also, the QTcg unit is most likely wholly Pliocene or even Mio-Pliocene.

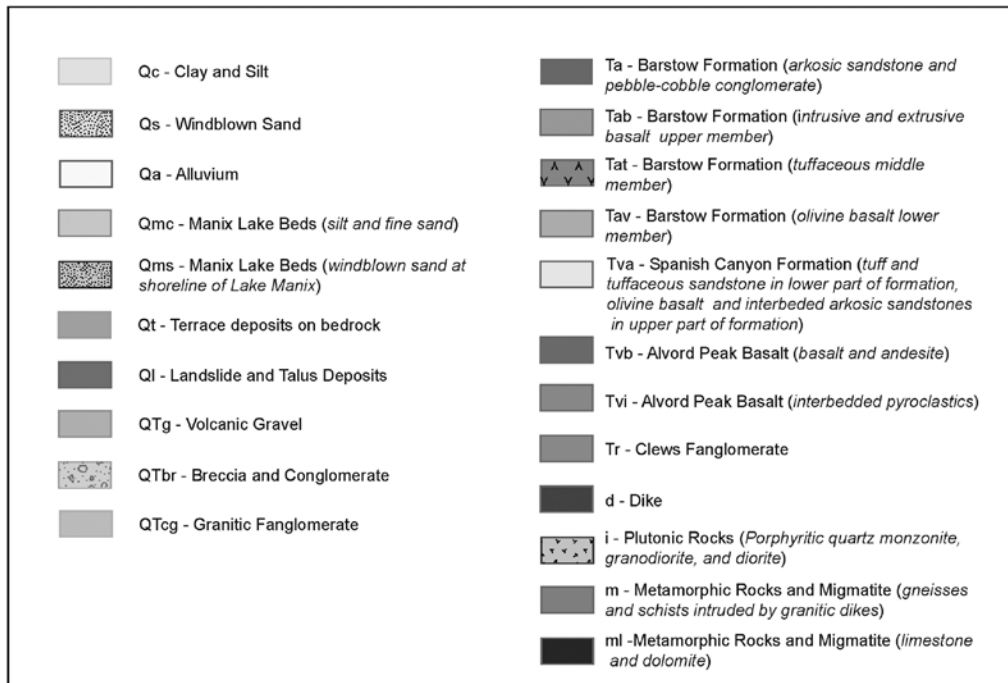


Figure 2a:
Geologic Unit Descriptions



Figure 3. Alvord Mountain viewed to SE; Mojave River in upper right.

which formed by northeastward extension starting in the Early Miocene. Clews Basin therefore is a structural basin with continental deposits in it that formed contemporaneously with sediments in the other basins to the west including the Pickhandle and Tropic Basins north of Barstow in the Calico Mountains and Gravel Hills respectively. This crustal extension thinned the crust and created

a pathway for voluminous amounts of basaltic lava to inject into the sediments and to form a central volcanic complex in each of the basins. Additionally, the northeastern limit of Clews Basin deepened as extension progressed and the Cronese Hills segment of the detachment upper plate broke away, forming an escarpment from which rapidly progressing alluvial fans grew southwestward. Fillmore (1993) describes the sedimentary sequence that is found in Clews Basin, and, in particular, describes the Clews Formation, the oldest sedimentary unit, as an Early Miocene synextensional continental deposit. Ross et al. (1989) define two episodes of clockwise rotation in

the central Mojave Desert. The earlier of these occurred in late Oligocene to early Miocene time and was confined primarily to rocks south and west of the Alvord Mountain and northern Cady Mountain block. This rotational episode occurred prior to the deposition of the Peach Spring Tuff in the central Mojave Desert. Additionally, Ross et

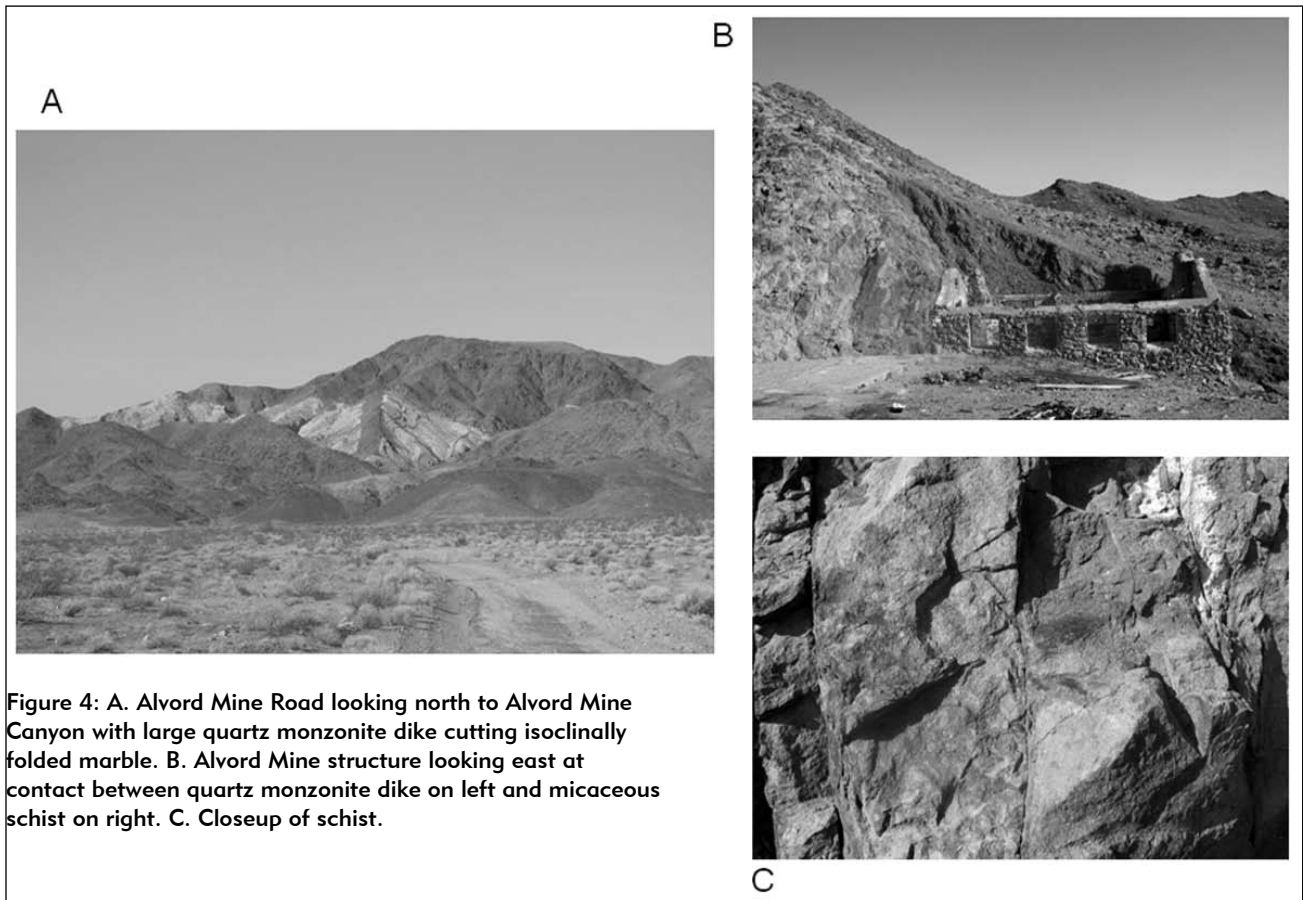


Figure 4: A. Alvord Mine Road looking north to Alvord Mine Canyon with large quartz monzonite dike cutting isoclinally folded marble. B. Alvord Mine structure looking east at contact between quartz monzonite dike on left and micaceous schist. C. Closeup of schist.

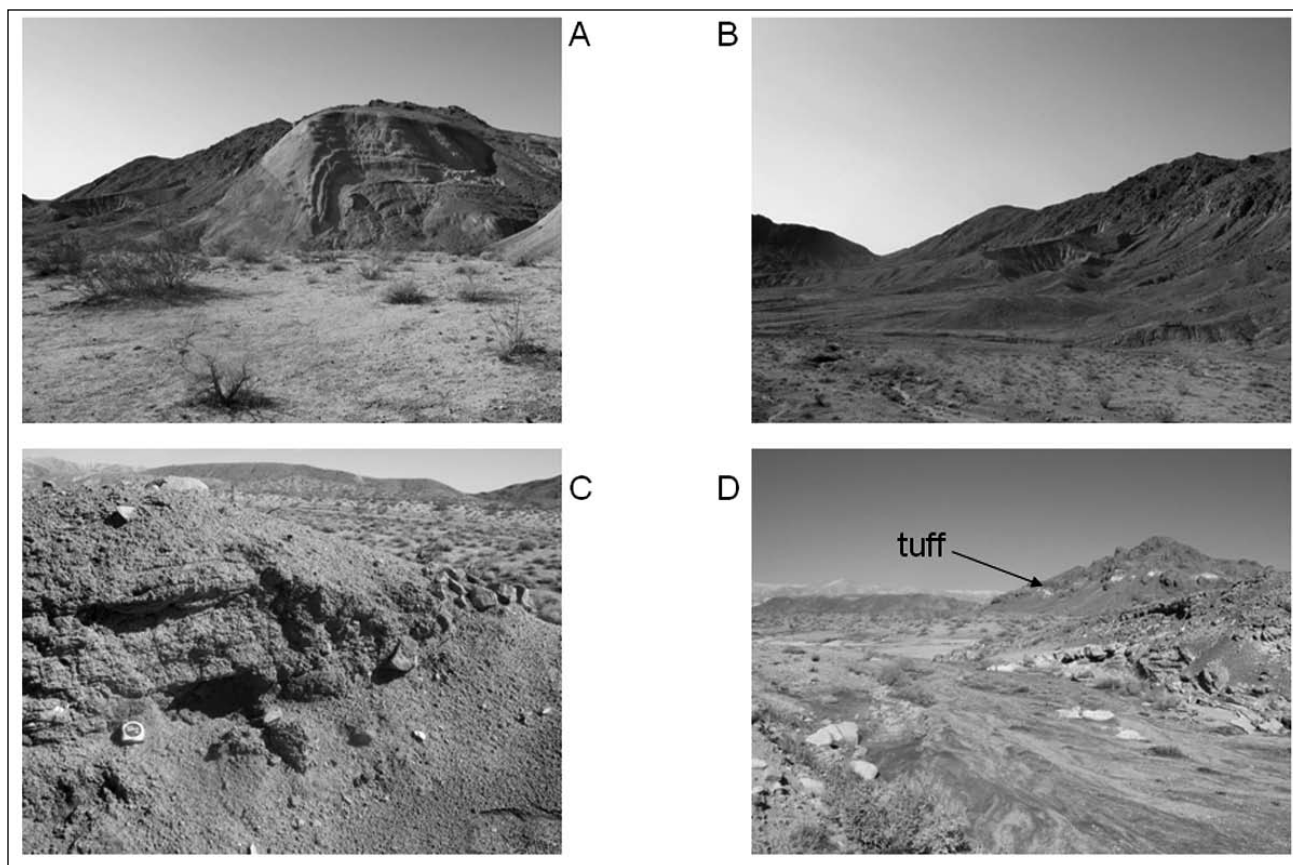


Figure 5: Clews Formation: A. Faulted and folded Clews lacustrine sediments: eastern area canyon. B: Clews Formation faulted against quartz monzodiorite in central Alvord Mountain region; view is west and in distance of A. C. Clews Formation in Spanish Canyon. D. Clews Formation lacustrine units in eastern canyon near location of A and B above, view is north with Peach Spring Tuff in background.

al. show that a later rotational event caused a clockwise rotation of 50° of the Alvord Mountain area as east-trending faults became active around 6 Ma. According to Miller and Walker (2002), the mid-Miocene and later extensions do not appear to have reoriented Mesozoic structures in the Alvord Mountain area, although they allow for some rotation to have occurred and are uncertain whether the data truly restrict such a rotational model. McFadden et. al. (1990) likewise suggest that little if any rotation has occurred in the Mud Hills.

The Clews Formation is an upward-coarsening continental deposit. Initial sedimentation formed in a depocenter where limestone, siltstone, and mudstone beds of lacustrine origin were deposited (exposed now mostly in the central portion of the anticline east of Spanish Canyon); see Fig. 5A,B,D. A local source of streams feeding the paleolake is indicated by conglomerates along the western margin of the anticline. These formed as small sheetflood-dominated alluvial fans that contain mostly granitic clasts similar to those in the Alvord Mountain basement to the west, indicating southeast-directed paleoflow (Fillmore, 1973). A more distant stream source is recognized from pebbly sandstones exposed in the

northern region of Spanish Canyon at the nose of the anticline where Clews Formation sediments contain metasedimentary rocks found in the Paradise Range north and northwest of Alvord Mountain (Fig. 5C). The upper Clews Formation formed as the Alvord Mountain–Cronese Hills block broke apart along a northwest-striking, southwest-dipping normal fault. The basin began to fill rapidly with coarse alluvial fan conglomerates and breccias including massive monolithologic debris flow deposits derived from the Cronese Hills (Fillmore, 1993). The Clews Formation is thickest in the east and overall approaches 312 meters thick at Clews Ridge, but is only 45 meters thick on the west in Spanish Canyon (see Fig. 1). Megabreccias associated with large runout debris flows are described by Fillmore (1993) from the area near Clew’s Ridge. The age of the Clews Formation is only roughly estimated in that it is overlain by what Fillmore (1993) calls the “two tuffs,” the upper of which is described as the Peach Spring Tuff dated at 18.5 ± 0.2 Ma by Nielson et. al. (1990). No fossils have been found within the Clews Formation (Woodburne, 1991).

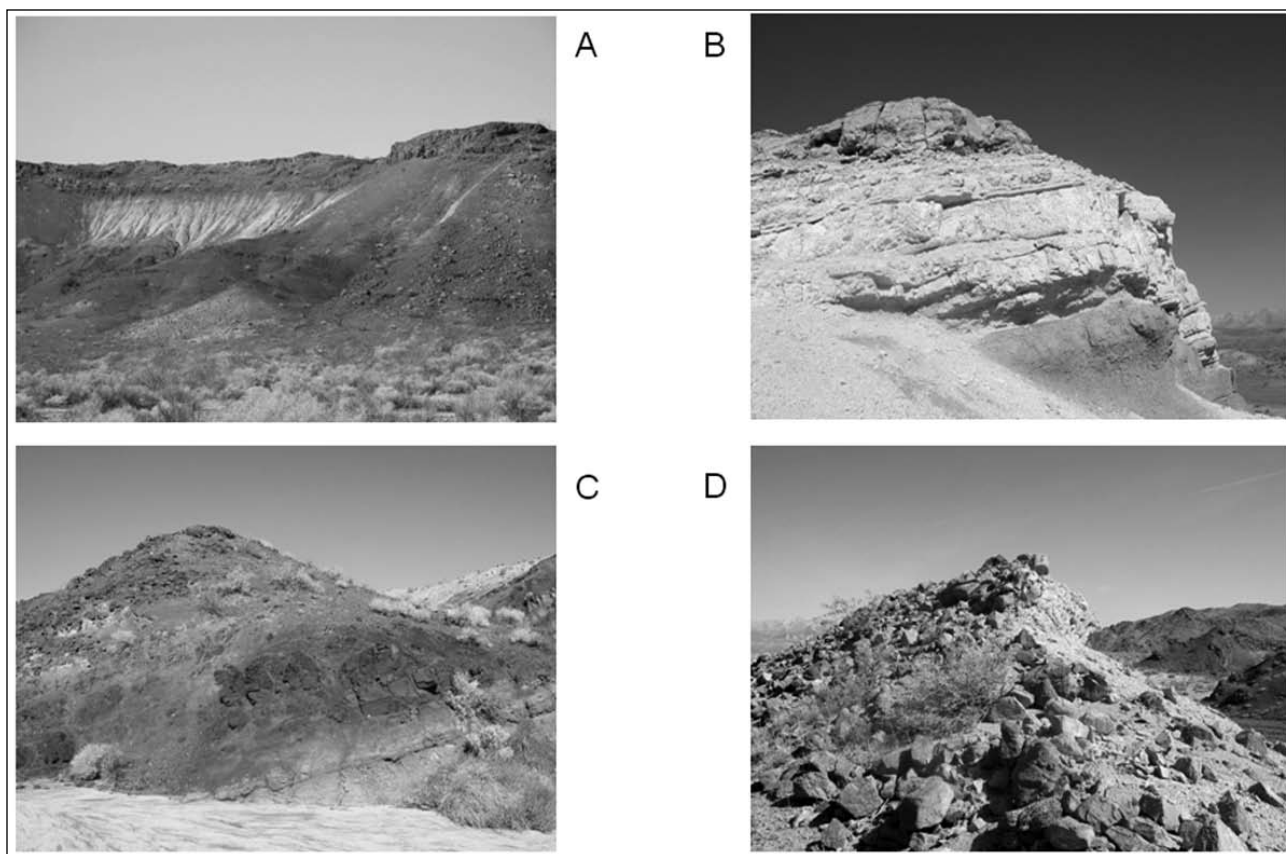


Figure 6: A. Spanish Canyon Formation basalts in northern Spanish Canyon. B. Spanish Canyon Formation lower gray and white tuffs. C. Spanish Canyon Formation uppermost basalt on baked sandstone, Barstow Formation above. D. Spanish Canyon Formation monolithologic breccia of quartz monzodiorite.

Spanish Canyon Formation

The Spanish Canyon Formation, 55 meters thick, is an easily recognized tuffaceous sandstone with a very white thick tuff near the base and two olivine basalt flows at the top. The uppermost basalt defines the top of the formation (see Fig. 6A, B and C) and has been dated by Dokka and Woodburne (1986) at 20.3 ± 0.8 Ma by whole rock K–Ar analysis. This age is in conflict with the age of the underlying Peach Spring Tuff. According to Byers (1960), the basalts are extra basinal in origin given their widespread constant thickness and ophitic texture, indicating a highly fluid magma flow. The basalt flows appear to be similar in petrology and texture to basalts higher in the Barstow Formation and are associated with pebble conglomerate whose clasts are from the Paradise Range. It seems most likely then that the basalts are derived from vents to the northwest and flowed rapidly southeast into Clews Basin.

Of particular interest is a monolithologic breccia composed of quartz monzonite and arkosic coarse sand and pebbles that lies between the two upper basalts. This unit was noted in the measured section of Spanish Canyon by Byers (1960). The breccia is a loose accumulation of yellowish gray boulders and cobbles of quartz

monzonite with a rare boulder of dark gray diorite that are mostly rectangular shaped or cubic in form and which are similar to the quartz monzodiorite and diorite dike rocks from the basement rocks to the southwest (Fig 6D). This unit crops out solely in the type section west of Spanish Canyon in Section 30 R4E T12N. The deposit represents a debris flow from a highland to the southwest into the fluvial low lying basin in which the finer grained sediments, tuffs, and basalts were being deposited.

The age of the Spanish Canyon formation is uncertain given the out-of-sequence age reported for the upper basalt. However the unit lies above the Peach Spring Tuff and conformably underlies the Barstow Formation, which has yielded a Miocene horse, *Paraplihippus carrizoensis*, from 150 meters above the Spanish Canyon top. This species does not extend younger than the late Hemingfordian LMA, about 16 Ma. Woodburne et.al. (1982) have reported a date on the lapilli tuff lying in the upper portion of the formation of 13.8 and 14.0 Ma (K–Ar). The Spanish Canyon Formation is generally Middle Miocene and no younger than late Hemingfordian in age.

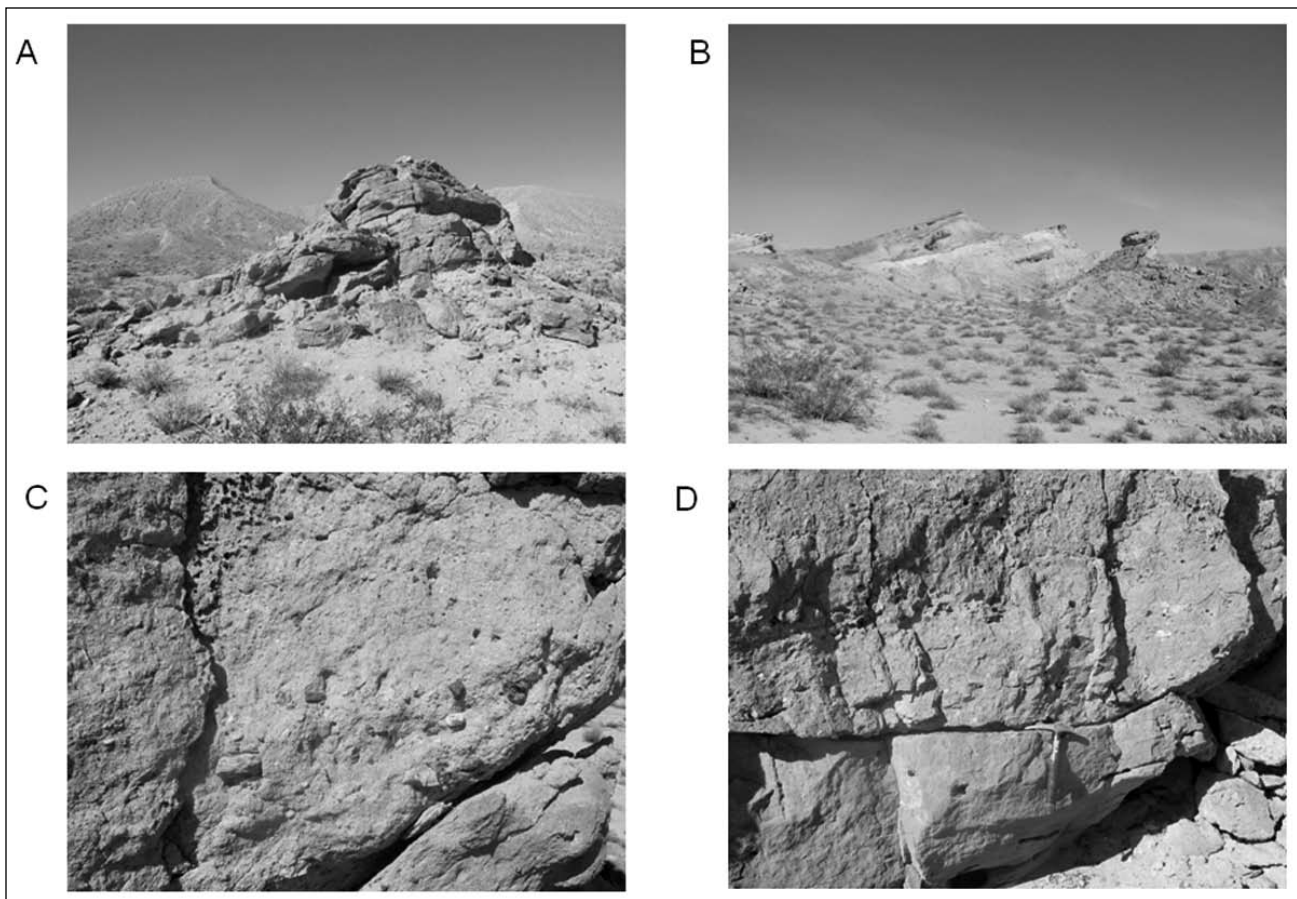


Figure 7: A and B are outcrops of paleochannel deposits in lower Barstow Formation: C and D are closeups of channel deposits.

Barstow Formation

The Barstow Formation in the Alvord Mountain hills is 300 meters thick and divisible into a lower, middle, and upper member (Byers, 1960). The lower Barstow contains three basalt layers of intrusive and extrusive origin (Byers, 1960) and is primarily composed of conglomerate lenses and massive sandstone beds which in the area of upper Spanish Canyon reveal a distinct set of very large paleochannel deposits (see Fig. 7A and B; Fig. 7C and D are close-up images of the channel deposits). The formation contains an 8-meter-thick distinctive tuffaceous silt to fine sand interval that contains well- rounded pebble-size white tuff clasts. This “lapilli tuff” unit is dated at around 13 to 14 Ma by Woodburne et.al. (1982) and is the middle member of the formation. The upper member consists of tuffaceous sandstone and contains an andesite cobble conglomerate that is up to 30 meters thick. The uppermost Barstow Formation contains fewer volcanic clasts and becomes coarser as it grades into the overlying gravel fanglomerate (Byers, 1960). Large paleochannels indicate that the lower Barstow Formation basin was formed in part by a large braided river system much like the Mojave River today and the cobbles that are found in

the channels appear to be more like the metasediments found north and northwest of Clews basin. This is similar to the paleocurrent directions noted for the Mud Hills Barstow Formation sediments as reported by Woodburne et. al. (1990).

Alvord Peak Basalt

The Alvord Peak Basalt is a massive olivine basalt, the source of which is local to the Alvord Mountain hills. Vertical basalt-filled pipes and vents areas can be seen south and east of Spanish Canyon in Sec 32, T12N, R4E. The basaltic mass under Alvord Peak is composed of many separate flows; in places the flows are associated with volcanoclastic sediments bearing blocks and bombs of basalt as well as autoclastic basaltic breccias (Byers, 1960) (fig. 8). The basaltic center contains weathered and replaced olivine and amygdules of opal and calcite within the vesicular portions of some flows. The basalt is 150 meters thick at Alvord Mountain but absent from the region at and north of Clews Ridge and east of the axis of Spanish Canyon Anticline. Basalt is primarily present at the mouth of Spanish Canyon and unevenly covers 3 km on a line striking north 45° west. It thickens to the north toward the area of Alvord Peak. Little to no Alvord Peak



Figure 8: Alvord Peak Basalt at Alvord Peak and as flows on reddish autobreccia in central Spanish Canyon.

Basalt is present east of the axis of the anticline, meaning that some control must have been exerted on the flows by the structure of the anticline.

The age of the Alvord Peak Basalt is undetermined although its field relations suggest it is younger than the Barstow Formation or perhaps an upper member of the Barstow as shown by Fillmore (1993). Schermer et.al (1996) reported ages for Pliocene basalts and basaltic centers north of Alvord Mountain in the Bicycle Lake and Bitter Springs areas as 5.57 ± 0.26 Ma and 5.5 ± 0.2 Ma, respectively; Miller and Yount (2002) reported an age of 3.4 ± 0.5 Ma for a basalt in Coyote Canyon west of these areas. However, an age of 13 to 14 Ma would seem to be the youngest likely age for the Alvord Peak Basalt as it lies decidedly below the granitic fanglomerate described below but within the upper Barstow Formation; shown by McFadden et.al. (1990) to be capped by a lapilli tuff that is 13.4 ± 0.2 Ma in the Mud Hills. Strona and Miller (2002) however, consider the Alvord Peak Basalt to lie below the Spanish Canyon Formation. They also consider the Peach Spring Tuff to be a part of the Spanish Canyon Formation which would make the Alvord Basalt synextensional and older than 18.5 Ma. The field relationships of the Alvord Peak Basalt are difficult to pin down and its proper stratigraphic position will be best understood when a valid date is obtained.

Granitic Fanglomerate

The uppermost sedimentary unit in the Alvord Mountain hills is the Granitic Fanglomerate of Byers (1960), which appears to be conformable above the Barstow. The fan-

glomerate is exposed as a folded, gently north- and east-dipping accumulation of granitic boulder lenses and coarse sand and gravel deposits that are poorly indurated. The deposit is over 300 meters thick and is coarser in the northwest than it is in the southeast, which indicates a source to the northwest. It is significant that this unit contains few if any volcanic clasts but does contain distinctive red and green calc-silicate metasediments, thin bedded quartzites, and granitoids with muscovite dikes which are found in the Goldstone area and display paleocurrent orientations that indicate a source area to the northwest (Miller and Yount, 2002). Miller and Yount (2002) describe the gravel formation as the deposits of a trunk stream or long, broad alluvial fan that originated in the Goldstone area and flowed as far as 55 km to Bitter Spring, east of

Alvord Mountain. The fanglomerate appears thickest in the northwest and in places appears to be conformable on the Barstow Formation but thins and is unconformable over most lower units on the southeast side of the Spanish Canyon anticline. No fossils are known from it and its total thickness is unknown. However, it is Pliocene in age because it contains an ash dated by Miller and Yount (2002) as 3.4 ± 0.2 Ma. This thick deposit indicates that the Alvord Mountain area was a depocenter for fluvial deposits well into the Pliocene. The dip directions and alignment of the Spanish Canyon anticline continue upwards through the Fanglomerate, indicating that it was folded during the Spanish Canyon fold event.

Structural and tectonic development

Nearly continuous sedimentation from the Early Miocene into the Plio-Pleistocene is represented by a nearly 1,000 meter thick sedimentary and volcanic rock-filled Clew's basin of the Alvord Mountain area. Subsequent to the deposition of the thick and extensive granite fanglomerate, the Alvord Mountain complex was uplifted, folded, and faulted. Miller and Yount (2002) consider the Alvord Mountain area to be within the ESF (east-striking fault) domain for which Schermer et.al. (1996) earlier identified 2–6 km of left slip on each of the main east-striking faults. Tectonic blocks in the domain underwent a vertical rotation of $\sim 40^\circ$ as the Eastern California Shear Zone deformed. In the area of Alvord Mountain these east-striking faults include the Bicycle Lake fault and Coyote

Lake fault north of Alvord Mountain and the Manix fault south of Alvord Mountain.

Schermer et al. (1996) noted that faulting in this ESF domain took place after the 10–12 Ma volcanic episodes in the area. Miller and Yount (2002) state that folds and strike-slip thrusts are created through nonrigid behavior within blocks separated by east-trending faults and such is likely the case for the domical uplift of the Alvord Mountain block and the folding event for the Spanish Canyon anticline. Their estimation of principal stress orientations are that primary stress was horizontal along a 020° azimuth and a horizontal least stress axis oriented 110° (90° to the principal stress). This stress orientation is consistent with an axial plane of the Spanish Canyon anticline that strikes $N20^\circ$ – 50° W and for the domical uplift axis of Alvord Mountain to be $N50^\circ$ E. Spanish Canyon anticline is highly dissected by east-trending ($N60^\circ$ E) normal faults that are mostly south-side-down and along which some strike slip motion is likely. Since the gravel fanglomerate has been shown to contain a volcanic ash that is 3.4 ± 0.2 Ma and since it is incorporated in the Spanish Canyon Anticline, the folding and uplift events must be late Pliocene or younger.

Conclusions

The base of the Clews Formation is poorly dated and the development of the depocenter would be clarified with the discovery and identification of fossils. The volcanic sequence within the Alvord Mountain region is composed of petrologically similar sequences of basalt whose ages are inverse to stratigraphic position. The highest and most voluminous basalt, the Alvord Peak Basalt, remains poorly dated as late Miocene or Pliocene. The relationship of the overlying Pliocene granitic fanglomerate to the Alvord Peak Basalt and the Barstow Formation is poorly constrained yet is very important for understanding the tectonostratigraphic events of the Late Miocene through Pleistocene.

The extensional and subsequent contractional development of the Alvord Mountain block has created a well-exposed suite of rocks with important and complex structural and tectonic implications. The basement geology is structurally elongate northwest to southeast and the dike rocks that cut through this basement, if related to other dike swarms mapped nearby, indicate that the basement did not rotate. However, analyses of Tertiary sediments and the modeling of this structural terrane within the Mojave Block rotational and tectonic framework (Schermer et al, 1996) suggests clockwise rotation of 40° . In addition, the regional analysis of the volcanic rocks in the central Mojave Desert by Ross et al. (1989)

shows Pliocene rotation of 50° for the Alvord Mountain and northern Cady Mountain area. These matters are a high priority if the Alvord Mountain geology is to be of more use in the understanding of the evolution of the central Mojave Desert.

Acknowledgements

This paper has benefitted greatly from the thoughtful and detailed review by Dr. David M. Miller. In addition, I want to thank the Mojave Water Agency who have supported a great deal of research at Cal State Fullerton and who were supportive of my efforts here to consolidate and synthesize the information about the Alvord Mountain region. I also would like to thank and acknowledge Bob Reynolds for suggesting the idea for the paper and encouraging me to write it. Lastly, I want to thank Rene Perez for creating the maps and other illustrations for me. Without his help I would have been hard-pressed to complete the project.

References

- Byers, Jr., F.M., 1960, Geology of the Alvord Mountain Quadrangle, San Bernardino County, California: U.S. Geological Survey Bulletin 1089-A, 71 p.
- Dokka, R.F. and Woodburne, M.O., 1986, Mid-Tertiary extensional tectonics and sedimentation, central Mojave Desert, California: LSU Publ. Geol. Geophys. Tectonics and Sedimentation, 1, 55 p.
- Fillmore, Robert P., 1993, Sedimentation and extensional basin evolution in a Miocene metamorphic core complex setting, Alvord Mountain, central Mojave Desert, California, USA: *Sedimentology*, v. 40, p. 721-742.
- Fillmore, R. P., Walker, J.D., Bartley, J.M., and Glazner, A.F., 1994, Development of three genetically related basins associated with detachment-style faulting: predicted characteristics and an example from the central Mojave Desert, California: *Geology*, v. 22, p. 1087-1090.
- McFadden, B.J., Swisher III, C.C., Opdyke, N.D., and Woodburne, M.O., 1990, Paleomagnetism, geochronology, and possible tectonic rotation of the middle Miocene Barstow Formation, Mojave Desert, southern California: *Geological Society of America Bulletin*, v.102, p. 478-493.
- Miller, D. M. and Yount, J. C., 2002, Late Cenozoic tectonic evolution of the north-central Mojave Desert inferred from fault history and physiographic evolution of the Fort Irwin area, California: in Glazner, A. F., Walker, J. D., and Bartley, J. M., eds., 2002, *Geologic evolution of the Mojave Desert and southwestern Basin and Range*: Geological Society of America Memoir 195, p. 173-197.
- Miller, Jonathon, S. and Walker, J. Douglas, 2002, Mesozoic geologic evolution of Alvord Mountain, central Mojave Desert, California: in Glazner, A. F., Walker, J. D., and Bartley, J. M., eds., 2002, *Geologic evolution of the Mojave Desert and southwestern Basin and Range*: Geological Society of America Memoir 195, p. 59-77.
- Nielson, J.E., Lux, D.R., Dalrymple, G.B., and Glazner, A.F., 1990, Age of the Peach Spring Tuff, Southeastern California and Western Arizona: *J. Geophys. Res.*, v. 95, p. 571-580.
- Ross, T.M., Luyendyk, B.P., and Haston, R.B., 1989, Paleomagnetic evidence for Neogene clockwise tectonic rotation in the central Mojave Desert, California: *Geology*, v. 17, p. 470-473.

- Schermer, E.R., Luyendyk, B.K., and Cisowski, S., 1996, Late Cenozoic structure and tectonics of the northern Mojave Desert: *Tectonics*, v. 15, p. 700-710.
- Strona, J. and Miller, J., 2002, Stratigraphy and petrology of Miocene volcanic rocks at Alvord Mountain, Central Mojave Desert, California: Geological Society of America Abstracts with Programs, Cordilleran Section Meetings
- Woodburne, M.O., 1991, The Mojave Desert Province, in Woodburne, M.O., Reynolds, R.E., and Whistler, D.P., eds., *Inland southern California: the last 70 million years*: San Bernardino County Museum Association Quarterly, v. 38, p. 60-77.
- Woodburne, M.O., Miller, S.T., and Tedford, R.H., 1982, Stratigraphy and geochronology of Miocene strata in the central Mojave Desert: Geological Society of America Cordilleran Section Field Trip Guidebook, p. 47-64.
- Woodburne, M.O. and Swisher, C.C., III, 1995, Land Mammal high-resolution geochronology, Intercontinental overland dispersals, sea-level, climate, and vicariance: in Berggren, W.A, Kent, D.V., Aubry, M-P and Hardenbol J., eds., *Geochronology, time scales and global stratigraphic correlations: unified temporal framework for an Historical Geology*: SEPM Special Publication No. 54, p. 335-364.
- Woodburne, M.O., Tedford, R.H., and Swisher, C.C., III, 1990, Lithostratigraphy, biostratigraphy, and geochronology of the Barstow Formation, Mojave Desert, Southern California: Geological Society of America Bulletin, v. 102, p. 459-477

Camel tracks from the Early Miocene Hector Formation, Cady Mountains, California

Robert E. Reynolds

Redlands, CA 92373, rreynolds220@verizon.net

Abstract

The first vertebrate ichnofossils reported from the Cady Mountains consist of tracks of a small camel representing ichnospecies *Dizygopodium quadracordatum*. Tracks are preserved in water-laid volcanoclastic sediments and are associated with imprints of raindrops. The tracks are in a thick sedimentary sequence that dates to >20 Ma, and possibly represent a camelid ichnospecies from the early Hemingfordian North American Land Mammal Age of the early Miocene.

Geologic setting

The western, central, and eastern Cady Mountains and the south-central Cady Mountains contain early Miocene sediments and late Arikareean (Woodburne, 1998) and Hemingfordian fossil faunas (Moseley, 1978; Miller, 1980; Williamson, 1980). A tuff mapped in these deposits (Dibblee and Bassett, 1966b; Moseley, 1978; Miller, 1980; Williamson, 1980) has been identified as the 18.7 Ma Peach Spring Tuff (Hillhouse and others, this volume). The Cady Mountains contain dated volcanic rocks and faunas that overlap the age of deposition of the Barstow Formation, but apparently do not contain a sedimentary sequence similar to the Barstow Formation.

This suggests that the Cady sedimentary sections were deposited under different conditions or in a different depositional basin.

Stratigraphy

Imprints of camel tracks are preserved at the base of a white ash in water-laid volcanoclastic sediments in the northern Cady Mountains, south of Afton Canyon, in the central Mojave Desert. Geological mapping of the Hector Formation in the northern Cady Mountains indicates that the tracks are in sedimentary Unit Ts3 (Moseley, 1978), described as “green to brown tuffaceous lacustrine sandstone, siltstone and mudstone with interbedded fluvial conglomerate.” This unit lies between two basalts: the underlying (Tb2) is a non-vesicular olivine basalt, and the overlying (Tb3) is a vesicular basalt. The lower (Tb2) has several dates averaging 21.9 Ma (Moseley, 1978; Woodburne and Reynolds, this volume) and the upper (Tb3) has several discordant dates (Moseley, 1978) and is considered to be between 19 and 20 m.y. old (Woodburne, 1998). The basalts, in turn, are overlain by the 18.7 Ma Peach Spring Tuff (Moseley, 1978; Hillhouse and others, this volume). This stratigraphic sequence suggests that the volcanoclastic sediments containing the tracks date between 19 or 20 Ma and 21.9 Ma. This range of dates falls within the early Hemingfordian North American Land Mammal Age (NALMA; Tedford and others,

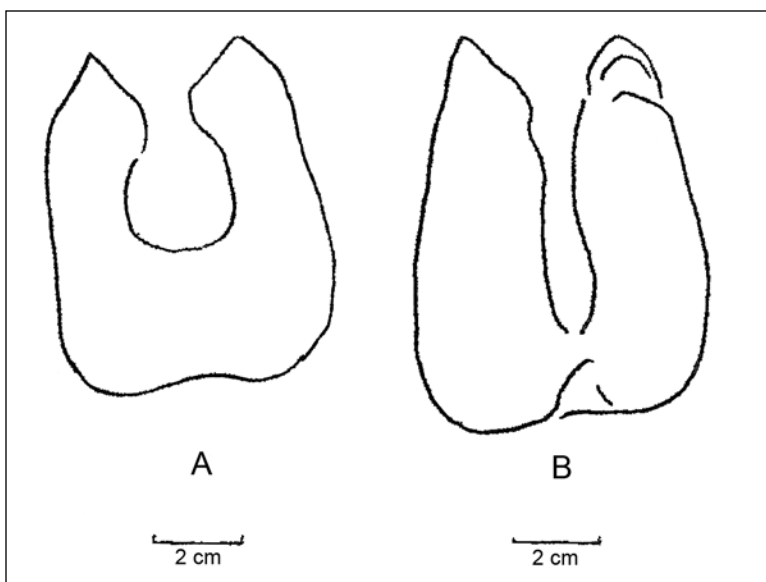


Figure 1. Drawing of the northern Cady Mountain camel tracks representing prints of the manus (A) and pes (B) of *Dizygopodium quadracordatum*.

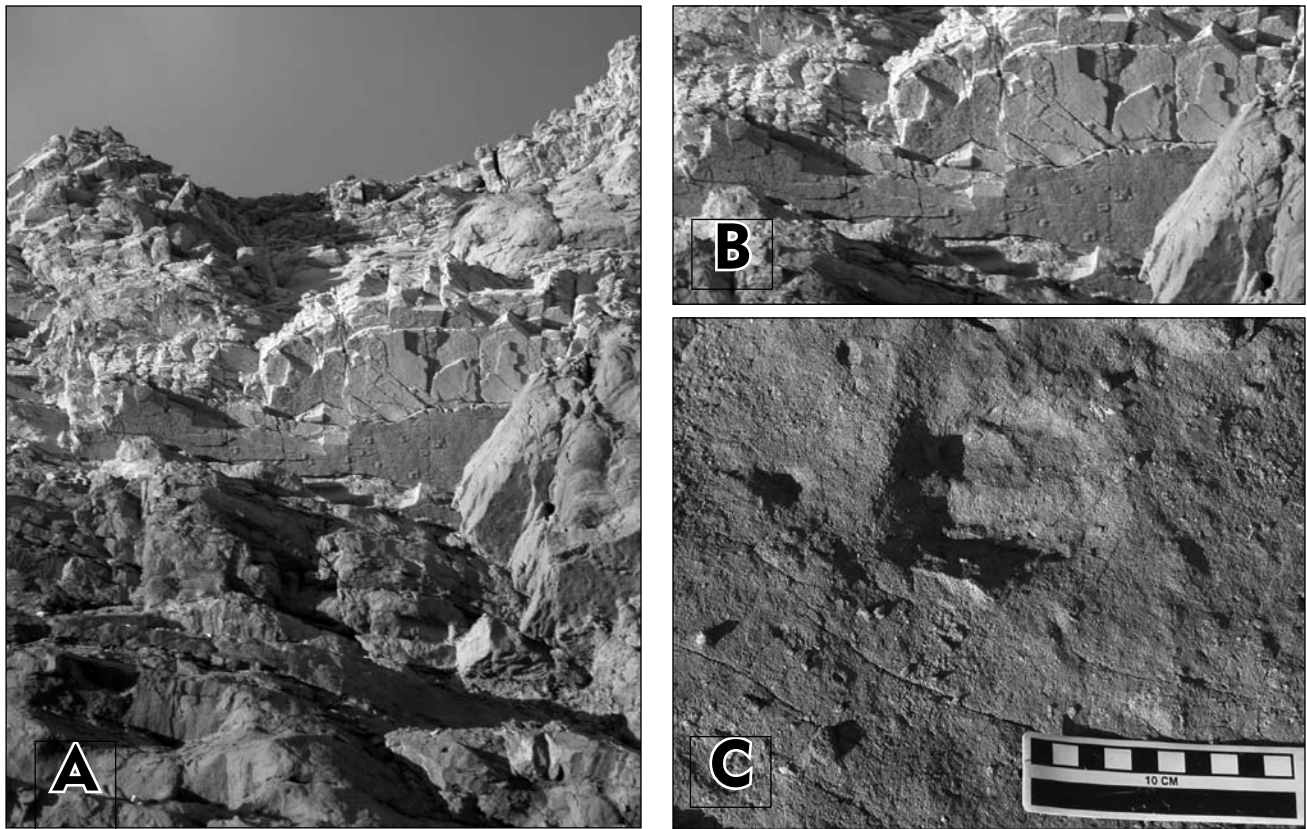


Figure 2. Camel tracks on overhang (2A and B, detail) and on a fallen block of sediment (C). R.E. Reynolds photographs.

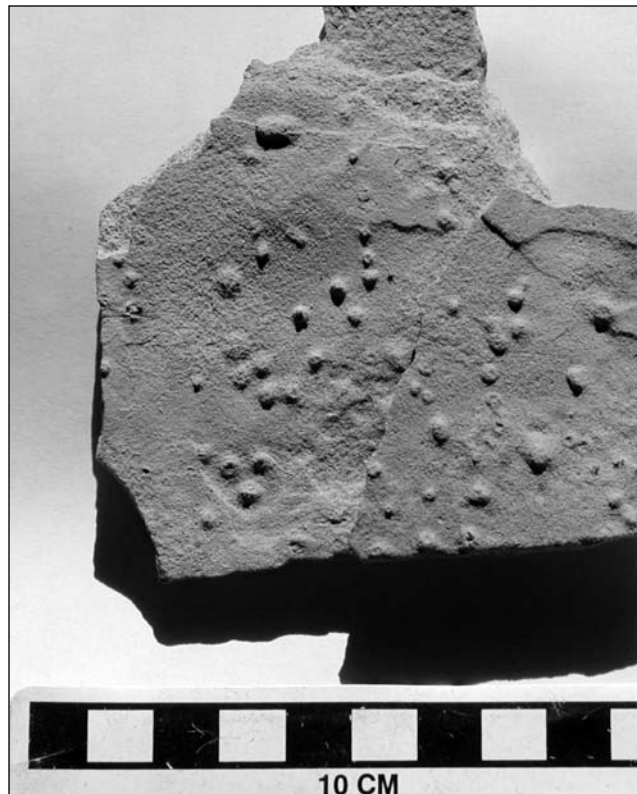


Figure 3. Raindrop impressions in fine-grained volcaniclastic sandstone. R.E. Reynolds photograph.

2004).

Description

The camel tracks from Unit Ts3 are small, measuring just 8.4 cm long and 6.5 cm wide. The tracks are quadrate and almost square, showing short, blunt claws in the manus, but claws in the pes are not apparent (Figs. 1, 2a and b). In the manus (Fig 1a), the anterior gap connects the anterior trough to a large medial pocket. In the pes, an interdigital furrow connects the interclavular gap to the proximal gap; there is no medial pocket. The morphology compares well to *Dizygopodium quadracordatum* (Fig. 17 a in Sarjeant and Reynolds, 1999); however, the size of this early Hemingfordian camel track is approximately 18 % smaller than the late middle Miocene Clarendonian NALMA figured in that report. The fine-grained volcaniclastic sandstone also contain imprints of rain drops (Fig. 3) and ripple marks, suggesting that the sediments were deposited in a ephemeral shallow lake or playa.

Associated faunas

The early Hemingfordian Lower Cady Mountains Local Fauna has not produced fossil bones of camels. The camels *Paramiolabis tenuis* (Kelly, 1992) and cf. *Aepycamelus* sp. are reported from the Upper Cady Mountains Local

Fauna, which is considered late Hemingfordian age or about 16–17 Ma (Woodburne, 1998; Woodburne and Reynolds, this volume) and therefore probably do not represent the makers of the camel tracks found lower in the section.

References cited

- Dibblee, T.W. Jr. and A.M. Bassett. 1966. Geologic map of the Cady Mountains quadrangle, San Bernardino County, California. U.S. Geological Survey Miscellaneous Geological Investigations Map I-467, 1:62,500.
- Hillhouse, J.W., D.M. Miller and B.D. Turrin. 2010. Correlation of the Miocene Peach Spring Tuff with the geomagnetic polarity time scale and new constraints on tectonic rotations in the Mojave Desert, California, *in* Overboard in the Mojave: 20 million years of lakes and wetlands, R.E. Reynolds and D.M. Miller (eds). California State University Fullerton, Desert Studies, this volume.
- Kelly, T.S. 1992. New middle Miocene camels from the Caliente Formation, Cuyama Valley badlands, California. *PaleoBios* 13(32):1-22.
- Miller, S.T. 1980. Geology and mammalian biostratigraphy of a part of the northern Cady Mountains, California. U.S. Geological Survey Open File Report 80-978:121p.
- Moseley, C.G. 1978. Geology of a portion of the northern Cady Mountains, Mojave Desert, California. University of California, Riverside, Department of Earth Sciences (Master's thesis).
- Sarjeant, W.A.S. and R. E. Reynolds. 1999. Camelid and horse footprints from the Miocene of California and Nevada, pp 3-20 in *Fossil Footprints*, R.E. Reynolds (ed). San Bernardino County Museum Association Quarterly, 46(2).
- Tedford, R.H., L.B. Albright, A.D. Barnosky, I. Ferrusquia-Villafranca, R.M. Hunt, Jr., J.E. Storer, C.C. Swisher, III, M.R. Voorhies, S.D. Webb, and D.P. Whistler. 2004. Mammalian biochronology of the Arikareean through Hemphillian interval (late Oligocene through early Pliocene epochs), pp. 169-231 in *Late Cretaceous and Cenozoic Mammals of North America: Biostratigraphy and Biochronology*, M.O. Woodburne, (ed.), Columbia University Press, New York.
- Williamson, D.A. 1980. The geology of a portion of the eastern Cady Mountains, Mojave Desert California. University of California, Riverside, Department of Earth Sciences (Master's thesis).
- Woodburne, M.O. 1991. The Mojave Desert province, pp 60-77 in *Inland Southern California: the last 70 million years*, M.O. Woodburne, R.E. Reynolds, and D.P. Whistler (eds). San Bernardino County Museum Quarterly 38(3&4).
- Woodburne, M.O. 1998. Arikareean and Hemingfordian faunas of the Cady Mountains, Mojave Desert province, California, pp 197-201 in *Depositional environments, lithostratigraphy, and biostratigraphy of the White River and Arikaree Gouups (late Eocene to early Miocene, North America*, D.O. Terry Jr., H.E. LaGarry, and R.M. Hunt Jr, (eds.), Geological Society of America Special Paper 325.
- Woodburne, M.O. and R.E. Reynolds. 2010. The mammalian litho- and biochronology of the Mojave Desert Province, *in* Overboard in the Mojave: 20 million years of lakes and wetlands, R.E. Reynolds and D.M. Miller (eds). California State University Fullerton, Desert Studies, this volume.

The Palm Borate Company operations, Calico

Larry M. Vredenburg

Bureau of Land Management, 3801 Pegasus Drive, Bakersfield, California 93308

Introduction

In 1991 Alan Hensher and I visited with Glen and Dorene Settle at their home at the Tropico Mine, just outside of Rosemond. During the course of our conversation the Settles mentioned that someone had dropped off a photo album with them, but they didn't know where the pictures were from. After looking through the album, Alan and I felt some of the photos were from the Calico area, but we weren't sure. Glen and Dorene allowed me to copy the photos in the album, and a short time later, with prints in hand, I visited Daggett and was able to confirm that some of the photos were taken there. But some of the most intriguing photos remained a mystery until I received an inquiry from Bob Reynolds in July 2006. I

supplied him with the photos and he was able to confirm the location of the mine railroad and mill shown in some of the photos. Recently, Mike Rauschkolb with Rio Tinto Minerals was able to identify the mine as the operations of The Palm Borate Company.

The Palm Borate Company operations

Much has been written on the borax deposits of Calico, but operations of the Palm Borate Company and its successor Borax Properties Limited are scarcely a footnote. The borate-bearing shale exposed in an unnamed canyon



Figure 1. Working on the Borax Properties Limited mine railroad



Figure 2. These two dare devils pose on the newly completed trestle

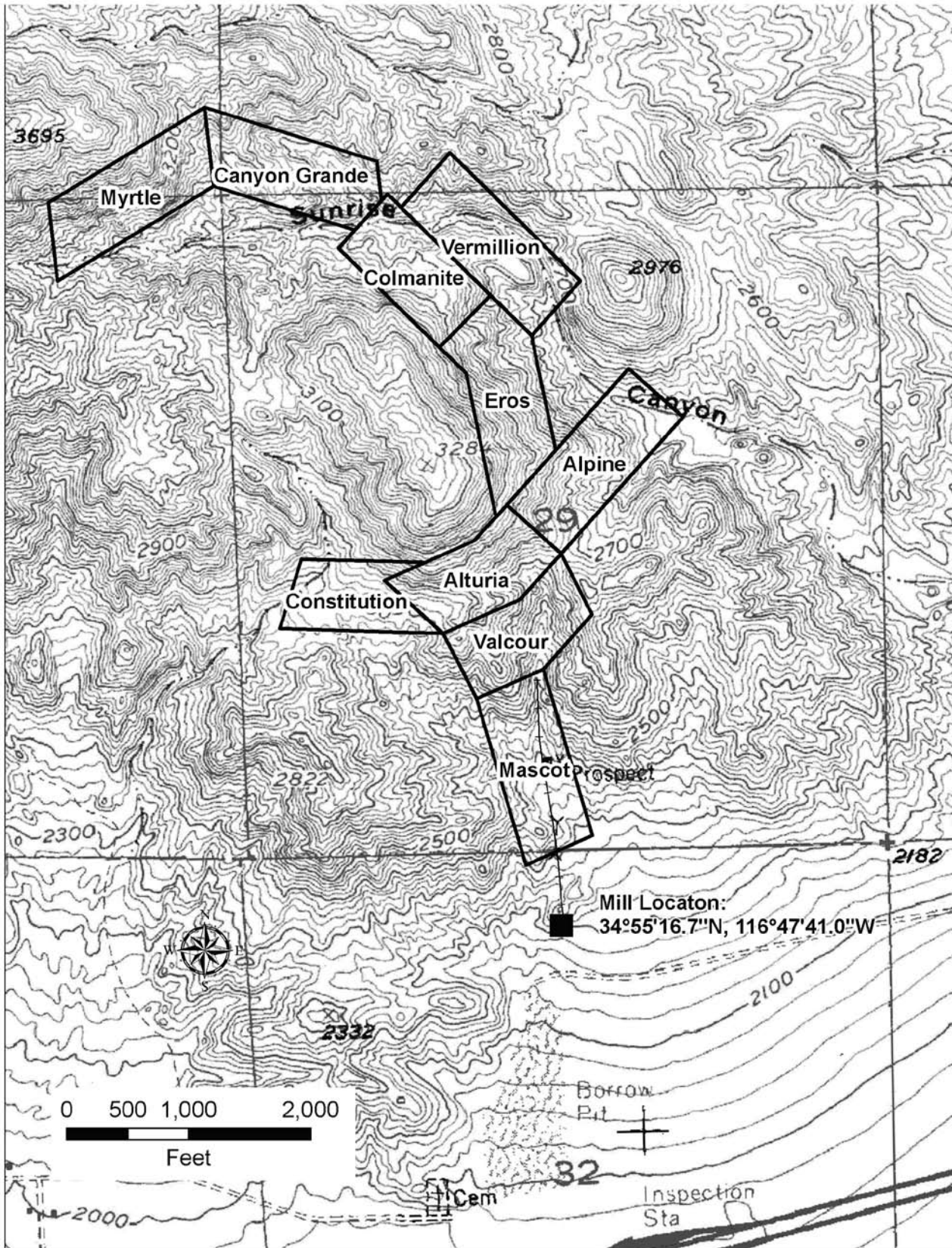


Figure 3. Map showing patented mining claims of the Paom Borate Company, T.10 N, R.2E, SBBM

Table 1
Patented Mining Claims of the Palm Borate Company

Claim Name Section	Mineral Survey	Date of Location	Patent Date Date	Patentee	Mine Name
Alpine	4174	29 Oct 1900	30 June 1905	American Board of Promoters Boracic Acid Co.	Colmanite Consolidated Mine
Canyon Grande	4174	8 Jan 1901	30 June 1905	American Board of Promoters Boracic Acid Co.	Colmanite Consolidated Mine
Alturia	4174	8 Jan 1901	30 June 1905	American Board of Promoters Boracic Acid Co.	Colmanite Consolidated Mine
Myrtle	4174	20 Apr 1901	30 June 1905	American Board of Promoters Boracic Acid Co.	Colmanite Consolidated Mine
Constitution	4705	20 Jun 1901	13 Oct 1910	Palm Borate Co.	
Colmanite	4174	21 Jul 1903	30 June 1905	American Board of Promoters Boracic Acid Co.	Colmanite Consolidated Mine
Eros	4174	21 Jul 1903	30 June 1905	American Board of Promoters Boracic Acid Co.	Colmanite Consolidated Mine
Vermillion	4707	8 Aug 1903	13 Oct 1910	Palm Borate Co.	
Mascot	4706	8 Aug 1903	13 Oct 1910	Palm Borate Co.	Mascot Consolidated Lode
Valcour	4706	8 Aug 1903	13 Oct 1910	Palm Borate Co.	Mascot Consolidated Lode



Figure 4. Mill and camp of the Borax Properties Limited about 1908. This mill was located at: 34°55'16.70"N, 116°47'41.00"W.

on the southeast end of the Calico Mountain appears to have been first developed by the Stephens & Greer mine, probably during the height of the borax boom in the Calico Mountains in the late 1890s (Bailey, 1902, p. 59).

The Palm Borate Company was organized in 1902 and was capitalized for \$500,000. The company eventually patented 10 mining claims (Table 1), which had been located between 1900 and 1903. Six of the claims were patented on June 30, 1905 under the company's previous name, the American Board of Promoters Boracic Acid Company (The Mining World, 6 June 1906, p. 4; Ver Plank, 1956, p. 282; Yale, 1905, p. 1022); the remainder were patented October 13, 1910, under the name Palm Borate Company (BLM Records).

In P. E. Nettleton's 1906 evaluation report of the six claims patented in 1905 he observed five shallow shafts with a total of 508 feet of workings, and an adit with a total of 1,550 feet of workings. The adit was on the Altura mining claim. The portal of the adit was just north of the south sideline of the claim, just west of the drainage that cuts the claim. He estimated the claims should yield 250,000 tons with an average of 8 percent boric acid, and that the mine should return an annual profit of \$131,000 per year.

In 1908, persuaded by this glowing report, The Borax Properties Limited of London, England (capitalized for £130,000) acquired an interest in the property. By June, construction was well under way on a gravity tramway, which required a trestle and tunnels to bring ore from mine to mill. In addition to the mill, which reportedly cost \$75,000, the company constructed a mess hall, a bunk house, and other administrative structures (Rhyolite Herald, 4 Nov. 1908; American Mining Review 20 June 1908, p. 15). A number of structures were also erected on the Mascot claim.

According to Wright (1953, p. 224) the plant operated only 10 days, when a drop in boric acid prices forced its closure. In 1910 the Palm Borate Company lost its corporate license to operate for failure to pay the required fees (Curry, 1911, p. 15).

References

- Bailey, Gilbert E., 1902, The Saline Deposits of California, California State Mining Bureau, Bulletin 24.
- Brelich, Henry, Sept. 16, 1907, 'Report on the Mines of the Palm Borate Company.' Unpublished report (on file at Death Valley National Park).
- Curry, C. F., 1911, Certified copy of compiled statement of domestic corporations whose charters have been forfeited at 4 O'clock p.m. November 30, 1910, California Secretary of State. p. 15
- Garrett, Donald, E. 1998, Borates, Handbook of Deposits, Processing, Properties, and Use, San Diego: Academic Press, p. 137
- Wright, Lauren A, et. al., 1953, Mines and Mineral Deposits of San Bernardino County, California, California Journal of Mines and Geology, Vol. 49 (January-April, 1953).
- Nettleton, P. E., February 1906, "Report on the Mines of the Palm Borate Company." Unpublished report (on file at Death Valley National Park).
- Ver Plank, W. E., 1956, "History of Borax Production in the United States." California Journal of Mines and Geology, Volume 52, Number 3, pp 280-283.
- Yale, Charles G., 1905, Borax: Mineral Resources of the United States. Calendar Year 1904, US Geological Survey, Minerals Yearbook, Vol. 1, p. 1022.
- Yale, Charles G., 1908, Borax: Mineral Resources of the United States. Calendar Year 1907, US Geological Survey, Minerals Yearbook, Part II, p. 634.

Geology and ore genesis of epithermal silver–barite mineralization in the central Mojave Desert, California

David R. Jessey

Geological Sciences Department, California Polytechnic University-Pomona

Abstract

Silver–barite deposits of the central Mojave Desert, near Barstow, CA have yielded more than \$20 million in silver and barite. Mineralization occurs in rocks ranging in age from Precambrian Waterman Gneiss to early-middle Miocene Barstow Formation, but nearly all of the recorded production is from the early Miocene Pickhandle Formation. Pickhandle volcanic rocks host numerous northwest-trending silver–barite veins emplaced along steeply dipping faults. The highest ore grades are encountered in the oxidized apex of the veins where silver chlorides and native silver are the chief ore minerals.

Geochronology indicates the main pulse of mineralization occurred at 17 Ma. This coincides with the waning stages of detachment faulting (Waterman Hills detachment fault) and likely predates the initiation of dextral shear along the Calico fault. Silver–barite ore was deposited as massive, open-space filling within listric faults in the Pickhandle Formation, while it is largely disseminated within siltstone of the Barstow Formation. Paragenetic studies suggest that mineralization happened was constrained to a single event, requiring a plumbing system that enabled fluids to rise upward, perhaps along faults in the Barstow Formation, and then migrate laterally along permeable horizons. Fluid inclusion geothermometry indicates Pickhandle veins were emplaced at depths in excess of one kilometer while Barstow ores were deposited very near the surface.

One model proposes meteoric water circulating downward along the detachment fault surface. At depth, the meteoric water would be heated by plutons that fed Pickhandle volcanism and move convectively upward along normal faults within the overlying plate of the Waterman Hills detachment fault. As the fluid cooled, ore would be deposited within the Pickhandle Formation. Where available conduits and favorable stratigraphic horizons presented themselves, ore deposition would also have occurred within the overlying Barstow Formation. Subsequent right oblique-slip and transpression along the Calico fault resulted in uplift of the Calico Mountains, creating the present day topographic and structural relationships. In deeply eroded areas, such as Wall Street Canyon, the apices of outcropping silver veins were oxidized generating the “high-grade” silver chloride ores of the Calico District.

Introduction

Silver–barite mineralization occurs along a 30 kilometer wide band stretching across the central Mojave Desert from the Bullion and Cady Mountains near Ludlow, CA northwest to the Mitchel Range, a distance of approximately 60 miles. Most of the prospects and nearly all of the significant production occurred within the Calico Mining District located along the southern flank of the Calico Mountains four miles northeast of Barstow, CA (Fig. 1). The largest mines were situated along the south flank of the Calico Mountains, in or immediately adjacent to Wall Street Canyon.

The history of the Calico Mining District has been documented by numerous authors; Weber (1966) provides an excellent summary. The district flourished from the 1880s to the close of the nineteenth century. The manpower shortage during World War I and subsequent Great Depression marked the end of significant mining activity, but not before the district had established itself as the largest silver producer in California. Total silver production is thought to have exceeded \$20 million. However, put in the perspective of a true giant like the Comstock Lode (total production = \$396 million) (Smith, 1943), the Calico District remains quite small. During the 1950s an economic boom and a renewed interest in silver resulted in the reopening of several of the district's mines, but production during that era was small.

Petroleum exploration and development in California made barite an economically attractive commodity in the 1950s. From 1957 to 1961 the Leviathan Mine, northwest of Wall Street Canyon, was the largest barite producer on the west coast. Substantial barite reserves remain, but discovery of the much larger Battle Mountain, Nevada deposits severely impacted the economic viability of Calico barite. In the early 1960s, ASARCO Inc. and Superior Oil Corp. began exploration and limited development of disseminated silver deposits along the southwest flank of the Calico Mountains; the Waterloo and Langtry. Development was hampered by environmental concerns and the low silver ore grade. The properties have been largely inactive for nearly two decades, but recent spikes in precious metal prices have caused renewed interest in both properties.

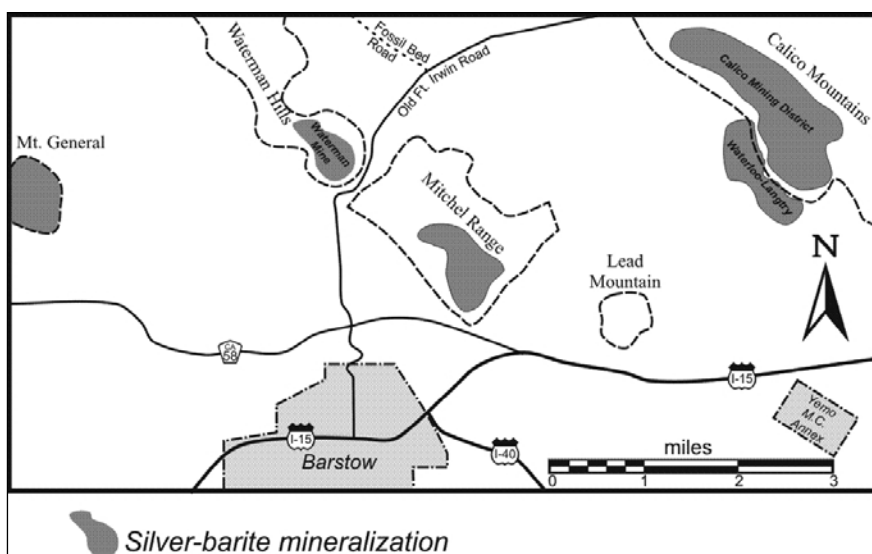


Figure 1. Silver–barite occurrences near the Calico Mountains.

General geology

Stratigraphy

Portions of the central Mojave Desert have been mapped by DeLeen (1950), McCulloh (1952, 1965), Weber (1965), Dibblee (1967, 1970), Mero (1972), Fletcher (1986), Payne and Glass (1987), Jessey and Yamashiro (1988), Cox and Wilshire (1993), Fletcher et al. (1995) and Singleton and Gans (2008). Basement rocks fall into one of three categories: (1) the Waterman Gneiss, an assemblage of Precambrian/Paleozoic meta-igneous and meta-sedimentary rocks; (2) the Coyote Group, a sequence of weakly metamorphosed, Paleozoic sedimentary rocks; and (3) Mesozoic intrusive and extrusive rocks (Fig. 2).

The Waterman Gneiss is exposed in the Waterman Hills and the Mitchell Range north of Barstow, as well as in isolated outcrops in the Hinkley Hills and in the low hills west of Harper Lake. Lithologically, the Waterman Gneiss varies from dioritic gneiss, to impure marble and quartzite, to various types of mylonite. When not affected by later retrograde events, Waterman Gneiss lies within the amphibolite facies of regional metamorphism. Subsequent retrograde metamorphism associated with regional detachment (see below) has locally superimposed a chlorite-grade greenschist facies (Glazner et al. 1988). The age of the Waterman gneiss is conjectural. Dibblee (1967) assigned it to the Precambrian, while Glazner et al. (1988) suggest late Precambrian–Paleozoic.

A sequence of sedimentary and metamorphosed volcanic rocks lies along the north flank of the Calico Mountains (McCulloh, 1952). It consists of marble, schist, conglomerate, quartzite, hornfels, metamorphosed andesite and minor metamorphosed basalt. McCulloh termed

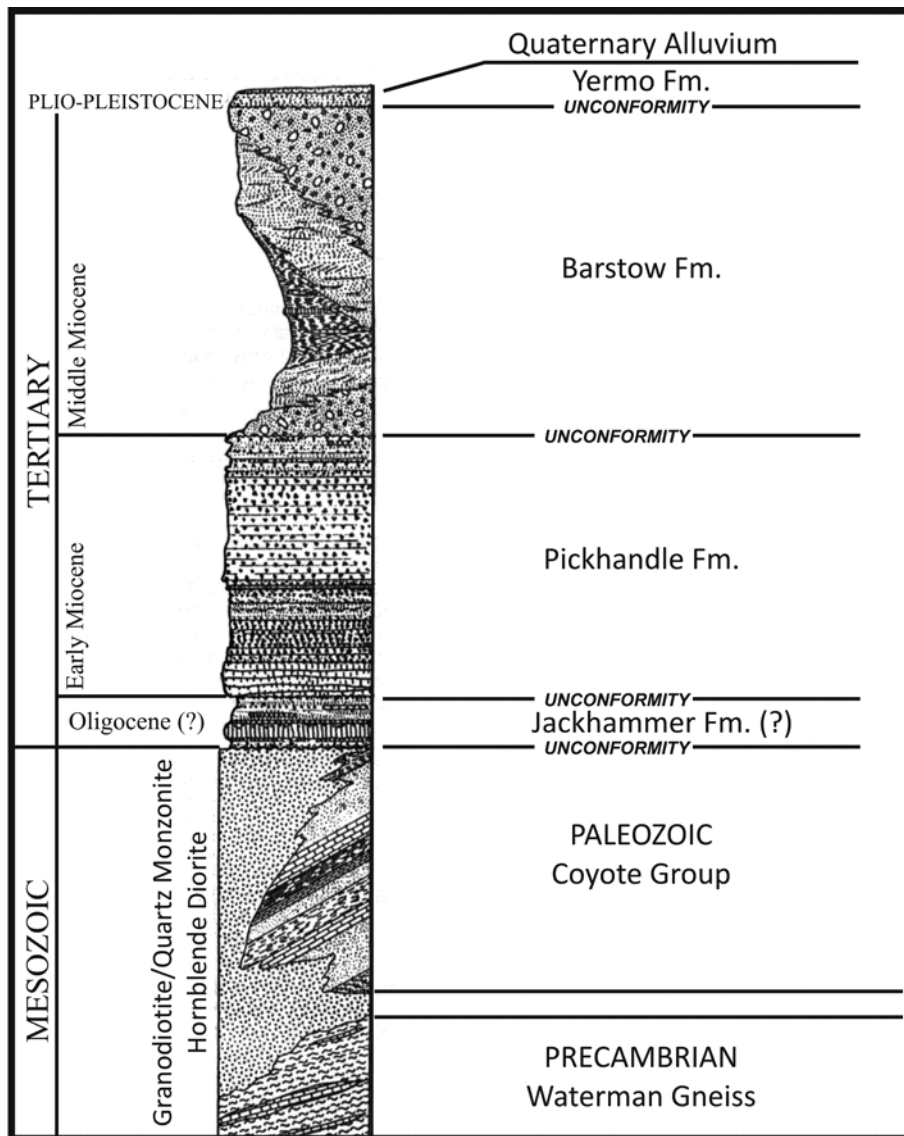


Figure 2. Generalized stratigraphic column for the Calico District (after Jessey and Tarman, 1988).

the rocks the Coyote Group for Coyote Dry Lake to the northeast of the outcrop area. Poorly preserved Paleozoic fossils were found near the middle of the sequence, but McCulloh (1952) believed it possible that the upper half of the Coyote Group is of Mesozoic age. Actual stratigraphic relationships are difficult to determine as both the top and bottom of the Coyote Group are truncated by Mesozoic intrusive rocks.

Small, scattered exposures of porphyritic volcanic rocks of uncertain age crop out at the southeast end of the Calico Mountains and south of I-15. These lithologically heterogeneous rocks have been correlated with several different rock units across the Mojave. Perhaps the closest match is the Jurassic Sidewinder Volcanic series (Bowen, 1954) near Victorville.

Plutonic rocks crop out at the east and northeast end of the Calico Mountains, as well as locally in the Water-

man Hills, Mitchel Range and near Mt. General. Most intrusions are middle to late Mesozoic in age, yielding radiometric dates that cluster around 90 Ma (Payne and Glass, 1987). However, Walker et al., (1995) report an age of 22-23 Ma for a pluton from the Waterman Hills (Walker et al., 1995). Compositionally, the intrusives range from diorite and quartz diorite to granodiorite and quartz monzonite. Characteristically, the intrusive rocks weather to form subdued terrains resulting in low hills, in direct contrast to the more rugged topography of Tertiary volcanic rocks.

McCulloh (1952) mapped a major unconformity separating basement rocks from overlying Tertiary units. Dokka (1986), Glazner, et al. (1988) and Jessey and Tarman (1988) reinterpreted the unconformity as a regional detachment fault: a low angle normal fault separating lower plate basement from upper plate Tertiary volcanic and sedimentary rocks. The basal unit above the discontinuity is the Jackhammer Formation (McCulloh, 1952). At its type section in Jackhammer Gap it consists of 700 feet of tuff, tuff breccia,

volcanogenic sedimentary rock, arkosic conglomerate, and basalt. Dokka and Woodburne (1986) report an age date of 25.6 Ma (Oligocene) for a basalt near the top of the Jackhammer. Subsequent research has questioned the need to separate the Jackhammer from the overlying Pickhandle Formation. The Pickhandle is generally regarded to be Miocene in age (24–18 Ma, Burke et al., 1982), however, the contact with the Jackhammer appears conformable and lithologically the two units are similar. This led Singleton and Gans (2008) to combine the Jackhammer and Pickhandle (Fig. 3).

The Early Miocene Pickhandle Formation (McCulloh, 1952) is named for exposures in Pickhandle Pass of the northern Calico Mountains. The Pickhandle is one of two major ore-bearing units in the Calico district, hosting much of the vein-type barite-silver mineralization. In

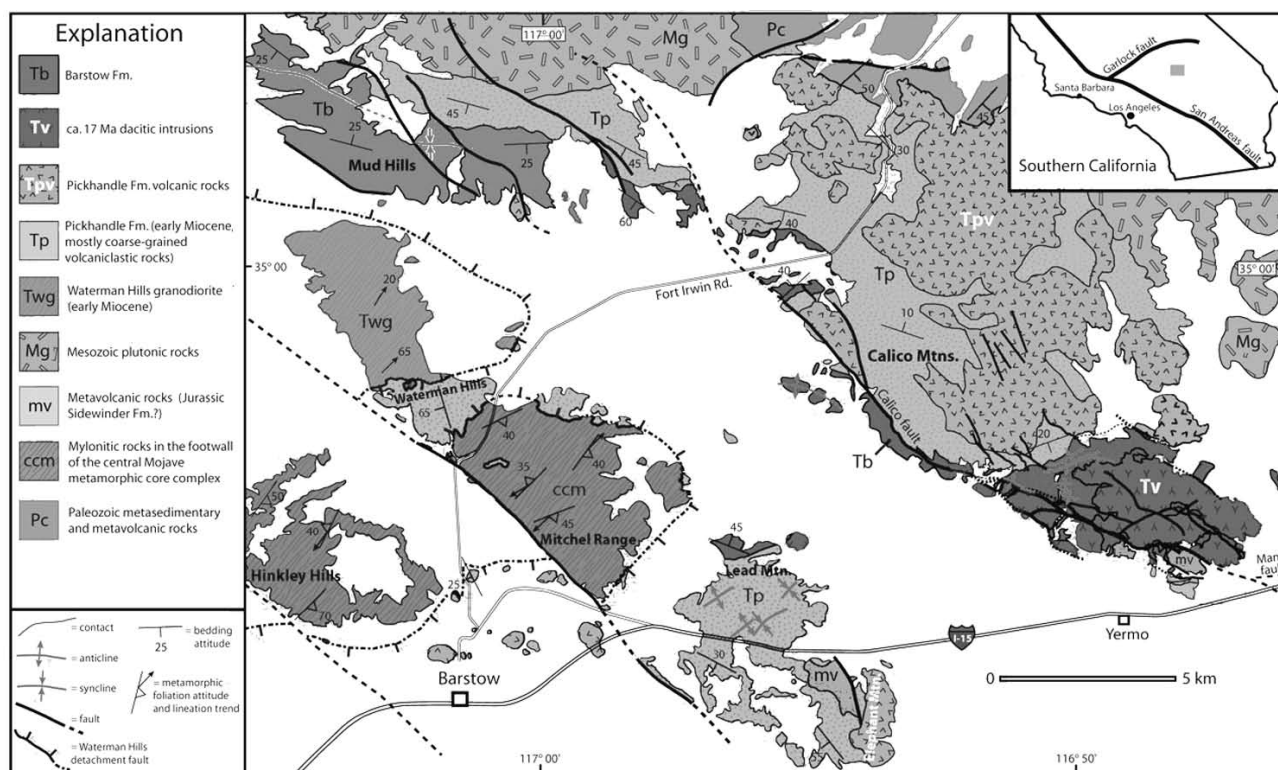


Figure 3. Generalized geologic map of the Calico Mountains area (modified from Singleton and Gans, 2008).

general the Pickhandle is a series of intercalated pyroclastic rocks and volcanic flows, the latter predominantly of rhyodacitic to dacitic composition. Minor volcaniclastic sedimentary units occur throughout the sequence, but are more common near the contact with the overlying Barstow Formation. Age dates by various authors suggest the bulk of the Pickhandle Formation was deposited by 19 Ma (Singleton and Gans, 2008); however a series of dacite domes near the southeast end of the Calico Mountains have yielded ages as young as 16.8 Ma (Singleton and Gans, 2008) indicating that the waning stages of Pickhandle volcanism may be coeval with deposition of the overlying Barstow Formation.

The Middle Miocene Barstow Formation overlies the Pickhandle volcanics, the basal contact marked by transition from volcanics to sedimentary rocks. While many researchers have argued that Barstow sedimentation began at 17 Ma (Woodburne, et. al., 1990, Glazner et al., 2002), field relationships suggest there is overlap between Pickhandle volcanism and Barstow deposition; the latter beginning, perhaps, as early as 18 Ma (Reynolds, pers. comm., 2009). The Barstow Formation was deposited in the east-trending, fault-controlled basin (Dokka, 1979). Thickness ranges from 2400 feet in the northwest part of the basin to 1200 feet in the Alvord Mountains (Byers, 1960; Dibblee, 1980). Lithologically, the Barstow consists of a wide array of sediment types reflecting deposition

in a shallow lake, tributary stream systems, and alluvial fans. In the Calico Mountains, a gradual upward coarsening can be observed with rocks grading from calcareous mudstones and sandstones to conglomerates. Impure limestones with thicknesses ranging from fractions of inches to a few feet are locally present at the base of the Barstow Formation in the Calico Mountains. The Barstow is the second major ore host in the region. The ore occurs as disseminated grains and randomly-oriented stockwork veinlets in permeable siltstones and porous sandstones.

Widespread gravels in the Yermo Hills, east of the Calicos and in canyons in the southern Calico Mountain have been termed the Yermo Formation (McCulloh, 1952). The gravels are Plio-Pleistocene in age.

Structure

Rocks of the central Mojave Desert have been complexly deformed. Pre-Cenozoic structures have been largely overprinted by Cenozoic extension and dextral shear. Early Miocene detachment and extension generally preceded the right-slip tectonics related to the North American–Pacific plate boundary. However, Glazner et al. (2002) point out that timing and magnitude of both events are controversial. Emplacement of silver-barite mineralization is directly related to at least one, and perhaps both, of these episodes of deformation.

The Waterman Hills detachment fault (Fig. 3) juxtaposes Cenozoic volcanic and sedimentary rocks in

the hanging wall over mylonitic basement rocks in the footwall. Timing and extent of extension is open to question. Dokka (1989) argues that extension occurred along an east–west zone across the entire Mojave Desert, while Glazner et al. (2002) suggest that extension was confined to a 25-km-wide belt centered within the central Mojave Desert. Glazner et al. (1989) propose 40–60 km of north-east transport for upper plate rocks; field observations do not tend to support this magnitude of movement. Available evidence indicates that the majority of detachment occurred between 25 and 19 Ma (Singleton and Gans, 2008). However, emplacement of a series of dacite domes at the southeast edge of the Calico Mountains suggests that volcanism and extension may have continued until as recently as 16.5 Ma.

Neogene strike-slip faulting along the Calico and Harper Lake faults appears to postdate extension in the central Mojave Desert. These faults are a part of a larger

set of northwest-trending right-lateral faults that accommodate plate motion between the Pacific and North American plates (Dokka and Travis, 1990b). While the cumulative dextral shear across the region is on the order of 50–75 km (Dokka and Travis, 1990a), the total shear across the Calico and Harper Lake faults is probably much less. Early studies proposed tens of kilometers of right slip along the Calico fault (Dibblee, 1980, Dokka, 1983), but Fletcher (1986) and more recently Singleton and Gans (2008) suggest no more than three kilometers of right-slip. The onset of strike-slip faulting is uncertain, but Bartley et al. (1990) believe it may have begun as early as 19 Ma. If so, this would necessitate some overlap with detachment. Right-slip motion continues to the present.

Local Neogene shortening in the central Mojave has been attributed to zones of transpression along northwest-striking dextral faults (e.g., Dibblee, 1980b,

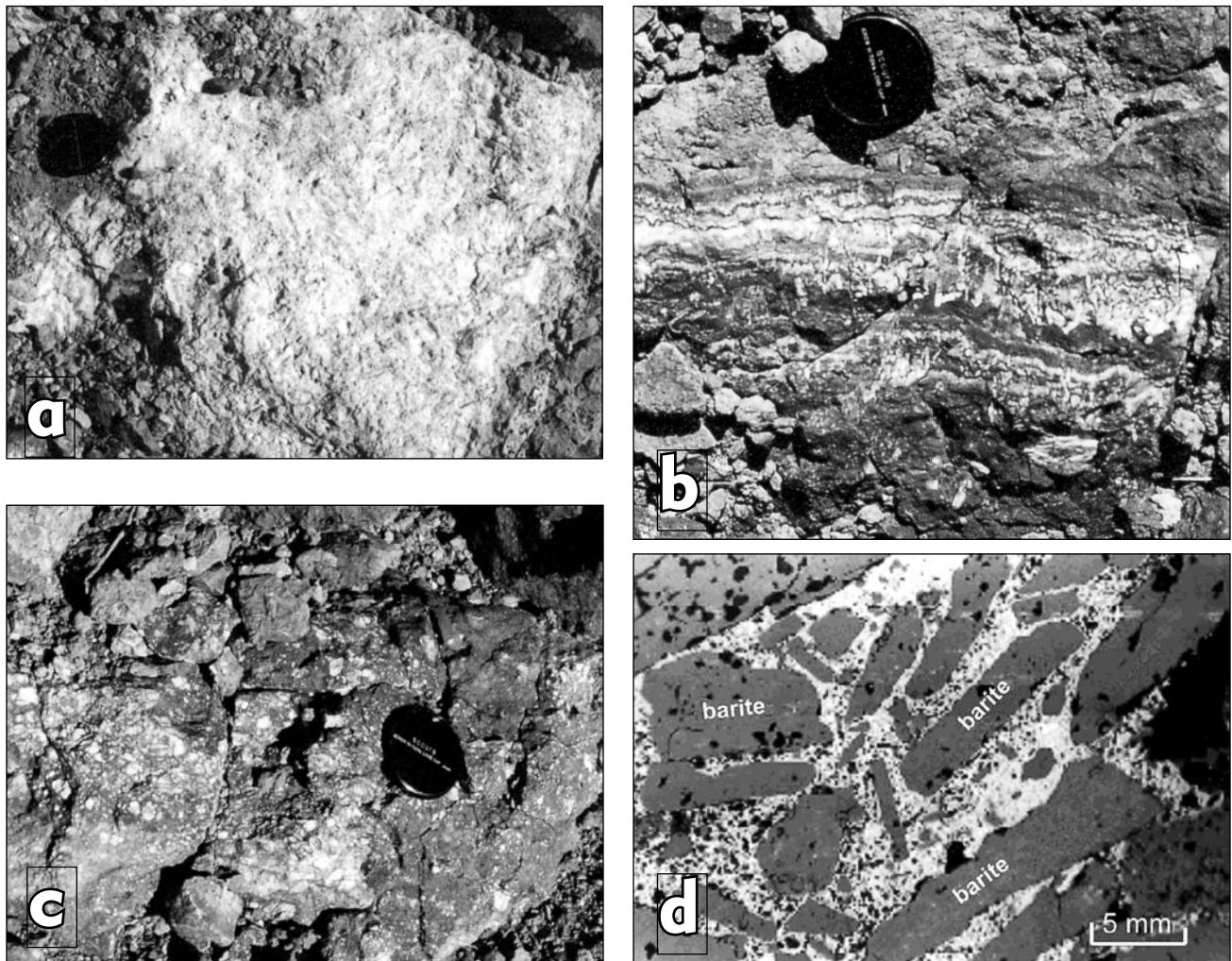


Figure 4. (a) Barite vein lacking sulfides or oxides; (b) comb-structured barite (white) with younger iron and Mg oxides (dark) lining the interior of the vein; (c) sulfide/oxide minerals (dark) brecciating and replacing barite (white); and (d) photomicrograph of barite (gray) veined and partially replaced by iron oxides and sulfides (white).

1994), or regional north–south contraction (Bartley et al., 1990). The shortening commonly manifests itself as a series of east–west trending folds in lacustrine/fluviol sedimentary rocks of the Miocene Barstow Formation, such as those exposed in the Calico Ghost Town parking lot (Tarman, 1988). The magnitude of shortening represented by these folds is not well documented, and the timing is uncertain. Tarman and Thompson (1988) argue that much of the folding occurred as a consequence of transpression, and Singleton and Gans (2008) state that a constraining bend in the Calico fault has caused the folding. Jessey and Yamashiro (1988) suggest that transpression was also responsible for a reverse fault mapped in the northwestern Calico Mountains. Singleton and Gans (2008) support the hypothesized thrusting and suggest the Calico fault may have undergone at least one kilometer of reverse movement (north side up).

Silver-barite mineralization

Occurrence

Calico District

The bulk of silver production from the central Mojave Desert has come from the Calico District and much of that from mines within or near Wall Street Canyon. Epigenetic silver–barite veins occur within the Early Miocene Pickhandle Formation (upper plate Waterman Hills detachment fault). The Pickhandle Formation is composed of flows of dacite to rhyodacite composition, associated tuffs, and volcanic breccias. All barite mineralization is localized within northwest trending (N25–75°W) faults which crosscut the gently dipping volcanic rocks at steep angles. Offset along fault planes is small, generally less than a few meters. Most veins dip to the southwest, but those in the northern portion of the district commonly dip to the northeast. These structural attitudes led Dibblee (1970) to propose that the Calico Mountain block is a northwest-trending anticline. Ore minerals have been deposited within open space and fill cavities in breccia. Barite crystals are euhedral, and often one centimeter or more in length. In many veins the barite has been brecciated and the interstices filled by iron oxides (magnetite and hematite) and sulfides. The near surface, oxidized portion of the veins yielded embolite (Ag (Br, Cl)) with lesser chlorargyrite, cerargyrite and native silver. Primary sulfides encountered at depth were below cutoff grade and rarely mined.

Three vein types have been recognized with gradations between all types (Jessey, 1986). Monomineralic veins of white barite (Fig. 4a) lacking any alteration are rare, occurring only to the northwest of Wall Street Canyon. These veins represent the earliest stage of mineralization.

Veins with comb-textured barite lining the walls and a mixture of jasperoid and rarer magnetite, hematite, and magnesium oxides filling the interstices are more common (Fig. 4b). These veins represent the second stage of mineralization. Although silver is present, the veins generally assay at less than 1 oz/ton.

Also common are “black-matrix” veins comprised of brecciated and partially replaced barite fragments in a matrix of iron and manganese oxides and sulfides (Fig. 4c and 4d). Magnetite occurs locally, with partial to total alteration to hematite. A variety of undifferentiated manganese oxide species are also present. Sulfides are rarer. They consist of pyrite and galena with trace chalcopyrite and tennantite. Silver assays as high as 1100 ounces per ton have been reported, but 3–5 oz/ton is closer to the norm. The silver-bearing species is uncertain. Samples of high grade silver ore were examined and found to contain a high proportion of galena suggesting argentiferous galena; however, other samples contained acanthite (silver sulfide) and native silver.

Propylitic alteration (chlorite+calcite±epidote) is the most pervasive and readily recognizable form of alteration in the district. The alteration is widespread and commonly not proximal to barite veins. Silicification, consisting of fine-grained chalcedony and jasperoid is more closely associated with the silver-barite mineralization. Payne and Glass (1987) also report hydrothermal alteration of amphibole to celadonite. Local kaolinization is present in host rocks containing phenocrysts of plagioclase.

Oxidized barite veins are thought to represent the supergene equivalent of the black matrix barite veins. They are easily recognized by the brick red alteration adjacent to and intimately associated with the barite. The alteration consists of jasperoid and secondary, fine-grained hematite. Most primary sulfides have been replaced, with only occasional, heavily corroded, pyrite remaining. Galena has altered to cerrusite. Magnetite has been replaced by hematite, leaving pseudomorphs of the original magnetite



Figure 5. Photomicrograph of disseminated barite (white) in siltstone (from Fletcher, 1986).

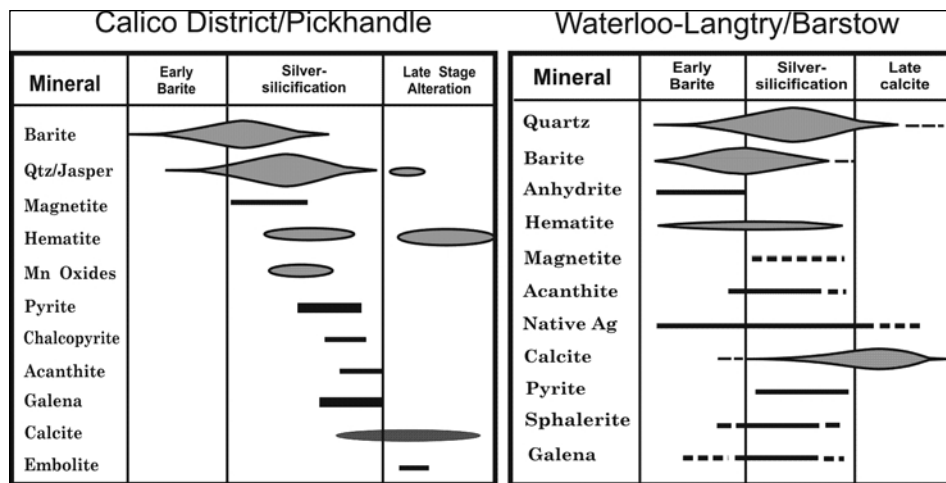


Figure 6. Detailed paragenesis for the vein-type ores of the Calico District and disseminated ores of the Barstow Formation (Waterloo/Langtry paragenesis after Fletcher, 1986).

crystals. Secondary silver minerals, particularly embolite and cerargyrite, are present in some veins but absent in others. Silver grades for the oxidized veins of Wall Street Canyon are quite high, exceeding 10 oz/ton.

Barstow Formation

Fletcher (1986) studied the silver-barite mineralization of the Middle Miocene Barstow Formation. He identified two separate tabular ore bodies: the Waterloo deposit, north of the Calico fault, and the Langtry, 2.5 kilometers to the west and on the south side of the Calico fault. He argued that the deposits were once part of a single ore body that was subsequently offset by right slip along the Calico fault. The Barstow Formation in this area is composed of lacustrine sandstones, siltstones and mudstones with minor interbedded volcanoclastic sandstones. Fletcher reports that barite occurs as disseminated grains and cementing material within siltstone (99%) (Fig. 5) and more rarely as veinlets (1%). The veinlets have a random orientation in contrast to the northwest strike of those in the underlying Pickhandle volcanics.

Alteration consists of bleaching that most commonly manifests itself as K-feldspar replacement of detrital feldspar grains. Although not termed as such by Fletcher, this alteration appears similar to the pervasive potassic alteration associated with many epithermal precious metal deposits. The alteration is correlative with an early stage of barite-quartz mineralization. Intense silicification, most commonly as colloform bands of jasperoidal chalcedony with minor recrystallized quartz is also present. For a more detailed discussion of the deposits see Fletcher (1986).

Waterman Hills

Silver ore in the Waterman Hills occurs in two northwest-striking veins that can be traced for a distance of over 500

meters along strike (Ireland, 1993). The veins lie within a sequence of volcanoclastic sedimentary rocks and volcanics of uncertain age. Dibblee (1952) correlated the rocks with the Miocene Pickhandle Formation. However, the more silicic nature of the volcanics (rhyolites) and the distinct northwest strike of the sedimentary units suggest that correlation is tentative. The veins, while concordant to bedding of the volcanoclastic sedimentary rocks at the surface, crosscut the strata in underground

workings. The Waterman Hills detachment fault crops out less than one kilometer north and south of the barite veins. Projections from surface outcrops indicate the fault contact should lie less than 150 meters below the present erosional surface. Mineralization resembles that of the Calico District wherein coarse-grained barite has been brecciated and replaced by a matrix of iron oxides. Little is known about the silver occurrence and no sulfides were observed either on the dumps or in underground workings.

Mitchel Range

The Waterman Gneiss (lower plate of the Waterman Hills detachment fault) hosts a group of northwest trending barite veins at the southeast end of the Mitchel Range. The veins crosscut the Waterman Gneiss at steep angles and postdate a prominent mylonitic fabric. Vein margins are slickensided, suggesting emplacement within fault zones. Barite occurs as coarse-grained rosette-shaped aggregates, often heavily iron stained. Minor iron oxides are present, as are trace amounts of galena. No silver mineralization was observed and the limited extent of the workings suggests ore grades were sub-economic.

Mt. General

The silver-barite mineralization of the Mt. General area, 8 kilometers north of Lenwood CA, is the most enigmatic of all occurrences in the central Mojave Desert. Mapping suggests host rocks are highly extended, felsic volcanic rocks lithologically similar to those of the Waterman Hills. Stratigraphic position of these rocks is uncertain. Perhaps the most unusual aspect of the Mt. General occurrence is the barite. It occurs within poorly defined northwest-trending shear zones which lack any continuity along strike. The dark, cloudy barite, com-

monly with inclusions of particulate matter has been extensively sheared and brecciated. Microscopic examination indicates the barite and minor associated calcite have undergone extensive recrystallization. Iron staining is common in underground workings, but iron oxides occur in only trace amounts. The limited underground workings and scattered prospect pits suggest low silver ore grades.

Paragenesis

Detailed paragenetic studies of the Calico vein mineralization (Jessey, 1992) and disseminated ores in the Barstow Formation (Fletcher, 1986) have shown only minor differences in the sequence of ore mineral deposition (Fig. 6). Less detailed studies of mineralization in the Waterman Hills, Mitchell Range and Mt. General indicate a similar, albeit less complex, paragenesis.

To generalize, barite and chalcedony/quartz were deposited early (Early Barite stage). This was followed by an episode of pervasive silicification (initial stage of Silver-Silicification) and subsequently by hematite, magnetite and manganese oxides. A variety of sulfides followed the oxides (late stage Silver-Silicification). The Late Stage Alteration included secondary iron oxides, silver chlorides, and calcite deposited by supergene fluids.

Fletcher (1986) and Rosso (1992) attribute the observed paragenetic sequence to changes in fluid pH. However, the elevated salinity of ore fluids (see Fluid Inclusion Geothermometry) makes this difficult to rationalize, as saline fluids typically buffer reactions and resist pH change. Another explanation would be a change in Eh. Early oxidizing ore fluids would deposit sulfates (barite) and oxides (magnetite/hematite). As conditions became more reducing, pyrite, galena and acanthite would form.

The overall similarity of paragenesis for the Pickhandle veins and Barstow disseminated ores is supported by the atypical nature of the paragenetic sequence (oxides early—sulfides late). Most epithermal deposits are characterized by early sulfide mineralization with later deposition of oxides and carbonates. As both the Calico veins and Waterloo/Langtry deposits are paragenetically atypical, it

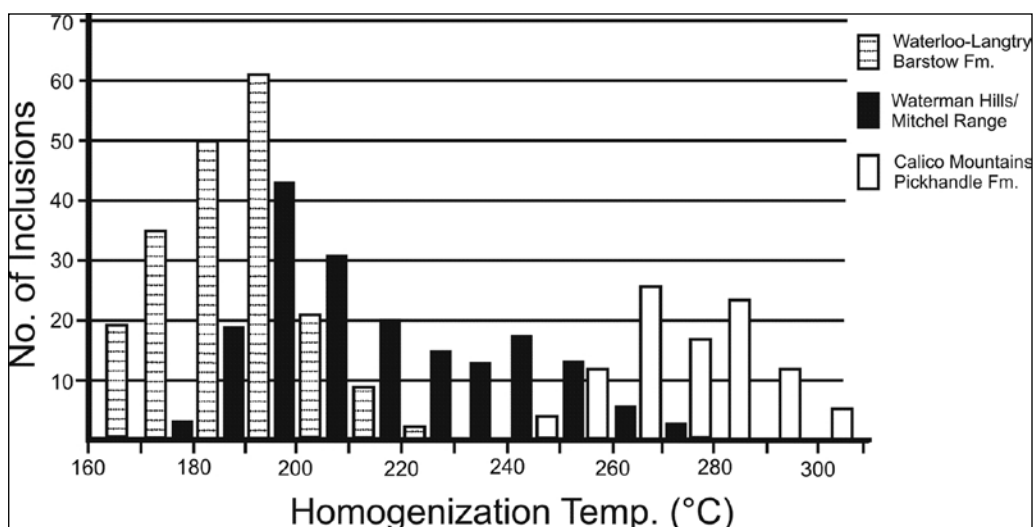


Figure 7. Fluid inclusion homogenization temperatures for central Mojave Desert barite occurrences (data for Waterloo/Langtry from Fletcher, 1986).

seems more likely there was a single evolving fluid responsible for ore deposition and not two unrelated events.

Fluid inclusion geothermometry

Fluid inclusion data are summarized in Figure 7. Homogenization temperatures for barite range from 160° to 310° C. Calico District barite fluid inclusions are often large (>50 microns) and are restricted to Type I, indicating the fluids were not boiling at the time of barite crystallization. Homogenization temperatures range from 240° to 310° C and salinities from 2–4 wt% NaCl. Utilizing paragenesis, barite geothermometry and salinity, and Eh-pH relationships, Rosso (1992) calculated depth of barite emplacement at one to 1.5 kilometers. Limited analysis of quartz (not shown) has yielded homogenization temperatures similar to but slightly higher than those for barite (average \approx 290°C).

Fletcher (1986) analyzed both barite and quartz from the Waterloo and Langtry deposits within the Barstow Formation. He reported homogenization temperatures from 140° to 230°C with salinities of 4–6 wt% NaCl. Barite inclusions were generally smaller than those from the Pickhandle veins and Type II inclusions, indicative of boiling, were common. Utilizing similar data and logic to that of Rosso, Fletcher (1986) calculated depth of emplacement of Barstow mineralization at no greater than a few hundred meters.

Figure 7 combines data for the Waterman Hills and Mitchell Range. While homogenization temperatures show considerable overlap (Waterman Hills 180° to 220° C and Mitchell Range 180–270°C) there are significant differences between the two localities (Jessey and Tarmen, 1994). Waterman Hills barite veins are hosted by upper plate rocks correlative to the Miocene Pickhandle Formation (upper plate WHDF). Inclusions were the

smallest observed in the central Mojave Desert (average diameter 10-15 microns) and evidence for boiling the most widespread. Although salinity could not be determined, field observation suggests barite was emplaced at shallow depths. Mitchel Range barite was deposited in veins cutting basement rocks (Waterman Gneiss) in the lower plate of the WHDF. Barite occurs as coarse-grained rosette-shaped aggregates, often heavily iron stained. Crystals are water-clear and contain large inclusions (>100 microns). There is no evidence of fluid boiling.

Mt. General barite (not shown in Figure 7) is dark, cloudy, and yields few usable inclusions. Furthermore, upon heating the inclusions often fracture making homogenization temperature difficult to measure. Data from a small number of observations (12) suggests the average homogenization temperature is approximately 210°C.

Discussion

Any model for the genesis of silver-barite mineralization must take into account structural and stratigraphic relationships, as well as mineral paragenesis and fluid inclusion geothermometry and geobarometry. The first question that must be addressed is the age of mineralization. Fletcher (1986) states that two samples of potassically altered Barstow Formation yielded K/Ar dates of 17.1 and 17.5 Ma respectively. This is consistent with timing of deposition of the Barstow Formation.

The relationship of the mineralization to structural evolution is more problematic. Singleton and Gans (2008) argue that unroofing of the Waterman Hills core complex occurred between 25 Ma and 19 Ma. Bartley et al. (1990) place the onset of dextral shear in the central Mojave at 19 Ma, although it is unclear if this includes the Calico fault. Singleton and Gans (2008) state, however, that active dacitic volcanism was occurring at the southeast end of the Calico Mountains as recently as 16.5 Ma. This is difficult to reconcile with a transpressive environment. Such a setting would result in compressional stress inhibiting the rise of silicic magmas to the surface. It seems more likely that dacitic volcanism was related to extension during the waning stages of detachment. Furthermore, the Waterloo and Langtry deposits are cleanly offset 2–3 kilometers by the Calico fault (Fletcher, 1986). This must have occurred after the ore was emplaced at 17–17.5 Ma. Had dextral shear been contemporaneous with ore emplacement it is expected the ore bodies would be elongated or “strung out” parallel to that fault, or that at least pods of mineralization would be found scattered between the two ore bodies. Neither is the case; field evidence supports the emplacement of mineralization before the Calico fault underwent right-slip motion.

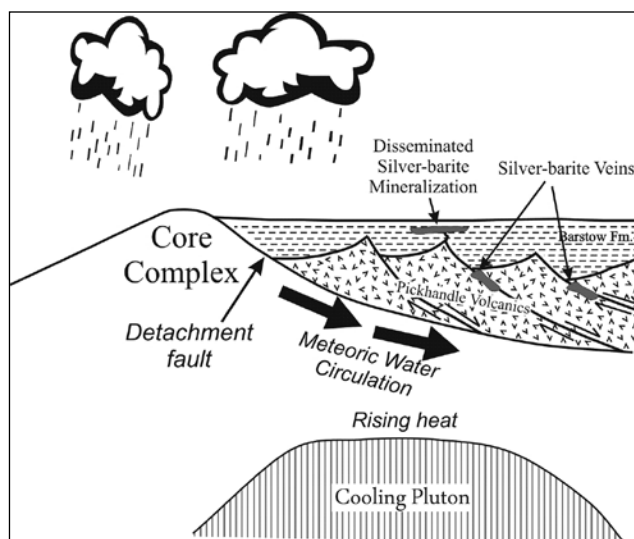


Figure 8. Cartoon model for silver-barite mineralization in the central Mojave Desert.

Figure 8 is a suggested model for silver-barite mineralization in the central Mojave. Northeast transport of the detached block from 25 to 19 Ma created a linear northwest-southeast-trending basin. Extension resulted in volcanism (Pickhandle Formation) followed by erosion and deposition of the Barstow Formation. Extension resumed around 17.5 Ma and continued until 16.5 Ma (Singleton and Gans, 2008). Volcanism occurred in the eastern Calico Mountains and heat from the pluton(s) drove a convective geothermal system to the west. Meteoric water moved downward along the detachment fault surface, was heated, and moved convectively upward along listric faults in the overlying plate. As the fluid cooled or was diluted, ore deposition occurred within the Pickhandle Formation. Evidence suggests ore deposition in the overlying Barstow Formation was contemporaneous. This requires that fluids rose locally along faults into the Barstow Formation. Impermeable horizons (shales?) would have prevented fluid migration unless they were breached by faulting. The nature of faulting in the Barstow Formation is uncertain. Fletcher (1986) states that 1% of all Barstow mineralization is vein controlled, but that the veins have a random orientation, unlike the northwest strike of all Pickhandle veins. Perhaps extension occurred when the lower Barstow was only poorly indurated and hosts for mineralization were porous and brecciated sediments of the Barstow Formation.

Right oblique-slip on the Calico fault then offset mineral zones developed in the Barstow Formation 2–3 kilometers. In addition, transpression caused reverse faulting of the block north of the Calico fault resulting in uplift and anticlinal arching of the Calico Mountains (Dibblee, 1980; Singleton and Gans, 2008). This model agrees with Rosso et al. (1992) and Fletcher (1986) who

argue that Pickhandle mineralization was emplaced at depths of one–1.5 kilometers while mineralization in the Barstow Formation was deposited at depths of less than a few hundred meters. The present topographic relationship requires at least a kilometer of uplift to account for Pickhandle silver–barite veins at the head of Wall St Canyon situated 300 meters above disseminated mineralization of the Waterloo deposit in the Barstow Formation.

While the above model fits much of the analytical data and field observation, some questions nonetheless remain. Barite mineralization in the Mitchel Range occurs in northwest-trending veins within Waterman Gneiss (lower plate of the Waterman Hills detachment fault). How can this be reconciled with the model? Furthermore, Dokka (1986) states that the Pickhandle Formation has undergone less than 5% extension, a conclusion that led him to exclude rocks north of the Calico fault from the extended terrain of the central Mojave Desert. If the Calico Mountains have undergone minimal extension, how was the fault-controlled open space for the Pickhandle mineralization created? The paragenetic sequence requires that ore fluids were initially oxidizing (barite/hematite/magnetite), becoming more reduced (pyrite, other sulfides) over time. What caused the Eh drop and why did it occur late in the ore fluid evolution?

References cited

- Bartley, J.M., Glazner, A.F., and Schermer, E.R., 1990, North-south contraction of the Mojave block and strike-slip tectonics in southern California: *Science*, v. 248, p. 1398-1401.
- Bowen, O.E., 1954, *Geology and mineral deposits of the Barstow Quadrangle, San Bernardino County, CA*: California Division of Mines and Geology Bulletin 165, p. 185.
- Burke, D.B., Hillhouse, J.W., McKee, E.H., Miller, S.T., and Morton, J.L., 1982, Cenozoic rocks in the Barstow basin area of southern California—Stratigraphic relations, radiometric ages, and paleomagnetism: *U.S. Geological Survey Bulletin* 1529-E, p. 1-16.
- Byers, F.M., 1960, *Geology of the Alvord Mountain Quadrangle, San Bernardino County, California*: U.S. Geological Survey Bulletin 1089A, 71 p.
- Cox, B.F., and Wilshire, H.G., 1993, *Geologic map of the area around the Nebo Annex, Marine Corps Logistics Base, Barstow, California*: U.S. Geological Survey Open-File Report 93-568, 36 p.
- DeLeen, J., 1950, *Geology and mineral deposits of the Calico Mining District*: Unpublished Master's Thesis, University of California-Berkeley, 86 p.
- Dibblee, T. W., Jr., 1967, *Areal geology of the western Mojave Desert, California*: U. S. Geological Survey Prof. Paper 522, 153 p.
- Dibblee, T. W., Jr., 1970, *Geologic map of the Daggett Quadrangle, San Bernardino County, California*: U. S. Geological Survey Misc. Geological Inv. Map I-592.
- Dibblee, T. W., Jr., 1980a, Cenozoic rock units of the Mojave Desert, in Fife, D.L., and Brown, A.R., eds., *Geology and mineral wealth of the California desert: Dibblee Volume*: Santa Ana, California, South Coast Geological Society, p. 41-68.
- Dibblee, T. W., Jr., 1980b, *Geologic structure of the Mojave Desert*, in Fife, D.L., and Brown, A.R., eds., *Geology and mineral wealth of the California desert: Dibblee Volume*: Santa Ana, California, South Coast Geological Society, p. 69-100.
- Dokka, R.K., 1979, *Styles and timing of late Cenozoic faulting, central Mojave Desert, California*: Geological Society of America Abstracts with Programs, Annual Meeting, p. 414.
- Dokka, R.K., 1983, *Displacements on late Cenozoic strike-slip faults of the central Mojave Desert, California*: *Geology*, v. 11, p. 305-308.
- Dokka, R.K., 1986, *Patterns and modes of early Miocene crustal extension, central Mojave Desert, California*, in Mayer, L., ed., 1986, *Extensional tectonics of the southwestern United States: A perspective in processes and kinematics*: Geological Society of America Special Paper 208, p. 75-95.
- Dokka, R.K., 1989, *The Mojave extensional belt of southern California*: *Tectonics*, v. 8, p. 363-390.
- Dokka, R.K., and Travis, C.J., 1990a, *Late Cenozoic strike-slip faulting in the Mojave Desert, California*: *Tectonics*, v. 9, p. 311-340.
- Dokka, R.K., and Travis, C.J., 1990b, *Role of the Eastern California shear zone in accommodating Pacific-North American plate motion*: *Geophysical Research Letters*, v. 17, p. 1323-1326.
- Dokka, R.K., and Woodburne, M.O., 1986, *Mid-Tertiary extensional tectonics and sedimentation, central Mojave Desert, California*: *L.S.U. Publications in Geology and Geophysics-Tectonics and Sedimentation*, no. 1, 55p.
- Fletcher, D. I., 1986, *Geology and genesis of the Waterloo and Langtry silver-barite deposits, California*: Unpublished Ph.D. Dissertation, Stanford University, 158 p.
- Fletcher, J.M., Bartley, J.M., Martin, M.W., Glazner, A.F., and Walker, J.D., 1995, *Large-magnitude continental extension: An example from the Central Mojave metamorphic core complex*: *Geological Society of America Bulletin*, v. 107, p. 1468-1483.
- Glazner, A. F., Bartley, J. M., and Walker, J. D., 1988, *Geology of the Waterman Hills detachment fault, central Mojave Desert, California* in D.L. Weide and M.L. Faber eds., *This Extended Land Fieldtrip Guidebook: Las Vegas, Nevada*, Cordilleran Publishers, p. 225-237.
- Glazner, A. F., Bartley, J. M., and Walker, J. D., 1989, *Magnitude and significance of Miocene crustal extension in the central Mojave Desert, California*: *Geology*, v. 17, p. 50-53.
- Glazner, A.F., Walker, J.D., Bartley, J.M., and Fletcher, J.M., 2002, *Cenozoic evolution of the Mojave block of southern California*, in Glazner, A.F., et al., eds., *Geological evolution of the Mojave Desert and southwestern Basin and Range*: Geological Society of America Memoir 195, p. 19-41.
- Ireland, Mary C., 1993, *Barite mineralization, alteration and geochemistry of the Waterman Mine, San Bernardino County, CA*: Unpublished Senior Thesis, California Polytechnic Univ.-Pomona, 40 p.
- Jessey, D. R., 1986, *A geologic investigation of the Leviathan-Silver Bow property, Calico mining district, San Bernardino County California*: Geological Society of America Abstracts with Programs, Cordilleran Section, p. 121.
- Jessey, D. R., and Tarman, Donald W., 1988, *Geology of the Calico Mountains in Geologic Excursions in the Eastern Mojave Desert*, Lori Gaskin editor: National Association of Geology Teachers Far Western Section, Spring Conference, p 1-20.
- Jessey, D. R., and Tarman, Donald W., 1994, *Implications of fluid inclusion P/T data from Miocene hydrothermal barite deposits, central Mojave Desert, San Bernardino County, California*: *Quarterly of San Bernardino County Museum Association*, 1994, v. 41, p. 22-23.
- Jessey, D. R., and Yamashiro, J.A., 1988, *Structure and stratigraphy of a portion of the western Calico Mountains in Geologic Excursions in the Eastern Mojave Desert*, Lori Gaskin editor: National Association of Geology Teachers Far Western Section, Spring Conference, p 57-68.

- McCulloh, T. H., 1952, Geology of the southern half of the Lane Mountain Quadrangle, California: Unpublished Ph.D. Dissertation, University of California-Los Angeles, 182 p.
- McCulloh, T. H., 1965, Geologic map of the Nebo and Yermo Quadrangles, San Bernardino County, California: U. S. Geological Survey Open-File Map OFR-65-107.
- Mero, A. L., 1972, Geology and ore deposits of the south-central Calico Mountains: Unpublished Master's Thesis, Calif. State University-San Diego, 74 p.
- Payne, J. G., and Glass, J. R., 1987, Geology and silver deposits of the Calico district, San Bernardino County, California: Guidebook for field trips to bulk mineable precious metal deposits, Geological Society of Nevada, pp. 31-44.
- Rosso, K. M., 1992, Fluid inclusion microthermometry and geochemistry of the silver-barite mineralization in the Calico mining district, San Bernardino County, CA: Unpublished Senior Thesis, California Polytechnic Univ.-Pomona, 32 p.
- Rosso, K. M., Jessey, D. R., and Tarman, D. W., 1992, Fluid inclusion geothermometry and geochemistry of silver-barite mineralization in the Pickhandle Formation, Calico Mountains, San Bernardino County, California: Geological Society of America Abstracts with Programs, Annual Meeting, p. 79.
- Singleton, J.S. and Gans, P.B., 2008, Structural and stratigraphic evolution of the Calico Mountains: Implications for early Miocene extension and Neogene transpression in the central Mojave Desert, California: *Geosphere*, v. 4, p. 459-479.
- Smith, G. H., 1943, The history of the Comstock Lode 1850-1920: Nevada Bureau of Mines Bulletin 37, 305 p.
- Tarman, D. W., 1988, Synopsis of the stratigraphy and structure of the central Mojave Desert, in *Geologic Excursions in the Eastern Mojave Desert*, Lori Gaskin editor: National Association of Geology Teachers Far Western Section, Spring Conference, p 21-30.
- Tarman, D. W., and Thompson, D.M., 1988, Folding of the Barstow Formation in the southern Calico Mountains, in *Geologic Excursions in the Eastern Mojave Desert*, Lori Gaskin editor: National Association of Geology Teachers Far Western Section, Spring Conference, p 31-42.
- Walker, J.D., Fletcher, J.M., Fillmore, R.P., Martin, M.W., Taylor, W.J., Glazner, A.F., and Bartley, J.M., 1995, Connection between igneous activity and extension in the central Mojave metamorphic core complex: *Journal of Geophysical Research*, v. 100, p. 10,477–10,494
- Weber, F. H., 1965, Reconnaissance of silver-barite deposits of the Calico Mountains and vicinity: California Division of Mines and Geology Open File Map, Los Angeles.
- Weber, F. H., 1966, Silver mining in Old Calico: Calif. Division of Mines and Geology Mineral Info. Service, v. 19, pp. 71-80.
- Woodburne, M.O., Tedford, R.H., and Swisher, C.C., 1990, Lithostratigraphy, biostratigraphy, and geochronology of the Barstow Formation of the Mojave Desert, southern California: *Geological Society of America Bulletin*, v. 102, p. 459–477.

Biota of the Willow Tank Formation, late Early Cretaceous (Albian) from southern Nevada

Joshua W. Bonde¹, David J. Varricchio², and Frankie D. Jackson²

¹Geoscience Department, University of Nevada, Las Vegas bondej@unlv.nevada.edu

²Earth Sciences Department, Montana State University, Bozeman MT, djv@montana.edu, frankiej@montana.edu

In 1999 the Willow Tank Formation of southern Nevada, Lower Cretaceous, was found to be vertebrate-bearing. Exposures of this formation are found in the North Muddy Mountains, Gale Hills, Bowl of Fire, and the Virgin Mountains (Bohannon, 1983). This study focuses on exposures from the North Muddy Mountains. The Willow Tank Formation represents deposits of an alluvial system deposited within the foredeep of the Sevier Foreland (Schmitt and Kohout, 1986; Schmitt and Ashoff, 2003; Bonde et al., 2008; Bonde, 2008). The presence of carbonate nodules suggests a seasonally arid climate during deposition of this unit (Bonde, 2008). In addition, Willis and Behrensmeier (1994) state that mottling and reddening of beds is indicative of a well drained flood plain environment, these features are observed within the Willow Tank Formation. Given the sedimentological context, the organisms preserved in southern Nevada lived in a seasonally arid well-drained floodplain proximal to the Sevier Fold and Thrust front.

Macro-plant fossils are preserved at two sites within exposures of the Willow Tank Formation in Valley of Fire State Park. One site preserves only two fern morphotypes. The first of these consists of a rachis of at least 55 mm in length with primary pinna extending up to 24 mm in length and becoming progressively shorter distally down the rachis. The primary pinna have secondary pinna with an amplitude of 1 mm and have complete articulation with

the primary pinna. This morphology is consistent with the morpho-genus *Cladoplebis* (figured in Tidwell, 1998). The second morphotype consists of a rachis up to 42 mm in length with primary pinna approximately 24 mm in length which are fully attached to the rachis and are of

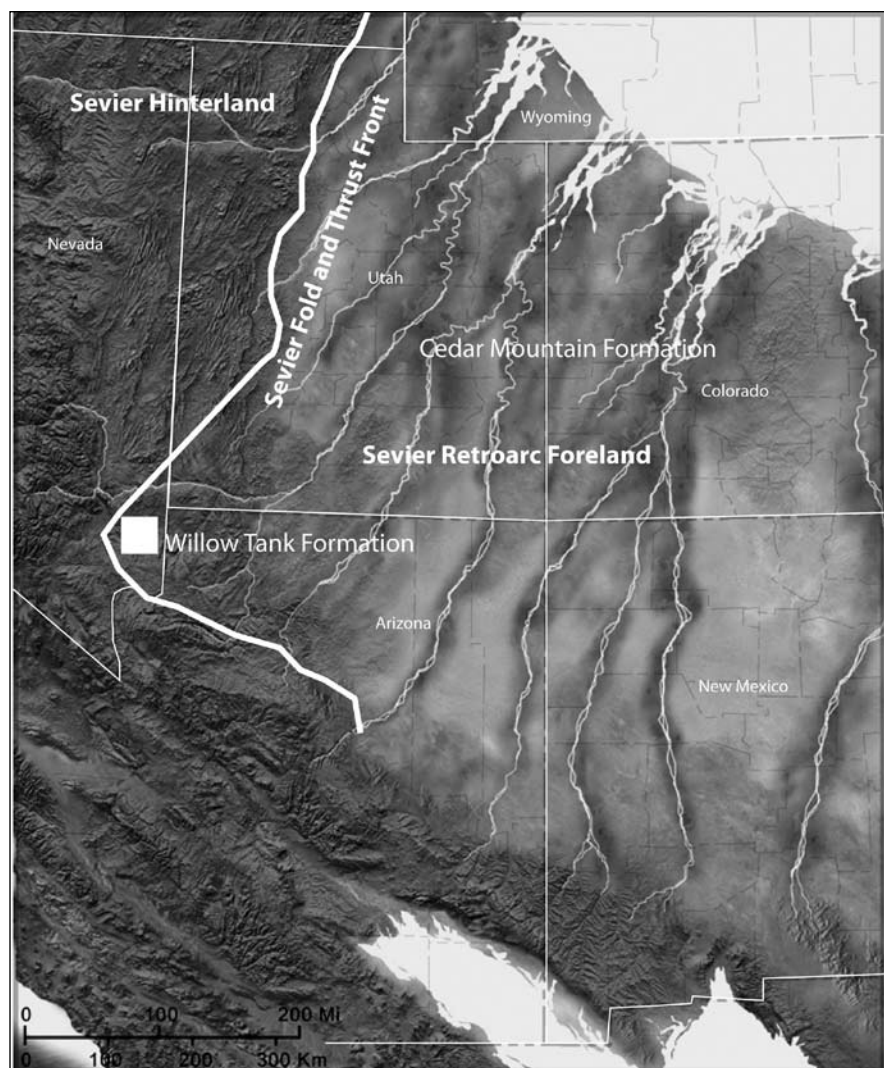


Figure 1. Paleogeography of the southwest (modified from Ron Blakey, NAU Geology) during late Early Cretaceous time, showing approximate relationship of the Willow Tank Formation to the Sevier Fold and Thrust Front (roughly after Currie, 2002).



Figure 2. *Cladoplebis* sp.

equal length from proximal to distal along the rachis. This morphotype is consistent with the genus *Matonia* (figured in Tidwell, 1998). The second site preserves a number of angiosperm leaves and reeds of sphenophytes. The most common sphenophyte has reeds >50 mm in length and up to 10 mm wide with parallel laminae along the length of the reed. These laminae are occasionally interrupted by joints, a morphology consistent with *Equisetum*. There are

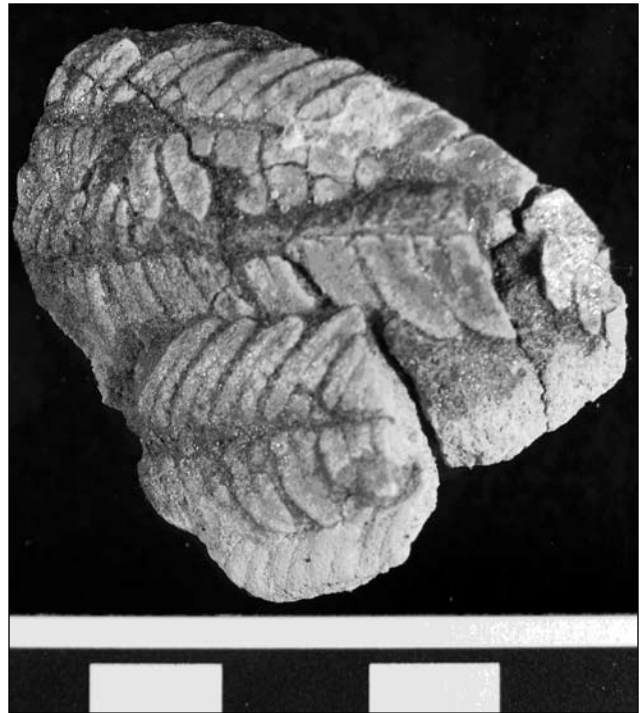


Figure 3. *Matonia* sp.

at least 5 angiosperm morphotypes, of which only two are currently identified. The first morphotype consists of a primary vein running the middle of the leaf with secondary veins staggered off of that and in

some cases tertiary veins at the very tips of the secondary veins. Secondaries are concave up. This leaf has an entire margin without serrations. This morphology is consistent with the morpho-genus *Magnolia*. The second leaf morphotype consists of groups of up to three leaflets. Each leaflet has a prominent primary vein with alternate secondary veins. The margins of these leaflets are entire. This morphotype is typical of the morpho-genus *Sapindopsis*, a common angiosperm plant of this time period. *Tempskya* has also been reported from the Willow Tank Formation (Ash and Read, 1976), but has not been observed by the authors. Based upon lithofacies assemblages and sedimentological data, the first site represents a floodplain setting which was buried in ash; the second site represents a floodplain pond which was buried by ash.

Of the vertebrates, Osteichthyes, Chelonia, Crocodyloformes, Ornithischia, and Saurischia are preserved.

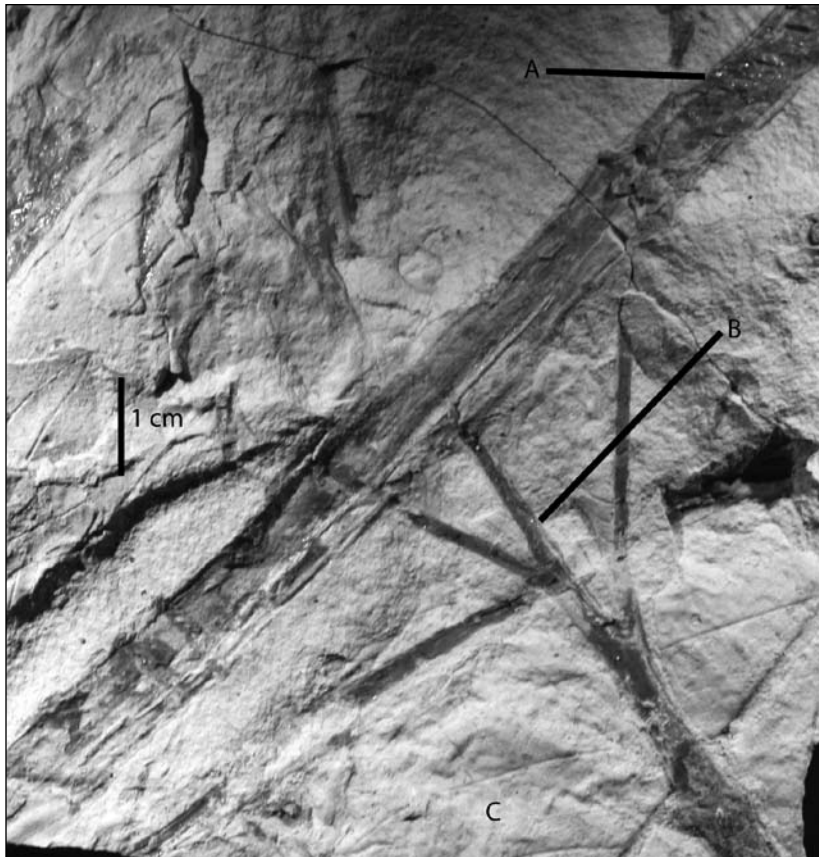


Figure 4. A. *Equisetum* sp., B. *Sapindopsis* sp., C. *Magnolia* sp.

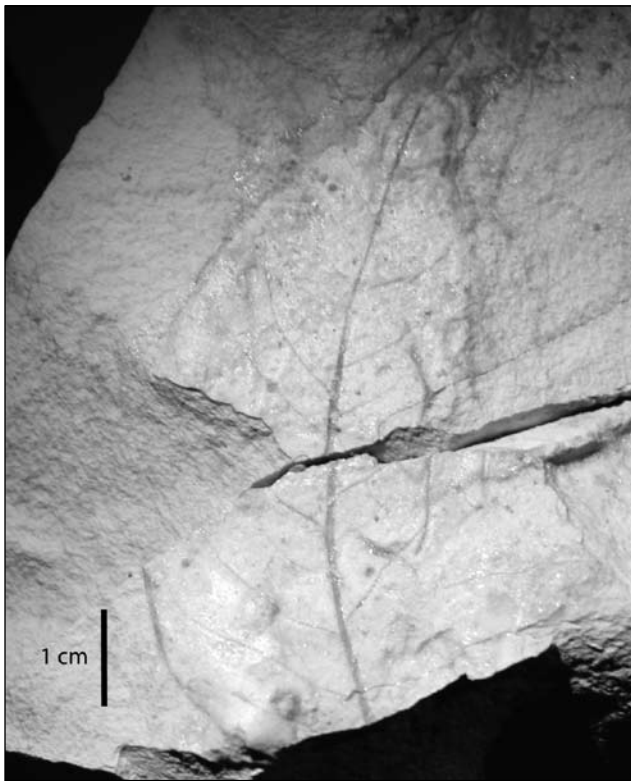


Figure 5. c.f. *Magnolia* sp.

One fish taxon is preserved as rhombohedral, enameled scales, consistent with the Lepisosteidae. The second is preserved as polygonal, enameled scales with a peg and socket articulation. This morphology is consistent with the Holostean A designation of Brinkman (1990). The third fish taxon is preserved as two tooth plates. These have a sharply undulatory pattern and are roughly triangular in shape. The ridges of the undulations come together at one corner of the triangle. This morphology is consistent with the Dipnoan, *Ceratodus*.

The group Chelonia is the most abundant group found within the Willow Tank Formation. There are 3, perhaps 4, different turtles from this formation. The first turtle has a carapace with pustule ornamentation: the pustules are random in distribution and are found over the entire carapace. This ornamentation is consistent with *Naomichelys speciosa*. The next turtle has an ornamentation which is weakly serpentine, and vein impressions along the carapace are more pronounced than ornamentation. This turtle is similar in morphol-

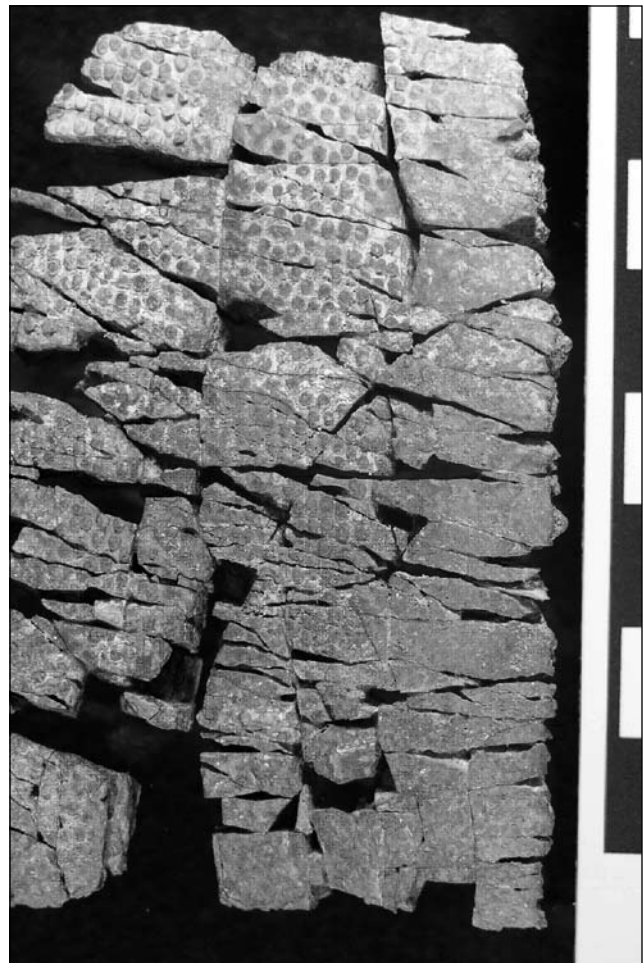


Figure 6. *Naomichelys speciosa*



Figure 7. Unknown baenid.

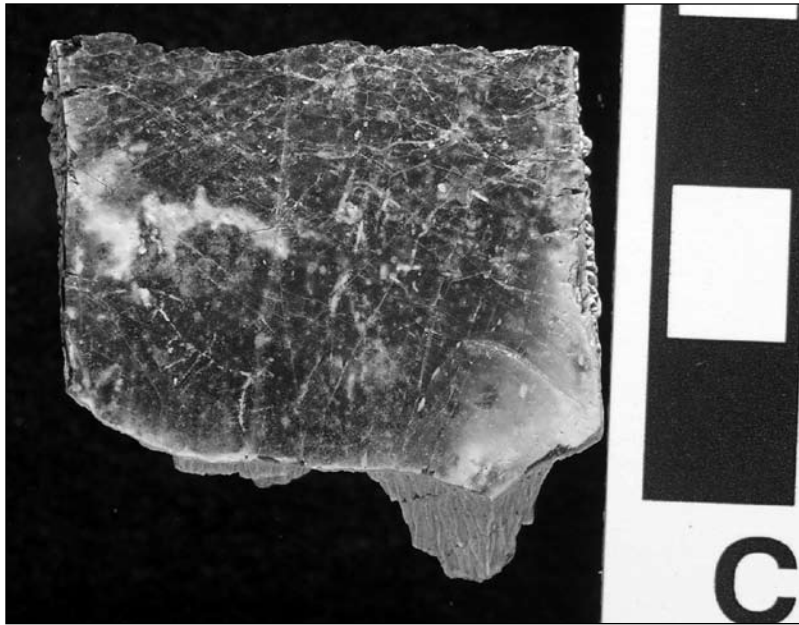


Figure 8. c.f. *Adocus* sp.

ogy to *Glyptops* but approaches the morphology of the Baenidae and may represent an intermediate form. A third turtle is represented by shell with a dimpled ornamentation. The “dimples” run in linear rows along the length of carapace elements and are very subtle. This morphology is similar to *Adocus* (Hay, 1908). The final, potential turtle type has deeply pitted ornamentation consistent with the Trionychoidea, but given the extremely fragmentary nature of the material a confident assignment cannot be made



Figure 9. c.f. *Goniopholis* sp.

at this time. If it were a true Trionychoidea, it would be the earliest occurrence of this taxon within North America.

Another group present in the formation is the Crocodyloformes. This group is represented by a number of different teeth and a single osteoderm. The teeth range from caniniform to short, round, and blunt. Teeth range from 4 cm to a few millimeters in length. The osteoderm has deeply pitted and irregular ornamentation on the dorsal surface and shows a number of annuli on the ventral surface. According to D. Fowler (pers. comm., 2005) the osteoderm is similar to those of *Goniopholis*.

There are four ornithischians from the Willow Tank Formation. The first type is preserved as two teeth. These teeth have a triangular shape with a sharp ridge around the base of the tooth. These teeth are similar

to those of the Thyreophora. Teeth of this group are exceptionally primitive and designation below the family level is not possible at this time. There are a number of ornithomimid elements from the Willow Tank Formation. The first ornithomimid is preserved as a pre-pubic process to the pubis, associated ribs, and ossified tendons. This material is not diagnostic to any particular taxon of ornithomimid.

The next ornithomimid is preserved as a series of dorsal vertebrae, two femora, and a number of phalanges and metapodials. The size and morphology of these elements is similar to the Hypsilophodontidae. A number of isolated teeth have been recovered that are up to 2–3 cm in width and 3–4 cm in length. These teeth are very robust and possess two prominent ridges up the axis of the tooth. This morphology is consistent with basal iguanodontians. Also from the Willow Tank Formation is a dental battery with teeth far less robust than the previously-described teeth. The teeth within the dental battery are 2 cm in width and up to 4 cm in length and possess a single prominent ridge down the middle



Figure 10 (above).
Thyreophoran tooth.



Figure 11 (right).
Iguanodontian pre-
pubic process.

of the tooth. The margins of the

teeth have rounded denticles. For each tooth family there are at least two replacement teeth in line with the active tooth. This state is a characteristic of an advanced iguanodontian (Horner et al., 2003).

Three saurischians have been recovered so far from the Willow Tank Formation. The first saurischian is preserved as a single tooth. This tooth is ~8 mm in length and is subcylindrical with a triangular wear facet. This is diagnostic of a basal titanosauriform. The next saurischian is also preserved as a single tooth. This tooth is ~8 mm in length with a D-shaped cross section. The lingual side has a prominent ridge which runs up the axis of the tooth and becomes less pronounced distally. On either side of this ridge are two carina which run the length of the tooth and become more pronounced distally, with denticles. This morphology is consistent with a premaxillary tooth from the Tyrannosauroidea. The final saurischian is represented by an associated femur and tibia as well as a number of teeth. The femur is slightly bowed and has a femur head which is perpendicular to the diaphysis of the bone. Teeth are laterally compressed with serrations on the posterior side of the tooth. These traits are consistent with the Dromaeosauridae.



Figure 12. Hypsilophodont elements.



Figure 13. Iguanodont tooth.



Figure 14. Hadrosauroid dental battery.

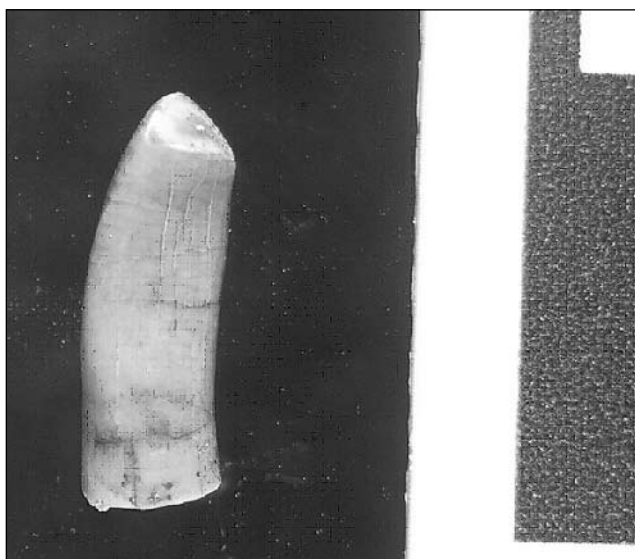


Figure 15. Titanosauriform tooth.

In addition to the body fossils two, perhaps three, egg shell morphotypes have been discovered. All eggshell material comes from paleosol horizons. Based upon SEM images of the cross-sections of these eggs, images show that they are all from theropods. One morphotype is identified as *Macroelongatoolithus*, known from Utah and Asia. Another morphotype is not identifiable, while the final morphotype is extremely thin and may actually be avian rather than non-avian theropod.

The significance of these finds, in addition to being the most complete assemblage from the state of Nevada from the Cretaceous, is that this biota occurs extremely close to the Sevier fold and thrust front and extends the range of Early Cretaceous taxa to the southwest. Similarly aged fauna from the Cedar Mountain Formation of Utah have been divided into three biozones, an Aptian–Albian

fauna, an Albian Fauna, and a latest Albian–early Cenomanian fauna (Kirkland et al., 1998). The fauna of the Willow Tank Formation do not conform to any one of the biozones of the Cedar Mountain Formation. Rather, fauna from the Willow Tank Formation consists of taxa which are present in each of the three biozones from the Cedar Mountain Formation. Basal Iguanodontia are indicative of the lowermost biozone, Titanosauriformes are indicative of the middle biozone, and Tyrannosauroida and advanced Iguanodontia are indicative of the uppermost biozone. Remains of basal iguanodontia are found upsection from the remains of the advanced iguanodontians. This is the first record, from North America, of basal and advanced iguanodontians temporally overlapping in a region. Also, an Albian occurrence of advanced iguanodontians and tyrannosauroids is the earliest occurrence of these taxa in North America. The uppermost biozone of the Cedar Mountain Formation, in addition to having tyrannosauroids and advanced iguanodontians, has pachycephalosaurs and ceratopsids. The Willow Tank Formation does not have either of these marginocephalians; this may represent a pulsed immigration from Asia as opposed to one event. Plant fossils are consistent with other floras of the time period by the dominance of the angiosperm *Sapindopsis* and are consistent with the pollen stage 2C of the Potomac group which, not only observed on the east coast,

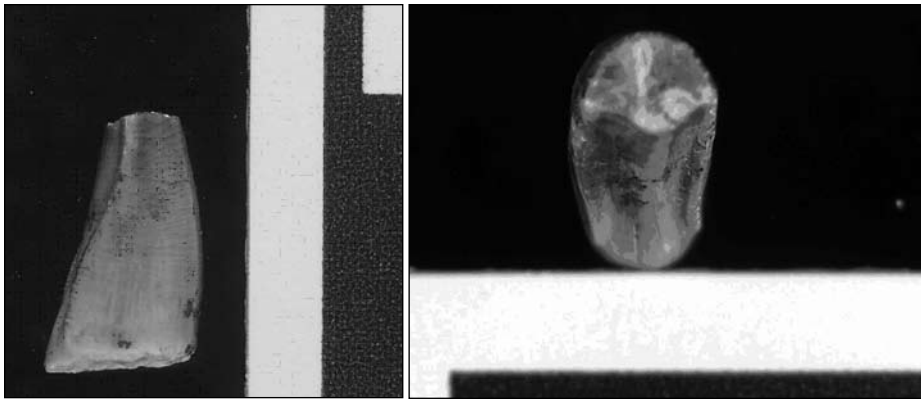


Figure 16. Tyrannosauroid premaxillary tooth. A. Lateral View, B. Occlusal View.



Figure 17. Femur of a dromaeosaurid.

is also seen in the Northern Rocky Mountains (Hickey and Doyle, 1977; Crabtree, 1987). In addition, the turtle species *Naomichelys speciosa* is also found nearly continent wide during this time period which is an extremely wide range for a vertebrate as small as a turtle. Likewise, the widespread distribution of a similar flora at roughly the same time may imply that there was a unique paleoecology that allowed for these continent-wide distributions of taxa which is not seen today. The dinosaurs, at the familial level, are also similar across the continent at this time and agree with the turtle and plant data. Research is ongoing on this formation and, with further study, more precise taxonomic assignments can be made and the relationships between differing biotas, temporally and geographically, can be analyzed.

References

- Ash, S.R. and Read, C.B., 1976, North American species of *Tempskya* and their stratigraphic significance: United States Geological Survey Professional Paper 874.
- Bohannon, R.G., 1983, Mesozoic and Cenozoic tectonic development of the Muddy, North Muddy, and northern Black Mountains, Clark County, Nevada, D.M. Miller, V.R. Todd, and K.A. Howards eds.: Tectonic and stratigraphic studies in the eastern Great Basin, Geological Society of America Memoir, v. 157, p. 125-148.
- Bonde, J.W., Varricchio, D.J., Jackson, F.D., Loope, D.B., and Shirk, A.M., 2008, Dinosaurs and Dunes! Sedimentology and paleontology of the Mesozoic in the Valley of Fire State Park, Geological Society of America Field Guide 11, p. 249-262.
- Bonde, J.W., 2008, Paleocology and taphonomy of the Willow Tank Formation (Albian), southern Nevada, MS Thesis Montana State University, Bozeman, MT, 107 p.
- Brinkman, D.B., 1990, Paleocology of the Judith River Formation (Campanian) of Dinosaur Provincial Park, Alberta, Canada: Evidence from vertebrate microfossil localities: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 78, p. 37-54.
- Crabtree, D.R., 1987, Angiosperms of the northern Rocky Mountains: Albian to Campanian (Cretaceous) megafossil floras, Annals of the Missouri Botanical Garden, v. 74, p. 707-747.
- Hay, O.P., 1908, The fossil turtles of North America: Carnegie Institute of Washington Publication 75.

- Hickey, L.J. and Doyle, J.A., 1977, Early Cretaceous fossil evidence for angiosperm evolution, *The Botanical Review*, v. 43, p. 1-102.
- Horner, J.R., Weishampel, D.B., and Forster, C.A., 2004, Hadrosauridae, D.B. Weishampel, P. Dodson, and H. Osmolska eds., *The Dinosauria* 2nd ed., p. 438-463.
- Kirkland, J.I., Lucas, S.G., and Estep, J.W., 1998, Cretaceous dinosaurs of the Colorado Plateau, S.G. Lucas, J.I. Kirkland, and J.W. Estep eds., *Lower and Middle Cretaceous terrestrial ecosystems: New Mexico Museum of Natural History and Science bulletin*, v. 14, p. 79-90.
- Schmitt, J.G. and Aschoff, J.L., 2003, Synorogenic sedimentation along the Cordilleran thrust belt front, southern Nevada: alluvial systems, sediment dispersal, and growth strata development: *Geological Society of America Annual Meeting Abstracts with Programs*, v. 35(6), p. 642.
- Schmitt, J.G. and Kohout, J.B., 1986, Early Cretaceous sedimentary evolution of the Sevier Foreland Basin, Muddy and North Muddy Mountains, Nevada: *Geological Society of America Annual Meeting Abstracts with Programs*, v. 18(2), p. 181.
- Tidwell, W.D., 1998, *Common fossil plants of western North America*, Smithsonian Institution Press, Washington D.C., 299p.
- Willis, B.J. and Behrensmeyer, A.K., 1994, Architecture of Miocene overbank deposits in northern Pakistan, *Journal of Sedimentary Research*, v. B64, p. 60-67.

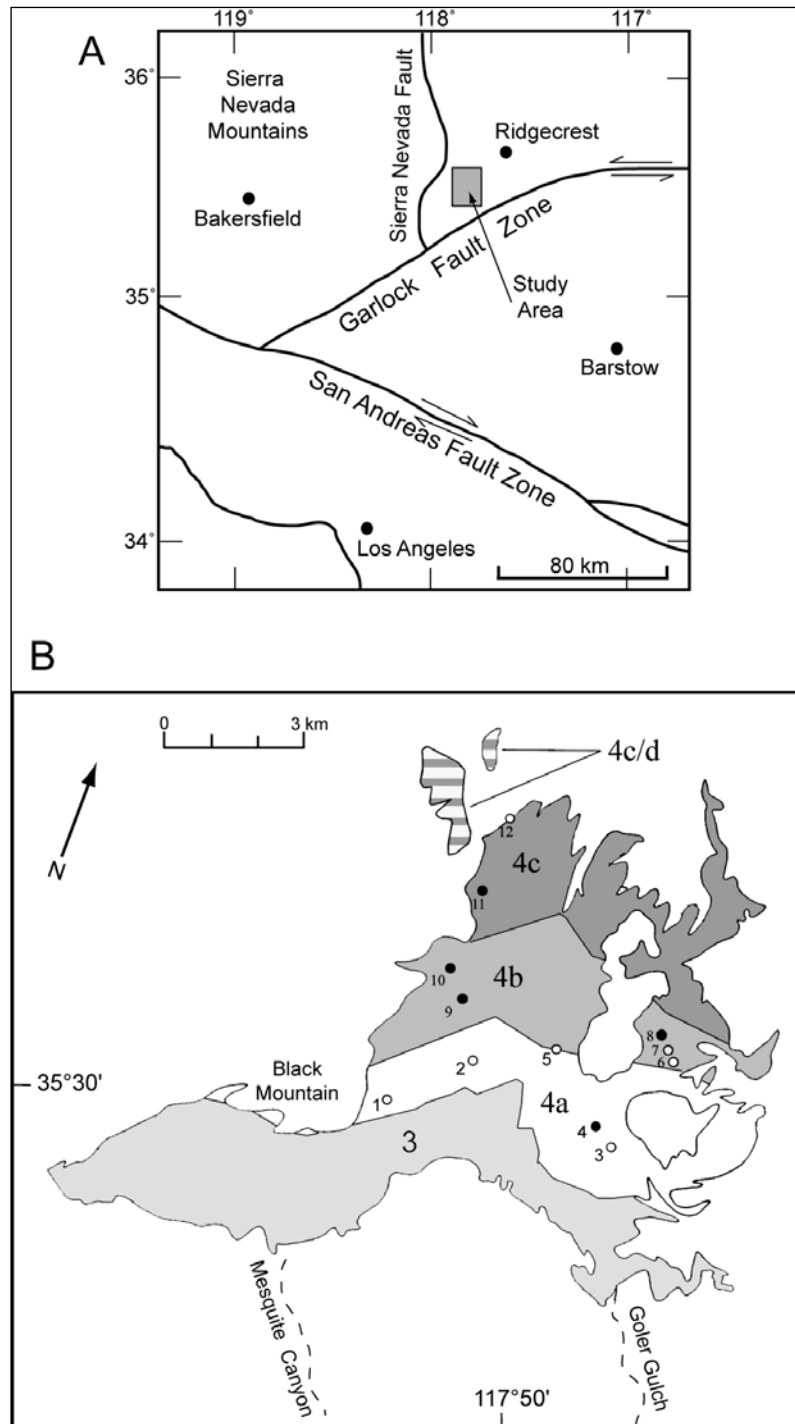
New marine sites from Member 4D of the Goler Formation of California

Donald L. Lofgren¹, Brent A. Silver², Theodora A. Hinkle², and Faizan Ali-Khan²
¹Raymond M. Alf Museum of Paleontology, 1175 West Baseline Rd. Claremont, CA 91711
²The Webb Schools, 1175 West Baseline Rd, Claremont, CA 91711

Introduction

The Goler Formation crops out in the El Paso Mountains in the northern Mojave Desert (Figure 1a). Underlain by Paleozoic and Mesozoic igneous and metamorphic rocks and overlain by sedimentary and igneous rocks of the Late Cenozoic Ricardo Group, the Goler Formation is over three kilometers thick and is subdivided into members (numbered one through four) comprised mainly of fluvial deposits (Dibblee 1952; Cox 1982, 1987). Originally, Goler sediments were thought to be Late Eocene to Miocene in age based on fossil plants (Axelrod 1949) and other evidence (Dibblee 1952) as the formation was thought to be virtually barren of invertebrate and vertebrate fossils. This misconception slowly changed based on collecting activities that began in the 1950s and extend to the present. Fossils are now known from members 3 and 4, although specimens from member

Figure 1a: Index map showing location of study area. Figure 1b: Outcrop map of the Goler Formation (adapted from Cox 1982; Cox and Diggles 1986) showing location of major fossil sites; members 1-2 not shown. Open circles are where strata underwent paleomagnetic analysis. Sites are numbered in approximate stratigraphic order; RAM locality numbers are given for all sites except 9 and 11. 1) Grand Canyon (V200510); 2) Land of Oz (V200001); 3) Phenacodus Pocket (V200612) and Honey Ridge (V200603); 4) R-Z Zone (V200304); 5) Laudate Discovery Site (V91014); 6) Edentulous Jaw Site (V98012); 7) Primate Gulch (V200202); 8) Lone Tooth (V200704); 9) UCMP locality V65170; 10) Honey Pot (V200120); 11) Cox-McDougall-Squires Marine Site; 12) Alf Museum Marine Sites (I200305, I200306, V200307).



3 are extremely rare. Member 4 is subdivided into four units, 4a, 4b, 4c, and 4d (Figure 1b), and samples of fossils have been collected from each unit except Member 4c.

Vertebrate fossils, especially mammals, have helped determine the age of the Goler Formation. It began in 1954 when M. McKenna found a mammal jaw at the Laudate Discovery Site (#5 in Figure 1) which indicated a Paleocene age for that part of the formation (McKenna 1955). Over the next 35 years only about a dozen identifiable vertebrate specimens were found, but they corroborated the Paleocene age of Member 4a and 4b (McKenna 1960; West 1970; McKenna et al. 1987). Recent work over the past two decades has yielded a large sample of Goler vertebrates from members 4a and 4b through intensive prospecting efforts and the employment of screen-washing techniques. Many of the taxa recovered were new records for the Goler Formation, including species of turtles, rays, lizards, crocodylians, and mammals (Lofgren et al. 1999, 2002, 2008; McKenna and Lofgren 2003; McKenna et al. 2008). The mammalian sample indicates that members 4a and 4b are Middle–Late Paleocene (Tiffanian Land Mammal Age) based on occurrences of age diagnostic species of plesiadapid primates, multituberculates, condylarths, and carnivores (Lofgren et al. 2008; Albright et al. in press).

In the 1980s, an interval of marine sediments (#11 in Figure 1b) was discovered in the uppermost part of the Goler Formation in Member 4d (Cox and Edwards 1984; Cox 1987). Initial collections of marine invertebrates from this interval yielded coccoliths, foraminifera, and mollusks. The age of Member 4d was tentatively interpreted to be late Paleocene (CP8) based on coccoliths (Reid and Cox 1989), late Paleocene–early Eocene based on mollusks (Squires et al. 1988), and early Eocene (P8 and ?P9) based on foraminifera (McDougall 1987).

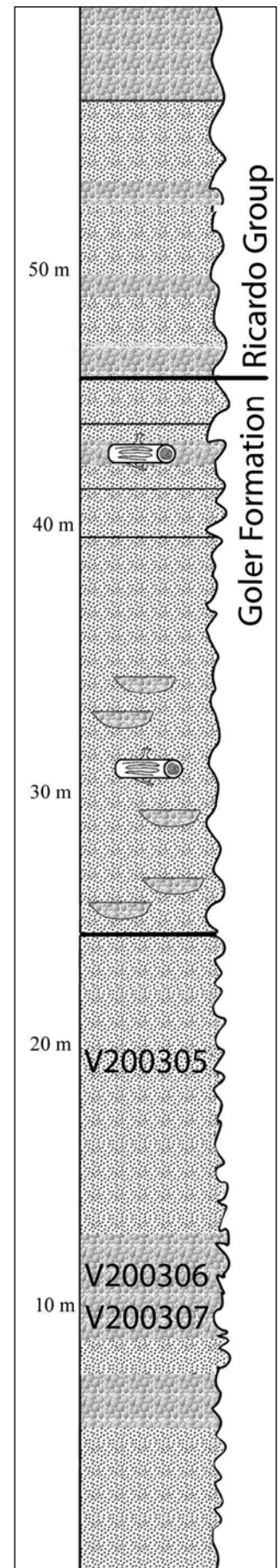
The initial mollusk sample from Member 4d totaled fifty-one specimens and all were molds or casts except for nineteen specimens of oysters (Squires et al. 1988). Recent work in the Goler Formation by the Raymond Alf Museum of Paleontology resulted in the discovery of three sites in Member 4d (#12 in Figure 1b) that yielded abundant and well-preserved marine invertebrates along with shark teeth, fossil wood, plant material, and a tooth of a terrestrial mammal. The discovery of new marine sites is an opportunity to further elucidate the Member 4d marine fauna and to describe a new section of Goler marine rocks.

Materials and methods

Since the early 1990s, Alf Museum crews have been systematically prospecting outcrops of Member 4 of the

Goler Formation. In 2003 R. Nydam found a large chunk of burrowed petrified wood on the north end of the El Paso Mountains. Subsequent searching of this site yielded numerous small specimens of well-preserved marine gastropods and pelecypods that were encased in eroded remnants of what apparently was a lens of sandstone. The discovery of this important new site (RAM I200305) spurred a return to the area a month later where two sites (RAM I200306 and RAM V200307) that also yielded marine mollusks were found about 100–200 meters to the east of I200305. I200306 yielded abundant oyster shells, casts and molds of other mollusks, and fragmentary plant fossils. V200307 yielded a gastropod, a tooth of a terrestrial mammal, abundant oysters, and shark teeth—the first record of shark teeth from the Goler Formation. A section was measured through strata containing the three new sites and up into rocks of the overlying Ricardo Group using a hand level and Jacobs staff (Figure 2).

Figure 2: Stratigraphic section measured through outcrops of the Goler Formation that contain the Alf Museum marine sites (#12 in Figure 1b). Dark lines in the section denote location of a fault (contact between Goler Formation and Ricardo Group) or a possible unconformity (23 m level in Goler Formation).



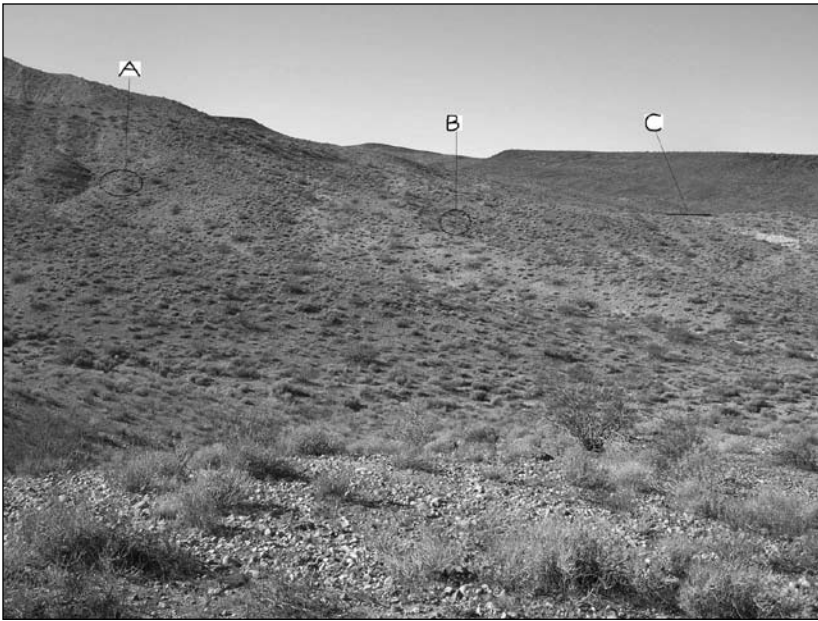


Figure 3: photo looking southwest at Goler marine outcrops with position of fossil sites noted by arrows; a) I200306, b) V200307, c) I200305

Preservation of mollusks from I200306 and V200307 is similar to that described by Squires et al (1988) from their main mollusk localities (M 9066, M 9067, M 9068) where fossils are molds and casts except for oysters. Specimens from I200305 are composed mostly of shelly material. Mollusk specimens from the three new sites were compared to mollusks described by Squires et al. (1988).

Digital calipers were used to measure teeth and shells. Dimensions for specimens shown in figures 7–14 are noted in the text description of each specimen. Measurements are in millimeters (mm); cm = centimeters, m = meters. The Raymond M. Alf Museum of Paleontology (RAM) records each vertebrate or invertebrate locality by using a V or I followed by numbers (e.g., V200305 or I200305). Specimens are given a numerical code preceded by RAM (e.g., RAM 7253). United States Geological Survey locality numbers begin with an M followed by four numbers (e.g. M 9067). UCMP refers to the University of California Museum of Paleontology. Premolars from the mammal jaw are denoted by a lower case p with a number following the p designating which premolar it represents (e.g. p4 would be a 4th lower premolar). RAM specimen numbers are provided for all taxa discussed in the text with the exception of specimens of *Acutostrea idriaensis* and *Turritella*

buwaldana which were too numerous (over 130) to list individually.

Stratigraphy and geology of newly discovered marine strata

Strata identified as marine rocks outcrop on the north end of the El Paso Mountains where the outcrop area is striped in Figure 1b. Marine rocks were also reported by Cox and Edwards (1984) and Cox (1987) from a small outcrop noted as number 11 on Figure 1b where virtually all the mollusks described by Squires et al (1988) were collected. The new Goler marine fauna reported here (#12 in Figure 2) is situated roughly between the previously identified marine outcrops and was mapped Member 4c by Cox and Diggles (1986).

Sites I200306 and V200307 occur near the head of gulleys on north-facing outcrops, while I200305 is southwest of V200307 on a low, south-facing exposure (Figure 3). I200306 and V200307 are at the same stratigraphic level (Figure 2). I200306 is situated in a two meter interval within a series of thick-bedded conglomerate and sandstone-pebble conglomerate lenses that exhibit cut and fill structures (Figure 4). Conglomerates are clast-supported with clasts well rounded and exceeding 10 cm in diameter in some cases. Angular rip-up clasts of green mudstone (marine?) are abundant locally. Slightly abraded to pristine oyster



Figure 4: Outcrop at I200306 showing erosive contact between sandstone wedge and conglomerate.



Figure 5: Conglomerate at V200307 where shark teeth and a nonmarine mammal tooth were found.

shell fragments to whole valves are common in conglomerate lenses while oysters (shelly material) and other mollusks (casts, and molds) and plant fragments are present

in sandstones. One thick conglomerate is laterally continuous and can be traced from I200306 west to V200307 where it exceeds two meters in thickness (Figure 5). Well-preserved oysters dominate the fauna at V200307 and unabraded whole valves are common. One of the three shark teeth (RAM 7256) recovered at V200307 was discovered in situ, partially exposed between two rounded cobbles.

I200305 is about seven meters stratigraphically above V200307. Pieces of what apparently once was a lens of indurated sandstone lie eroded on a slope of siltstones and sandstones. Abundant well-preserved mollusks occur in the sandstone lens and specimens are all of similar small size, indicating sorting had occurred.

Additional Goler Formation strata, higher in elevation and 50 m to the southwest of I200305, were measured and these rocks may represent a continuation of the sedimentary section exposed at I200305. This is uncertain as these sedimentary

Figure 6: Table showing the number of pelecypods, gastropods, and other organisms recovered from sites I200305, I200306, and V200307. X denotes the presence of plant fragments or fossil wood.

	200305	200306	200307
Pelecypoda			
<i>Acutostrea idriaensis</i>	10	65	29
<i>Barbatia biloba</i>	0	1	0
<i>Corbula</i> aff. <i>C. dickersoni</i>	7	1	0
<i>Ledina duttonae</i>	0	1	0
<i>Nemocardium linteum</i>	12	0	0
<i>Thracia</i> aff. <i>T. condoni</i>	3	2	0
Gastropoda			
<i>Calyptrea diegoana</i>	3	0	0
<i>Turritella buwaldana</i>	22	4	1
Gastropod A	1	2	0
Gastropod B	1	0	0
Gastropod C	1	0	0
Gastropod D	1	0	0
Gastropod E	1	0	0
Gastropod F	1	0	0
Vertebrata			
<i>Phenacodus</i> sp. cf. <i>P. vortmani</i>	0	0	1
<i>Striatolamia</i>	0	0	3
Plantae			
Plant Material		X	
Fossil Wood	X		

rocks differ in color and texture from those below and are dominated by cross-bedded coarse sandstone with lenses of pebble-cobble conglomerate. Fossil wood is present but mollusks were not evident. These higher strata appear to be non-marine. The sequence of apparently non-marine strata above marine strata indicates that the contact between the units (23 m level in Figure 2) could be fault bounded or unconformable. Either scenario is uncertain as the critical interval is covered. At the 44.5 m level a fault is evident, which places sedimentary rocks of the Ricardo Group on older Goler strata (Figure 2).

Paleontology

Pelecypoda

The six pelecypod taxa identified in the RAM collections from the three new Member 4d sites (Figure 6) were described by Squires et al. (1988). In contrast to those described by Squires et al. (1988), pelecypods from the new sites are more numerous, and at I200305 many well-preserved specimens consisting of shelly material are present.

Acutostrea idriaensis

The oyster *Acutostrea idriaensis* is the most common mollusk in the Goler Formation with over 100 specimens recovered. Fragmentary valves are very common at I200306 and V200307 and whole or nearly whole valves can be found with careful searching. Only complete valves were collected from I200305. They have widths of 13–25 mm and heights of 19–36 mm. Valves from I200306 and V200307 are much larger and have widths of 24–66 mm and heights of 38–91 mm.

Barbatia biloba

Only one specimen of *Barbatia biloba* was positively identified. RAM 9633 was collected at I200306 and it measures 8.6 mm in width and 8.2 mm in height.

Corbula aff. *C. dickersoni*

Specimens of *Corbula* aff. *C. dickersoni* from I200305 have widths from 9.6–23.8 mm and heights from 7.0–12.4 mm (RAM 9567, 9593, 9594, 9611, 9563, 9610, 9576). RAM 9626 from I200306 has a width of 24.1 mm and a height of 15.6 mm.

Ledina duttonae

Only one specimen, RAM 9627 from I200306 of *Ledina duttonae* was recovered. It measures 19.1 mm in width and 14.1 mm in height.

Nemocardium linteum

Twelve specimens of *Nemocardium linteum* were recovered from I200305 (9573, 9614, 9565, 9582, 9585, 12248, 12249, 12250, 12251, 12252, 12253, 12254). Measure-

ments of these valves vary from 9.0–30.1 mm in width and 8.5–21.3 mm in height.

Thracia aff. *T. condoni*

Five specimens of *Thracia* aff. *T. condoni* were identified, three from I200305 (RAM 9601, 9615, 9586), two from I200306 (RAM 9619, 9625). Widths are 10.8–19.1 mm and heights are 7.5–14.1 mm.

Gastropoda

In contrast to the pelecypod fauna, only two of the eight types of gastropods recovered from the Alf Museum marine sites (Figure 6) were reported by Squires et al. (1988). Of the six new records, none could be tentatively identified to genus.

Calyptrea diegoana

Three specimens of *Calyptrea diegoana* were identified from I200305 (RAM 9584, 9599, 9613). These fossils have a size range of 13.8 to 18.4 mm in width and 6.2 mm to 9.8 mm in height.

Turritella buwaldana

The most abundant gastropod is *Turritella buwaldana*, with 27 identifiable specimens, including complete shells. Site I200305 produced twenty-two specimens with widths of 3.4 to 9.0 mm, heights from 7.4 to 15.5 mm. Sites I200306 and V200307 yielded five specimens, with widths of 5.0 to 22.5 mm, and heights of 14.7 to 47.0 mm.

Gastropod A

Three specimens were identified of this type of gastropod; two from I200306, RAM 9621 and RAM 9622 (Figure 7), the other from I200305 (RAM 9600). Measurements of these specimens range from 5.8–18.6 mm in width and 3.7–7.2 mm in height.



Figure 7: Gastropod A; RAM 9622



Figure 8: Gastropod B; RAM 9592

Gastropod B

RAM 9592 (Figure 8) is well preserved and shows five whorls that have well-developed ribs. It resembles *Admete*, but is some other genus. RAM 9592 is 6.7 mm in width and 12.0 mm in height.

Gastropod C

RAM 9591 (Figure 9) has seven whorls with well defined costa in the body whorl. The width of RAM 9591 is 5.9 mm and the height is 13.1 mm.

Gastropod D

RAM 9616 (Figure 10) is 7.8 mm in width and 6.5 mm in height and has little ornamentation on the shell. It



Figure 9: Gastropod C; RAM 9591

resembles *Gyrodes* in some ways but a tentative identification was not possible.

Gastropod E

RAM 9602 (Figure 11) has a width of 6.9 mm and a height of 13.5 mm. Only part of the shell profile is exposed, but it is clear that strong ribbing and costa are developed in the body whorl.

Gastropod F

RAM 9604 (Figure 12) is a cast with four whorls. It is 4.9 mm in width and 8.8 mm in height.



Figure 10: Gastropod D; RAM 9616



Figure 11: Gastropod E; RAM 9602



Figure 12: Gastropod F; RAM 9604

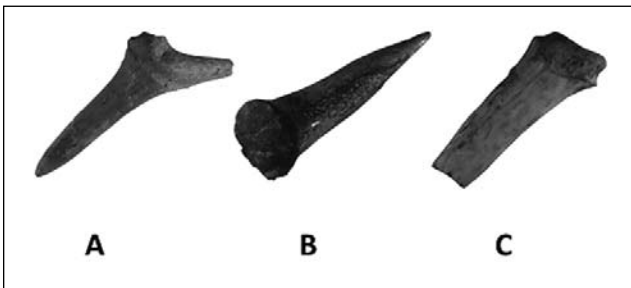
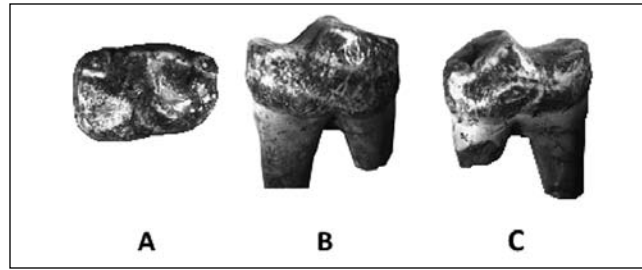
Chondrichthyes

Striatolama sp.

Specimens: RAM 7256, RAM 7255, RAM 7254 (Figure 13); all from RAM locality V200307.

Description: The morphology of these teeth is very similar to those of *Striatolamia*. RAM 7256 is an anterior tooth that is well preserved except only one lateral root is present. The intact crown is 13.7 mm in length and exhibits well-developed striations. A tiny lateral cusplet is present on the side of the crown where the lateral root is intact. This cusplet is minute so it is only visible under high magnification. RAM 7255 is an anterior tooth whose crown is intact but the root is almost entirely missing. The crown measures 15.3 mm in length and striations are evident, but they are not as well developed as those on RAM 7256. RAM 7254 is an anterior tooth with a broken crown measuring 8.4 mm in length. A small fragment of the root is present. The crown enamel is weathered and striations are present but faint.

Specimens of *Striatolamia striata* and *S. marcota* are widespread in the Upper Paleocene to Upper Eocene of Europe and the United States (Cappetta 1987). During its geological span *Striatolamia* tends to reduce its lateral

Figure 13: anterior teeth of *Striatolamia* sp.; a) RAM 7256; b) RAM 7255; c) RAM 7254; all views lingual.Figure 14: *Phenacodus* sp.cf. *P. vortmani*; RAM 7253 right p4; a) occlusal, b) labial, c) lingual views

cusplets (Capetta 1987), so the Goler specimens appear to be more aligned to *S. marcota* because of the highly reduced lateral cusplet evident on RAM 7256.

Mammalia

Phenacodus sp. cf. *P. vortmani*

Specimen: RAM 7253, right p4 (Figure 14), from RAM locality V200307.

RAM 7253 is slightly worn and has the typical bunodont cusp morphology of *Phenacodus*. The tooth is 10.2 mm in length and the width of the trigonid and talonid are 6.5 mm and 6.4 mm respectively. There are four trigonid cusps. The protoconid and metaconid are large and about the same height and size. Two smaller cusps occupy the position of the paraconid, with the labial cusp twice the size of the lingual cusp. The talonid is wide, with a large hypoconid and small entoconid. A hypoconulid is not present.

The morphological features of an isolated p4 of *Phenacodus*, like RAM 7253, are not fully diagnostic for determining species. However, size is helpful as RAM 7253 is too small to be *P. intermedius*, *P. magnus*, or *P. grangeri*, and too large to be *P. matthewi*. The length and width of RAM 7253 approach the maximums recorded for p4s of *P. bisonensis*; maximum length and width are 10.3 mm and 6.8 mm respectively (see tables A15–17 in Thewissen 1990). RAM 7253 is also similar to *P. vortmani*, as the size range of p4s of this species vary from 7.8–10.5 mm in length to 5.1–7.2 mm in width. The p4s of *P. bisonensis* differ from those of *P. vortmani* in having narrower talonid basins and the common presence of an entoconid, while entoconids are commonly absent on *P. vortmani* (Thewissen 1990). The p4 of RAM 7253 has a wide talonid and a small entoconid. The broad talonid and well developed basin on RAM 7253 align it with *P. vortmani*. RAM 7253 has a small entoconid and this feature is present (but not common) in *P. vortmani*. Based on the large size of RAM 7253 and minor differences in morphology in comparison to *P. bisonensis*, RAM 7253 is tentatively referred to *Phenacodus* sp. cf. *P. vortmani*.

Plantae

Pieces of fossil wood (some highly burrowed) were found at I200305 with the largest piece measuring 20 cm in length and 25 cm in diameter. Unidentifiable plant fragments and a partial leaf impression were recovered from a sandstone lens at I200306.

Discussion of Goler marine strata

The new Alf Museum marine localities extend the known area of the marine tongue that deposited a significant part of Member 4d. Also, fossils from members 3, 4a and 4b indicate that the Goler Basin was probably in close proximity to the Pacific Ocean throughout much of its history before the marine transgression evident in strata of Member 4d occurred. A single ray tooth (RAM 7180), suborder Batoidea, cf. *Hypolophodon* (Lofgren et al. 2002; Lofgren et al. 2008), was recovered from the Edentulous Jaw Site (Member 4b; #6 in Figure 1b). Batoids are characteristically nearshore marine fishes that are quite tolerant of brackish water and are often found in rivers (Bryant 1989). Also, the species represented by a section of turtle shell found at UCMP locality V81035 (Member 4a) is thought to be related to sea turtles (McKenna et al. 1987). Finally, a concretion from UCMP locality V5250 (Member 3) contains a bass-shaped osteichthyan fish whose fin morphology resembles beryciform fish (McKenna et al. 1987) whose extant representatives are marine (Zehren 1979). Cumulatively, these fossils suggest a marine influence through much of the history of deposition of the Goler Formation.

The sequence of marine and nonmarine sediments with marine invertebrates described by Cox (1987), McDougall (1987), and Squires et al (1988) (# 11 in Figure 1b) has a sequence of sedimentary rocks somewhat similar to the new marine section reported here (#12 in Figure 1b). The older known sequence consists of sandstones and conglomerates with marine mollusks, overlain by a marine siltstone unit with planktic and benthic foraminifera (indicating deposition at water depths of 50–150 meters), then capped by nonmarine sandstones and conglomerates that unconformably overlie the marine siltstone unit (Cox 1987; section 1 in figure 3, McDougall 1987).

In the new marine section, sandstones and conglomerates with marine mollusks, shark teeth, and a mammal tooth are overlain by a finer-grained sequence of sandstone and siltstone, and possibly capped by a sequence of nonmarine sandstone and conglomerate (Figure 2). The finer-grained sequence has an abundant marine mollusk fauna (at I200305) that resembles that from the underlying coarser sediments. Water depths of deposition for this

coarser sequence are interpreted to be 5–50 meters based on mollusks (McDougall 1987; Squires et al. 1988). Thus, the marine siltstone unit of Cox (1987) and McDougall (1987) that was deposited at greater depths appears to be missing from the new marine section. Sampling for planktic and benthic foraminifera at I200305 would provide additional data to access at what water depth it was deposited.

The lower conglomerate unit described by Cox (1987) and McDougall (1987) from their marine section (#11 in Figure 1b) was interpreted to represent a braided river delta (Cox 1987). The lower strata at the new marine section (0–15 meter level in figure 2) very likely represent this environment of deposition. The only apparent difference is paleontological (referring to specimens from I200306 and V200307). Lower strata at the new marine section have more abundant mollusks (many are whole valves or shells that show little abrasion), as well as shark teeth and a nonmarine mammal tooth. Also, whole oysters and vertebrate fossils are found in the conglomerates, indicating that little transport of the oysters occurred. Sites like V200307, with both nonmarine and marine vertebrates and marine invertebrate fossils, are rare and usually indicate that the strata represent a near-shore marine facies where nonmarine fossils survive transport just long enough to be deposited in a delta lobe. Thus, the lower strata at the new marine section may represent a more proximal delta facies than the lower conglomerate unit of Cox (1987).

A slightly deeper water facies is probably represented by the strata that contain I200305 (15–23 meter level in Figure 2) rich in well-preserved mollusks that apparently underwent some sorting but not great transport. Some of the mollusk species from I200305 occur lower in the section at I200306. The marine rocks at the new section are overlain by a presumed sequence of nonmarine sedimentary rocks (those above the 23 m level in Figure 2).

The marine unit in Member 4d was tentatively dated as late Paleocene (CP8) based on coccoliths (Reid and Cox 1989), late Paleocene–early Eocene based on mollusks (Squires et al. 1988), and early Eocene (P8 and ?P9) based on foraminiferal assemblages (McDougall 1987). The mammal tooth (RAM 7253) identified as *Phenacodus* sp. cf. *P. vortmani*, is the only one known from the marine unit of Member 4d. Based on mammalian faunas from the Rocky Mountain states, *P. vortmani* has a age span of Late Paleocene to Early Eocene (Thewissen 1990). If RAM 7253 is *P. vortmani*, then it supports a Late Paleocene–Early Eocene age for V200307, an interpretation that does not refine the earlier age estimates for the marine unit of Member 4d. The case is similar with the three *Striatolama*

teeth as this genus also has an age range of Late Paleocene to Early Eocene (Cappetta 1987).

Strata at the new marine section were sampled for paleomagnetic analysis and four samples with good results were all of reversed polarity (Albright et al. in press). Two of the samples stratigraphically bracket V200307, the other two bracket I200305. Because of the lack of precise age determination provided by marine and nonmarine fossils (Late Paleocene to Early Eocene), multiple chrons are possible correlatives, from Chron 25r to Chron 22r. Thus, the Late Paleocene to Early Eocene age estimate for the marine unit remains unchanged based on all available data. However, when the six specimens thought to represent new records of gastropods from the Alf Museum marine section are identified, they may provide a more precise age determination for Member 4d.

Acknowledgements

We thank B. Albright, R. Nydam, R. Baum, R. Lynas, C. Johnson, M. Torres, numerous Webb Schools students, and many others for assistance; J. Reed, R. Kaldenburg, D. Storm, and D. Lawler of the Bureau of Land Management for collecting permits; and the Mary Stuart Rogers Foundation and Goler Research Fund of the Raymond Alf Museum of Paleontology for financial support. Goler research was led and funded by the late Malcolm McKenna for nearly 60 years. We thank him for his unwavering interest and support.

Literature cited

- Albright, L.B., D.L. Lofgren, and M.C. McKenna. Magnetostratigraphy, mammalian biostratigraphy, and refined age of the Goler Formation (Paleocene), California. In press.
- Axelrod, D. I. 1949. Eocene and Oligocene formations in the western Great Basin. *Bulletin of the Geological Society of America* 60:1935-1936.
- Bryant, L. J. 1989. Non-dinosaurian lower vertebrates across the Cretaceous-Tertiary boundary in northeastern Montana. *University of California Publications in Geological Sciences* 134:1-107.
- Cappetta, H. 1987. *Chondrichthyes II, Mesozoic and Cenozoic Elasmobranchii*. Handbook of Paleichthyology, volume 3B, H-P. Schultze ed., Gustav Fischer Verlag, Stuttgart and New York, 193 pages.
- Cox, B. F. 1982. Stratigraphy, sedimentology, and structure of the Goler Formation (Paleocene), El Paso Mountains, California: Implications for Paleogene tectonism on the Garlock Fault Zone. Ph.D. dissertation, University of California, Riverside, 248 p.
- Cox, B. F. 1987. Stratigraphy, depositional environments, and paleotectonics of the Paleocene and Eocene Goler Formation, El Paso Mountains, California—geologic summary and road log; pp. 1-30 in Cox, B.F. (ed.), *Basin Analysis and Paleontology of the Paleocene and Eocene Goler Formation, El Paso Mountains, California*. SEPM (Society for Sedimentary Geology) Pacific Section Publication 57.
- Cox, B. F., and L. F. Edwards. 1984. Possible marginal-marine deposits in the Goler Formation (Paleocene), El Paso Mountains, California. *Eos* 65:1084.
- Cox, B. F., and M. F. Diggles. 1986. Geologic Map of the El Paso Mountains Wilderness Study Area, Kern County, California. United States Geological Survey Miscellaneous Field Studies Map, MF-1827, scale 1:24,000.
- Dibblee, T. W. 1952. Geology of the Saltdale Quadrangle, California. California Division of Mines Bulletin 160:7-43.
- Lofgren, D. L., M. C. McKenna, S. L. Walsh. 1999. New records of Torrejonian-Tiffanian mammals from the Paleocene-Eocene Goler Formation, California. *Journal of Vertebrate Paleontology* 19 (supplement to 3):60A.
- Lofgren, D. L., M. C. McKenna, S. L. Walsh, J. H. Hutchison, R. L. Nydam, and J. G. Honey. 2002. New records of Paleocene vertebrates from the Goler Formation of California. *Journal of Vertebrate Paleontology* 22 (supplement to 3):80A.
- Lofgren, D. L., J. G. Honey, M. C. McKenna, R. L. Zondervan, and E. E. Smith, 2008. Paleocene Primates from the Goler Formation of California: Science Series Number 41, Natural History Museum of Los Angeles County: 11-28.
- McDougall, K. 1987. Foraminiferal biostratigraphy and paleoecology of marine deposits, Goler Formation, California; pp. 43-67 in Cox, B.F. (ed.), *Basin Analysis and Paleontology of the Paleocene and Eocene Goler Formation, El Paso Mountains, California*. SEPM (Society for Sedimentary Geology) Pacific Section Publication 57.
- McKenna, M. C. 1955. Paleocene mammal, Goler Formation, Mojave Desert, California. *Bulletin of the American Association of Petroleum Geologists* 39:512-515.
- McKenna, M. C. 1960. A continental Paleocene vertebrate fauna from California. *American Museum Novitates* 2024:1-20.
- McKenna, M.C. and D.L. Lofgren. 2003. *Mimotricentes tedfordi*, a new Arctocyoniid from the Late Paleocene of California: *Bulletin of the American Museum of Natural History* 279:632-643.
- McKenna, M.C., J. H. Hutchison, and J. H. Hartman. 1987. Paleocene vertebrates and nonmarine mollusca from the Goler Formation, California; pp. 31-42 in Cox, B.F. (ed.), *Basin Analysis and Paleontology of the Paleocene and Eocene Goler Formation, El Paso Mountains, California*. SEPM (Society for Sedimentary Geology) Pacific Section Publication 57.
- McKenna, M. C., J. G. Honey, and D. L. Lofgren. 2008. *Goleroconus alfi*, a new small periptychid (Mammalia, Eparctocyona) from the late Paleocene of California: *Science Series Number 41, Natural History Museum of Los Angeles County*: 29-42.
- Reid S. A., and B. F. Cox. 1989. Early Eocene uplift of the southernmost San Joaquin Basin, California. *American Association of Petroleum Geologists Bulletin* 73:549-550.
- Squires, R. L., B. F. Cox, and C. L. Powell, II. 1988. Late Paleocene or Early Eocene mollusks from the uppermost part of the Goler Formation, California; pp. 183-187 in M. V. Filewicz and R. L. Squires (eds.), *Paleogene Stratigraphy, West Coast of North America*, Society of Economic Paleontologists and Mineralogists Pacific Section 58.
- Thewissen, J. G. M. 1990. Evolution of Paleocene and Eocene Phenacodontidae (Mammalia, Condylarthra). *University of Michigan Papers on Paleontology* 29:1-107.
- West, R. M. 1970. *Tetraclaenodon puercensis* (Mammalia: Phenacodontidae), Goler Formation, Paleocene of California, and distribution of the genus. *Journal of Paleontology* 44:851-857.
- Zehren, S. J. 1979. The comparative osteology and phylogeny of the Beryciformes (Pisces: Teleostei). *Evolutionary Monographs* 1:1-389.

A juvenile specimen of *Archaeohippus mourningi* (Perissodactyla: Equidae) from the Cajon Valley Formation (Middle Miocene) of California

Darrin C. Pagnac¹ and Robert E. Reynolds²

¹Museum of Geology, South Dakota School of Mines and Technology, Rapid City, SD 57701, darrin.pagnac@sdsmt.edu
²Redlands, CA 92373. rreynolds220@verizon.net

Abstract

A new specimen of *Archaeohippus mourningi* has been recovered from late Hemingfordian North American Land Mammal age sediments in Unit 3 of the Cajon Valley Formation (Tcv 3) in Cajon Pass, southern California. This specimen includes a partial skull containing juvenile dentition with one erupting adult molar. Associated is an isolated dp2 and fragments of scapula and pelvis. The associated juvenile dentition is described, as is the biostratigraphic range and geographic distribution of *Archaeohippus* in the Mojave Desert.

Background

The sedimentary series exposed in Cajon Valley, San Bernardino County, California, has played a notable role in southern California geology. In particular, the fluvial clastic sediments exposed unconformably atop the Cosy Dell Formation (Kcd; Morton and Miller, 2003; formerly San Francisquito) and Vaqueros (Oligocene–Miocene) Formation, and in part, equivalent to the middle Miocene Crowder Formation (Reynolds and others, 2008), have significantly added to knowledge of regional paleontology, geology, and tectonics. Formerly attributed to the Punchbowl Formation (Noble 1954a; 1954b; Dibblee 1967; Woodburne and Golz 1972), this sequence is now referred to the Cajon Valley Formation (Morton and Miller 2003). The Cajon Valley Formation is a 9000 foot sequence of coarse-grained epiclastic fluvial and alluvial sediments derived from source material from the Mojave Desert province to the north-northeast.

The geology of Cajon Valley was originally outlined in works by Noble (1954a; 1954b) who assigned the term “Punchbowl Formation” to the aforementioned sediments, correlating them with the Punchbowl Formation proper from the type area at Devil’s Punchbowl, 25 miles northwest of Cajon Pass near Valyermo. More detailed studies of stratigraphy were conducted by Tamura (1961) and Swinehart (1965). Dibblee (1967) discussed the stratigraphy and structure of Cajon Valley and again

makes reference to the “Punchbowl Formation.” The correlation of the Cajon Valley “Punchbowl Formation” with that in the type area at Devil’s Punchbowl was used as a measure of slip on the San Andreas Fault. Displacement of these correlative beds suggested 25 to 40 miles of local right lateral slip (Dibblee, 1967).

However, questions arose as to the valid correlation of these alluvial deposits on either side of the San Andreas Fault. Tedford and Downs (1965) suggested that the Punchbowl Formation proper, in the type area at Devil’s Punchbowl, was considerably younger than its “equivalent” beds in Cajon Pass. Detailed study of the lithostratigraphy and paleontology by Woodburne and Golz (1972) confirmed Tedford and Down’s hypothesis. Although the sediments in the “Punchbowl Formation” on either side of the San Andreas are similar lithologically, their mammalian faunal assemblages are temporally disjunct. The assemblage from the Punchbowl Formation proper is early Clarendonian (Cl 1–2) in age; the Cajon Valley beds in Cajon Valley are late Hemingfordian to early Barstovian (He2–Ba1). Woodburne and Golz also suggest a source area for sands in the Cajon Valley beds from exposures to the north-northeast near Victorville, separate and distinct from the source area at the Punchbowl type area.

Morton and Miller (2003) produced a preliminary geologic map of the San Bernardino 30' x 60' quadrangle in which they assign the Tertiary sequence in Cajon Pass to the Cajon Valley Formation. They divide the Cajon Valley Formation into six sub-units (Tcv1–6) based on those originally described (Woodburne and Golz, 1972; Woodburne, 1991).

Biostratigraphy

Woodburne and Golz (1972) presented the most complete summary of the mammalian fauna of the Cajon Valley Formation. Although their work focused on stratigraphy, they mention several key mammalian taxa important in assigning faunas to a North American Land Mammal Age (NALMA). The lower exposures of the Cajon Valley Formation were assigned a late Hemingfordian age based on the co-occurrence of the equid *Parapliohippus carrizoensis* (= *Merychippus tehachapiensis*) and the dromomerycid *Bouromeryx milleri*. Overlying exposures were assigned an early Barstovian designation based on the occurrence of the equid *Scaphohippus* (= *Merychippus*) *intermontanus*, the oreodont *Brachycrus buwaldi*, and the camelid *Aepycamelus* cf. *A. alexandrae*.

Reynolds (1991; Reynolds and others, 2008; Wagner and Reynolds, 1983) provides summary and faunal lists from several localities from Cajon Valley which were compiled from over a decade of paleontological monitoring and research throughout Cajon Valley. Of particular note are late Hemingfordian (He2) and early Barstovian (Ba1) faunal assemblages, which include an occurrence of *Archaeohippus mourningi*.

Cajon Valley occurrences of *Archaeohippus*

Archaeohippus remains are quite common in the Cajon Valley Beds and have been reported by a number of authors. The earliest reports come from Bode (1933) who described USNM 12244, an incomplete *A. mourningi* dentary with right i3–m2 and left p1–m2 collected by Dr. C. L. Gazin from the Miocene beds of Cajon Pass southeast of Alray.

Woodburne and Golz (1972) list five specimens of *Archaeohippus mourningi* from their “Punchbowl Formation,” specifically in Cajon Valley, at various stratigraphic levels. Most of these specimens consist of partial dentaries and associated dentition or postcrania. Woodburne and Golz utilize the occurrence of *A. mourningi* to assign an early Barstovian age to the younger assemblages from Cajon Valley, thereby correlating with the Green Hills Division and Second Division of the Barstow Formation (*Plithocyon barstowensis*/*Aelurodon asthenostylus*, *A. asthenostylus*/*Ramoceros brevicornis*, *R. brevicornis*/*Megahippus mckennai* interval zones of Pagnac, 2009).

Other *Archaeohippus* specimens from Cajon Pass are from the “Dip Slope” locality (Wagner and Reynolds, 1983; SBCM collections), He2 Unit Tcv3, of similar age to the new specimen described herein. The new locality, however, is a mile northwest of the Dip Slope locality and a mile southwest of the original discovery at Alray (Bode, 1933). Reynolds also comments on occurrences of *Archaeohippus* from the adjacent Crowder Formation in the He2-age Wye Local Fauna (Reynolds, 1991; Reynolds and others; 2008), and the Ba2-age Sulfur Spring Local Fauna in Unit 2 of the Crowder Formation. Reynolds lists *Archaeohippus* from Ba1-age Tcv5 sediments at Tower 19, 1482 feet above the basal contact with Unit Tcv3, north-northeast of Cosy Dell in Cajon Valley. The University of California Riverside vertebrate paleontology collection, recently relocated to UCMP, contains numerous specimens of *Archaeohippus* from Cajon Valley. The bulk of these specimens were collected from RV 7515, from the north side of a stream just north of Cajon Junction. The Cajon Valley and Crowder formations both contain numerous occurrences of *Archaeohippus* that span late Hemingfordian to early Barstovian time. This abundance lies in stark contrast to the notable paucity of *Archaeohippus* from the age-correlative Barstow Formation.

Abbreviations:

DPOF—dorsal preorbital fossa

MSTHT—mesostyle crown height

RV—University of California, Riverside, vertebrate locality

SBCM—San Bernardino County Museum, Redlands, CA

UCR—University of California, Riverside. Vertebrate holdings now housed at UCMP

UCMP—University of California Museum of Paleontology, Berkeley, CA

Systematic paleontology

CLASS: Mammalia Linnaeus 1758

ORDER: Perissodactyla Owen 1848

FAMILY: Equidae Gray 1821

GENUS: *Archaeohippus* Gidley 1906

Archaeohippus mourningi Merriam 1913

Description

The specimen (No. Tcv3 - 31; FN. 07JB 11-16-01) consists of a partial skull and isolated dp2. The skull has been diagenetically crushed. Due to lateral flattening the palate is obscured and the upper tooth rows are closely oppressed and partially obscured by remnant matrix. The left side of the skull is much more complete than the

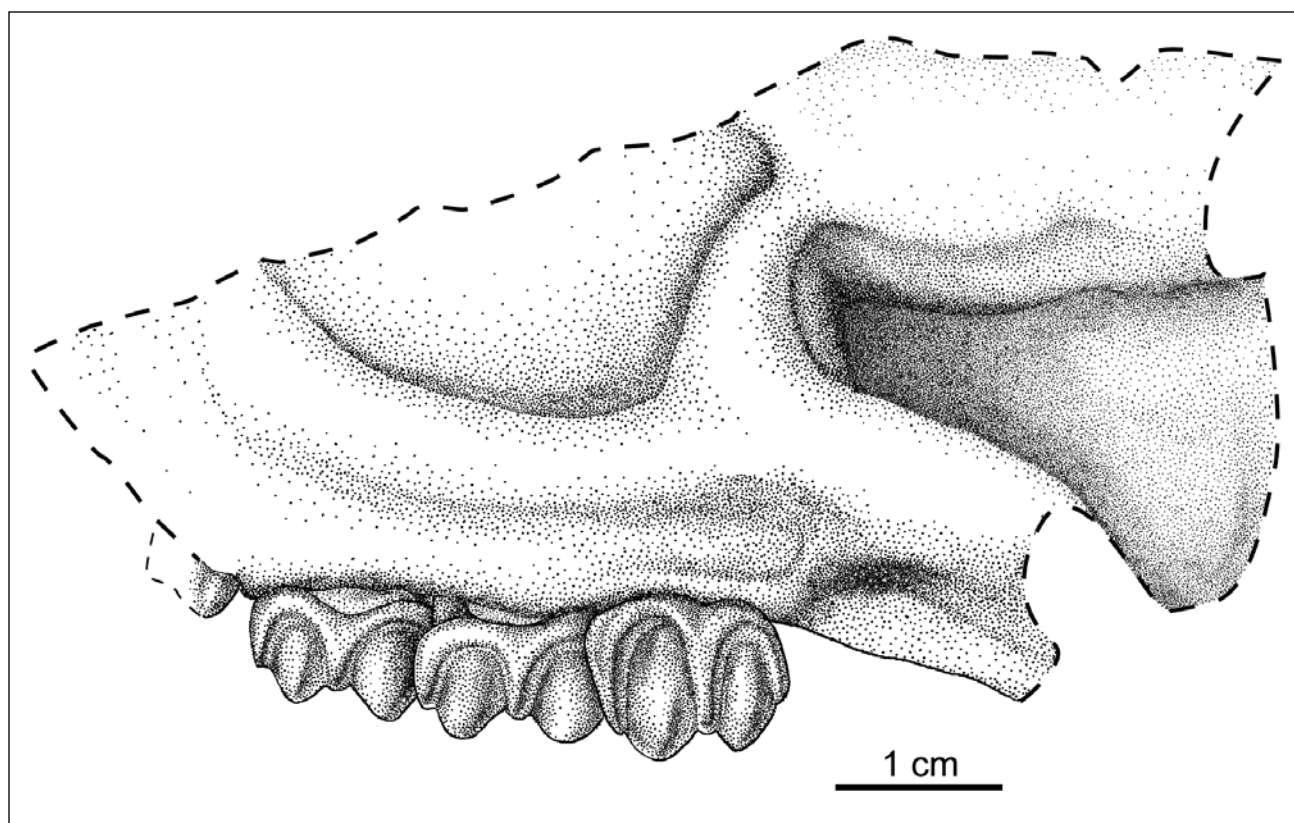


Figure 1: Left lateral view of the skull No. Tcv3 - 31; FN. 07JB 11-16-01 (*Archaeohippus mourningi*). The figured portion of the skull represents portions that are complete and unobscured by matrix.

Table I: Skull measurements for Tcv3 - 31; FN. 07JB 11-16-01 (*Archaeohippus mourningi*)

	mm
A-P skull length	155.0
Skull width	38.5
Skull height	53.5
Orbit length	42.9
Orbit height	24.0
Orbit depth	15.3
DPOF length	33.5
DPOF height	24.0
DPOF depth	7.2
UTRL	53.8

right, but much more of the right dentition is visible. On the left side, the anterior skull is missing from the anterior portion of the dorsal pre-orbital fossa (DPOF) and DP1 and anterior. The left braincase is missing as is the posterior portion of the left orbit, which is badly crushed. The left tooth row consists of DP2–4 and M1. The right side consists of a complete deciduous tooth row (DP1–4) and the lateral portion of the palate.

The left side of the skull (Fig. 1) is 155 mm long, measured from the anterior portion of the DPOF to the posterior margin of the orbit (Table I). The DPOF is 33.5 mm long, 24.0 mm high and 7.2 mm deep. Due to diagenetic alteration, it is difficult to discern any details of morphology within the DPOF. The posterior and ventral borders are quite sharp and defined. The DPOF does not exhibit noticeable nasomaxillary, nasolacrimal, or malar

Table II: Upper tooth measurements for Tcv3 - 31; FN. 07JB 11-16-01 (*Archaeohippus mourningi*).

	Right P1	Right DP2	Right DP3	Right DP4	Left DP3	Left DP4	Left M1
Length	10.7	15.0	12.8	13.9	12.0	13.8	13.3
Width	7.9	12.5	13.7	11.9	6.3	6.9	13.2
MSTHT	-	5.5	4.5	5.5	-	-	9.7
Protocone length	-	3.7	3.0	5.0	-	-	2.0
Protocone width	-	4.5	2.5	3.0	-	-	2.3

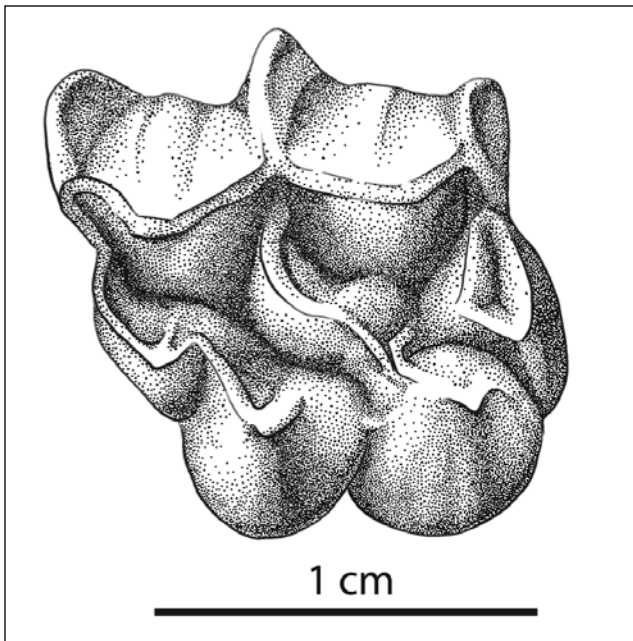


Figure 2: Occlusal view of the left M1 of No. Tcv3 - 31; FN. 07JB 11-16-01 (*Archaeohippus mourningi*). Anterior to left, labial up.

subdivisions, because their presence and expression may be obscured through diagenesis. The orbit is crushed and the postorbital bar, although present in *Archaeohippus*, was not preserved in this specimen.

The left tooth row is partially obscured by matrix, limiting the visible elements to the DP3–DP4 and M1. A miniscule portion of the posterior DP2 is visible, but detailed description is not possible. Tooth measurements can be found in Table II. The DP3 is almost completely encased in matrix, but the visible lateral surface shows prominent, well-developed labial styles. All exposed cheek teeth lack a lingual cingulum. The DP3 is 12.0 mm long with a mesostyle crown height (MSTHT) of 6.3

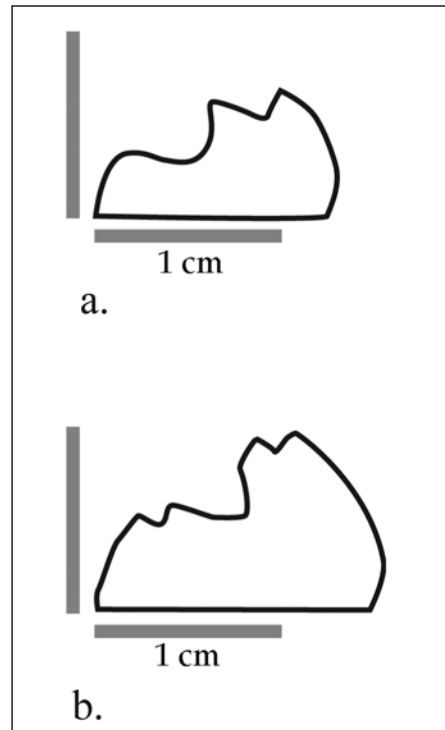


Figure 3: Medial/lateral cross section of upper cheek teeth from Tcv3 - 31; FN. 07JB 11-16-01 (*Archaeohippus mourningi*): a) left DP4, b) left M1.

mm. The DP4 is 13.8 mm long with an MSTHT of 6.9 mm. The anterolingual quadrant of this tooth is covered in matrix, but enough of the occlusal area is present to reveal a worn surface lacking a crochet between the protoloph and metaloph. The completely exposed M1 is 13.3 mm long and 13.2 mm wide, with a MSTHT of 9.7 mm, considerably taller than the deciduous dentition. The occlusal surface is unworn with a sharp protoloph and metaloph, lacks a crochet, and exhibits a rounded protocone and hypocone (Table II). As in the deciduous

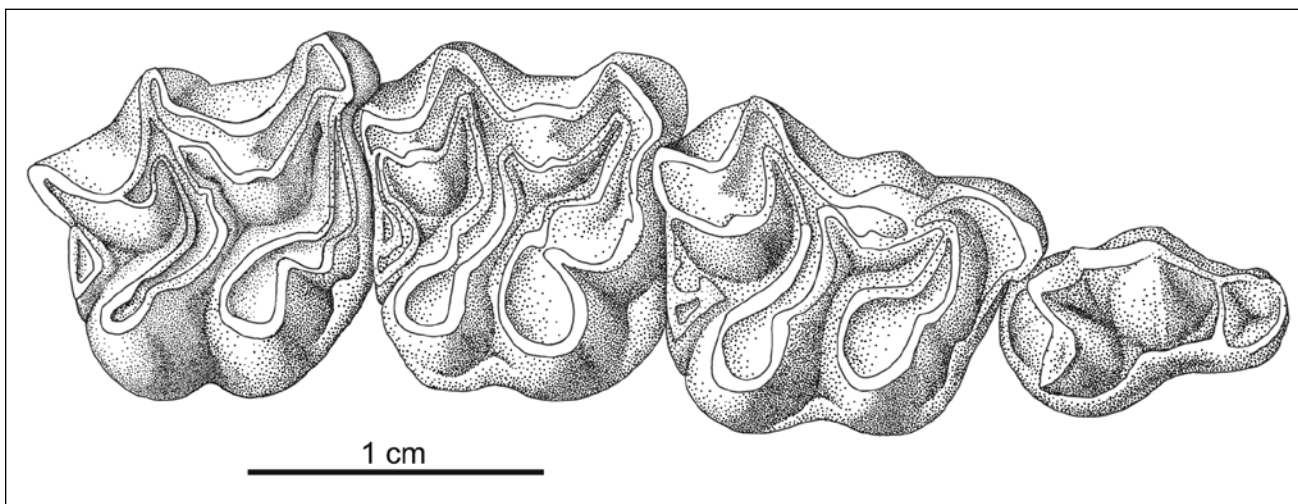


Figure 4: Occlusal view of the left DP2-DP4 of No. Tcv3 - 31; FN. 07JB 11-16-01 (*Archaeohippus mourningi*). Anterior to right, labial up.

Table III: Lower tooth measurements for Tcv3 - 31; FN. 07JB 11-16-01 (*Archaeohippus mourningi*).

	mm
Left dp2 Length	17.0
Left dp2 Width	6.7
Left dp2 Height	4.1

premolars, the M1 (Fig. 2) exhibits an extremely robust parastyle, mesostyle, and metastyle, thereby confirming this specimen's referral to *Archaeohippus mourningi*. Figure 3 illustrates a medial/lateral cross section of the DP4 and M1. Note the difference in crown height exhibited between the two teeth.

The right cheek tooth row (Fig. 4) is complete and un-obscured, containing DP1–DP4. As is typical of *Archaeohippus*, the teeth contain no cement. The DP1 is typical in size and shape, 10.7 mm long and 7.9 mm wide. The DP2 is 15.0 mm long and 12.5 mm wide, with an MSTHT of 5.5 mm. The occlusal surface shows considerable wear with dentine exposed on the protocone, paracone, metacone, hypocone, protoloph, and metaloph. The DP2 lacks a crochet. The protocone and hypocone are rounded (Table II). As on the left dentition, the labial styles on all the right cheek teeth are prominent and well-developed. The DP3 is 12.8 mm long and 13.7 mm wide, with a MSTHT of 4.5 mm. Again, the occlusal surface is well-worn with dentine visible on the all cones and lophs. The protocone and hypocone are rounded and a crochet is absent. The DP4 is 13.9 mm long and 11.9 mm wide, with a MSTHT of 5.5 mm. As with all of the other deciduous teeth from this specimen, the occlusal surface is well worn, there is no crochet, and the protocone and hypocone are rounded.

The isolated left dp2 (Fig. 5) is 17.0 mm long, 6.7 mm wide, and has a crown height of 4.1 mm (Table III). It is

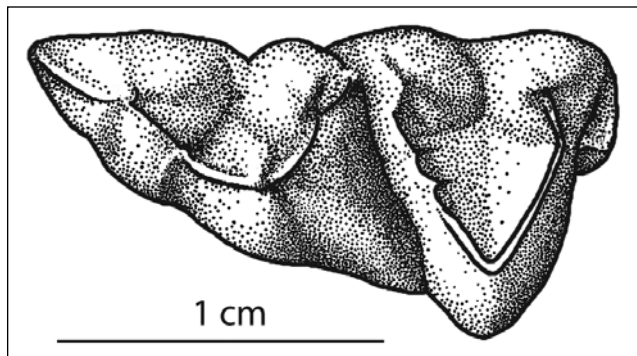


Figure 5: Occlusal view of dp2 of No. Tcv3 - 31; FN. 07JB 11-16-01 (*Archaeohippus mourningi*). Anterior to left, labial down.

typical of the species with no cement; prominent cusps and sharp lophs indicate little wear. The metaconid and metastylid are distinct and separate, and the hypoconulid is a low, flat shelf immediately posterior to the entoconid.

Remarks

Juvenile specimens of *Archaeohippus mourningi* dentition are notably rare. The type specimen, UCMP 19480 (= *Para-hippus mourningi* Merriam, 1913), consists of a partial left maxilla with DP3–4 and partially erupted M1. The only other juvenile tooth known is UCMP 32219, a left DP2 from the previously mentioned UCR locality RV 7515 north of Cajon Junction. The new specimen not only has a much more complete set of teeth and a partial cranium available, but it also represents the first record of deciduous lower dentition for the species.

Discussion

The specimen of *Archaeohippus* described above is typical of the material derived from the Cajon Valley and Crowder formations. It is more complete, albeit diagenetically altered, than the majority of isolated teeth or postcranial elements obtained from these sediments. The compilation of *Archaeohippus* material from Cajon Valley provides a sizable sample and contains some of the only known deciduous dentition of the genus.

Archaeohippus remains are quite abundant within the Cajon Valley and Crowder formations within Cajon Pass. At least a dozen specimens have been obtained from both formations within Cajon Valley (Bode, 1933; Woodburne and Golz, 1972; Reynolds, 1991, Reynolds and others, 2008). *Archaeohippus* is a standard component of most Miocene assemblages obtained from Cajon Valley. In contrast, very few specimens (possibly five) have been positively identified from the correlative Barstow Formation to the east (Merriam, 1913; Osborn, 1918; Pagnac, 2005; 2009). The difference in abundance of *Archaeohippus* fossils between the Cajon Valley Formation and the Barstow Formation may imply distinct depositional conditions indicating an environmental preference for the genus.

The Barstow basin drained internally in early Miocene time, slowly filling with 1000 m of sediment over a period of 4 m.y. (Woodburne, 1991b). These sediments, primarily silts and silty sands, have pulses of fanglomerate along basin margins. In contrast, the Cajon Valley Formation (Woodburne, 1991a) and the Crowder Formation (Reynolds and others, 2008) consist of pulses of arkosic alluvium derived from the Victorville Highland to the northeast (Woodburne and Golz, 1972) and deposited on southwest-facing slopes. These high energy pulses of coarse clastic material punctuate many periods of weathering on stable surfaces that developed red-brown

paleosols (fossil soil horizons) and wetland facies. Despite periods of surface stability with little or no deposition, the Cajon Valley Formation developed a thickness of 2,735 m over a period of 4 m.y. (Woodburne, 1991a). Thus, even with multiple periods of stability indicated by paleosols, the depositional environment of the Cajon Valley Formation had much higher transport energy than that in the Barstow basin, where paleosols are absent.

The weathering surfaces developed stable vegetation for periods of several thousand years. Margins of these “red-beds” were bounded by channels, (cut and fill of Woodburne and Golz, 1972) that may have contained riparian vegetation. These paleosols are important depositional environments that produce most of the invertebrate and vertebrate fossils, including *Archaeohippus* in the Crowder and Cajon Valley formations (SBCM Collections).

The middle member of the Barstow Formation (from which all specimens of *Archaeohippus* have been derived) contains fine-grained sandstone and mudstone beds with algal limestones deposited under relatively low energy fluvial or lacustrine conditions (Steinen, 1966; Woodburne et al., 1990). The depositional basin’s position on the east side of a Victorville highland (Woodburne and Golz, 1972) may have resulted in a nominal rain shadow effect.

The Barstow and Cajon Valley formations were deposited at about the same time, but under considerably different conditions. Deposition of the fluvio-lacustrine Barstow Formation was apparently at a slow, continuous rate, as indicated by the absence of paleosols. The presence of paleosols represents depositional gaps in the Cajon Valley and Crowder formations and appears to have provided vegetated habitats that favored *Archaeohippus*.

Plant information can provide insight into environmental conditions, but paleobotanical records from both formations are minimal at best. Alf (1970) described a small flora from the Barstow Formation including *Juniperus* (juniper) and *Toxodendron* (poison oak) wood impressions, *Quercus* (oak) leaves, and a palm frond. Plant macrofossils identified from a restricted lacustrine facies in the Cajon Valley Formation include *Celtis* (hackberry), *Platanus* (sycamore), and *Umbellularia* (California laurel) (Reynolds, unpublished data). Unfortunately, all of these taxa are too cosmopolitan in distribution for detailed environmental interpretations.

Faunal analysis yields little insight as well. The diverse fossil assemblages from both formations provide excellent correlative utility based on similarity. Unfortunately, this similarity prevents analysis of distinct faunal differences facilitating environmental comparison. If discrete environmental differences did exist between the two depositional basins, these differences would likely be shown

through microfaunal analyses. Although such studies are currently being conducted, the overall similarity of even the microfauna of the two formations prevents identification of specific environmental differences.

The issue of *Archaeohippus* environmental preference can likely be addressed through stable isotope analysis. Examination of carbon isotope signatures will identify differences in locally occurring flora as preserved in dietary signals. *Archaeohippus* remains could be sampled from both formations to detect disparate carbon isotopic signals. Comparative isotope data from abundant herbivorous taxa, particularly hypsodont equids, may reveal contrasting isotopic signatures between additional taxa from the two formations, thereby confirming potential environmental differences. Comparative studies of vertebrates from the Cajon Valley and Barstow formations provide a unique opportunity to investigate local variations in deposition, environment, and biota, provided these studies are kept within a properly detailed stratigraphic and temporal framework.

Acknowledgements

The authors acknowledge the constructive review provided by Michael Woodburne. Additional thanks to Katura Reynolds for preparing the line drawings of the specimen.

References cited

- Alf, R. M. 1970. A preliminary report on a Miocene flora from the Barstow Formation, Barstow, California. Southern California Academy of Sciences 69:183-188.
- Bode, F. C. 1933. Anchitherine horses from the *Merychippus* zone of the North Coalinga District, California. Contributions to Paleontology: Carnegie Institute of Washington 440:43-58.
- Dibblee, T. W. 1967. Areal geology of the western Mojave Desert, California. US Geological Survey Professional Paper 522:1-153.
- Dibblee, T. W. J. 1968. Geology of the Fremont Peak and Opal Mountain quadrangles, California. California Division of Mines and Geology Bulletin 188.
- Merriam, J. C. 1913. New Anchitherine horses from the Tertiary of the Great Basin area. University of California Publications Bulletin of the Department of Geological Sciences 7:419-434.
- Morton, D. M., and F. K. Miller. 2003. Preliminary Map of the San Bernardino 30' x 60' quadrangle, California. US Geological Survey Open File Report 190.
- Noble, L. F. 1954a. Geology of the Valermo quadrangle and vicinity, California. US Geological Survey Quadrangle Map GQ-50.
- Noble, L. F. 1954b. The San Andreas fault zone from Soledad Pass to Cajon Pass, California. California Division of Mines and Geology Bulletin 17:37-48.
- Osborn, H. F. 1918. Equidae of the Oligocene, Miocene, and Pliocene of North America; iconographic type revision. Memoirs of the American Museum of Natural History 2:1-330.
- Pagnac, D. C. 2005. A systematic review of the mammalian megafauna of the Middle Miocene Barstow Formation, Mojave Desert, California. Ph.D. thesis/dissertation, Department of Geology, University of California, Riverside, Riverside, CA, 384 pp.

- Pagnac, D. C. 2009. Large mammal biostratigraphy of the Barstow Formation (Middle Miocene), California. *Paleobios* 29:48-59.
- Reynolds, R. E. 1991. Biostratigraphic relationships of Tertiary small mammals from Cajon Valley, San Bernardino County, California. *San Bernardino County Museum Association Quarterly* 38:78-80.
- Reynolds, R. E., R. L. Reynolds and E. H. Lindsay. 2008. Biostratigraphy of the Miocene Crowder Formation, Cajon Pass, southwestern Mojave Desert, pp. 237-254 in X. Wang and L. G. Barnes (eds.), *Geologic and vertebrate paleontology of western and southern North America. Science Series 41*, Natural History Museum of Los Angeles County.
- Steinen, R. P. 1966. Stratigraphy of the middle and upper Miocene Bartow Formation, San Bernardino County, California. Unpublished Master's Thesis thesis/dissertation, University of California, Riverside, 150 pp.
- Swinehart, J. R. 1965. Geology of a portion of the lower Cajon valley, San Bernardino County, California. Unpublished Bachelor's thesis thesis/dissertation, Department of Geologic Sciences, University of California, Riverside.
- Tamura, A. Y. 1961. Stratigraphy of a portion of the upper Cajon Valley, California. Unpublished Bachelor's thesis thesis/dissertation, Department of Geological Sciences, University of California, Riverside.
- Tedford, R. H., and T. Downs. 1965. Age of the Punchbowl Formation, Los Angeles and San Bernardino County, California. *Geological Society of America Special Paper* 87:234.
- Wagner, H.M. and R.E. Reynolds. 1983. *Leptarctus ancipidens* (Carnivora: Mustelidae) from the Punchbowl formation, Cajon Pass, California: *Bulletin of the Southern California Academy of Sciences*.
- Woodburne, M. O. 1991a. The Cajon Valley; pp. 49-51 in M. O. Woodburne, R. E. Reynolds, and D. P. Whistler (eds.), *Inland Southern California: the last 70 million years*. San Bernardino Museum Association Quarterly 38(3-4):49-51.
- Woodburne, M. O. 1991b. The Mojave Desert Province; pp. 60-77 in M. O. Woodburne, R. E. Reynolds, and D. P. Whistler (eds.), *Inland Southern California: The last 70 million years*. San Bernardino Museum Association Quarterly 38(3-4):60-77.
- Woodburne, M. O., and D. J. Golz. 1972. Stratigraphy of the Punchbowl Formation, Cajon Valley, California. *University of California Publications in Geological Sciences* 92.
- Woodburne, M. O., R. H. Tedford, and C. C. Swisher. 1990. Lithostratigraphy, biostratigraphy, and geochronology of the Barstow Formation, Mojave Desert, southern California. *Geological Society of America Bulletin* 102:459-477.

Early Pleistocene range extension for *Microtus mexicanus* in the southwestern Mojave Desert

Robert E. Reynolds

Redlands, California 92373, rreynolds220@verizon.net

Abstract

The presence of the small microtine rodent, *Microtus mexicanus*, in ancestral Mojave River strata in northwestern Victorville, California, represents a temporal range extension for the species back to the Irvingtonian North American Land Mammal (early Pleistocene Epoch) and a geographic range extension into southern California. The presence of this taxon helps constrain the age of the sediments in which it was found to less than 1.4 Ma. It also assists with the interpretation of both the geographic range of *Microtus* species at that time, and the early Pleistocene habitat between Victorville and the Transverse Ranges of southern California.

Introduction

A lower first molar (M/1) (Natural History Museum of Los Angeles County Vertebrate Paleontology Section [LACM] 154595) representing a small-bodied species of *Microtus* was recovered from 10–11 feet below the surface in a trench excavated during construction of the Dr Pepper Snapple Group, Inc., West Coast beverage production and distribution facility at the Southern California Logistics Centre (formerly George Air Force Base) in northwestern Victorville, San Bernardino County, California (Lander, 2010). The locality (LACM 7786) was in a sequence of strata of late Pliocene to early Pleistocene age in the Adelanto 7.5-minute Quadrangle (unit Qo of Bortugno and Spittler, 1986; unit Qvof of Morton and Miller, 2003; unit Qoam of Hernandez and others, 2008). Strata in the vicinity of the fossil locality appear to represent cyclic wetlands that interfinger with the developing fluvial sediments of the Mojave River and the overlying sequence of arkosic fanglomerates of the Victorville fanglomerates that are derived from the San Gabriel Mountains near Victorville basin Cajon Pass (Meisling and Weldon, 1989). The transition from the sandy silts of the Victorville basin to the fluvial sequence of the Mojave River has been summarized previously (Reynolds and Cox, 1999; Reynolds and Miller, this volume), as has the geologic age of the Victorville Fan sequence (Cox and Tinsley, 1999; Izbecki, 1999).

Geologic setting

The typical stratigraphy seen in the vicinity Southern California Logistics Centre consists of three stratigraphically superposed units (lower, middle, upper). The lower unit consists of a pebble-cobble arkosic sand deposited by a southwardly flowing axial stream system. Clasts of Jurassic and early Miocene volcanic rocks are derived from the Mojave Desert, not from the Transverse Ranges, and include lithologies found in the Kramer Hills and near Barstow. Borehole magnetostratigraphy suggests that the top of the northerly derived, lower unit is about 2.5 Ma old (Cox and others, 1999; Reynolds and Cox, 1999). Such an age would be within the 4.1–1.5 Ma age range of the Phelan Peak Formation on the northern slope of the San Gabriel Mountains west of Cajon Pass. The depositional history of the Phelan Peak Formation records a reversal in its drainage direction from southward to northward (Meisling and Weldon, 1989), corresponding approximately to the age of the upper limits of the northerly derived, lower unit in Victorville.

The middle unit consists of lacustrine sandy silt and is seen in outcrops and in bore holes over a wide area. To the south, magnetostratigraphic data from bore holes indicate that the middle unit is late Pliocene to early Pleistocene or from 2.5 to ~2.0 Ma in age (Cox and others, 1999, 2003).

The upper unit is an upward-coarsening fluvial sequence of silt, sand, and gravel. Clast lithologies indicate

a source for these ancestral Mojave River gravels in the northwestern San Bernardino Mountains. A suggested depositional history is that stream flow from the southeast deposited the lacustrine middle unit in the Victorville basin north of the San Bernardino Mountains. As the ancestral Mojave became a well-established drainage and the mountains continued to rise, sediments overwhelmed the Victorville basin, the river proceeding to flow northward and then eastward to terminate at Lake Manix (Reynolds and Cox, 1999; Sibbett, 1999).

Stratigraphic occurrence

The fossil specimen was recovered from 10–11 feet below the surface in a greenish-gray, clayey to silty, fine-grained sandstone containing calichified root casts (Figure 1; Lander, 2010). The fossil-bearing horizon is a paleosol in a 7-foot-thick deposit of greenish-gray, fine- to coarse-grained, silty sandstone with caliche bands (Figure 1; Lander, 2010). This fine-grained sequence occurs stratigraphically below a 10-foot-deep channel of tan pebbly sand to poorly indurated sandstone with

east-southeasterly trending, prograded foreset bedding (Figure 1; Lander, 2010). The sand, in turn, is capped by a white caliche zone with very well indurated sandstone that fills rodent burrows penetrating underlying units (Figure 1; Lander, 2010), suggesting that soil carbonate developed over a long period of time. The entire sedimentary sequence is unconformably overlain by deeply weathered, bioturbated reddish-brown sediments (Figure 1; Lander, 2010) that might be a distal lateral equivalent of the Victorville fanglomerate (Meisling and Weldon, 1989).

Analysis

The presence of a *Microtus* M/1 that is 25% shorter in length than the mean of that parameter for *M. californicus* prompted a study of a number of Holocene species for comparison. The small fossil tooth was identified based on its “chevron morphology,” which is most like that of *Microtus* sp., and because it contains a posterior loop preceded by three closed basic triangles. The complex anteroconid has labial triangles confluent with buccal triangles (dental terminology after Repenning 1992:fig. 10). The common extant species in the Victorville area, *M. californicus*, was ruled out as a possible candidate because of the smaller size of the Victorville tooth. The Victorville M/1 was

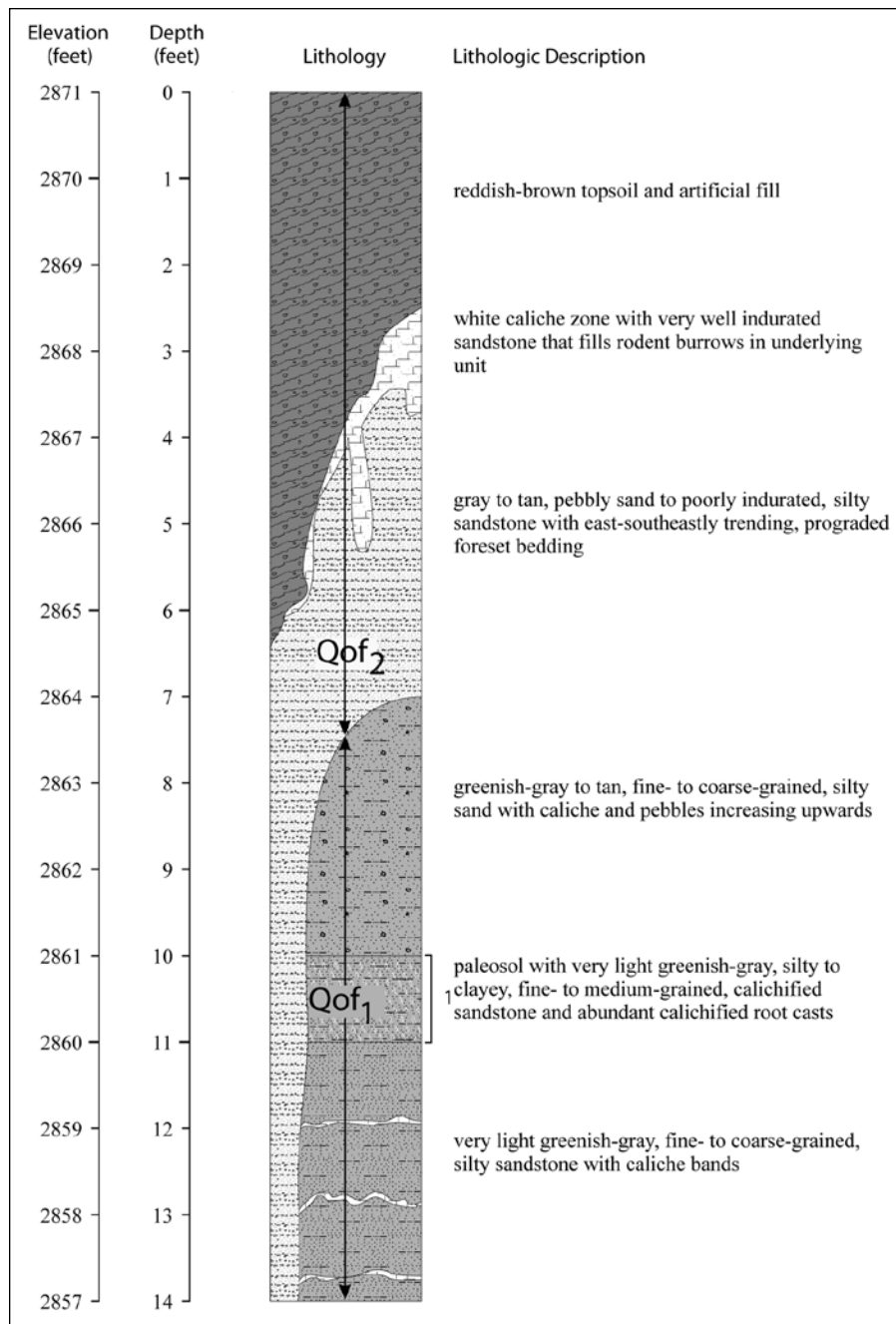


Figure 1. Stratigraphic columnar section, Southern California Logistics Centre, Victorville, San Bernardino County, California, showing stratigraphic level of recorded fossil locality. Qof₂: older Pleistocene Mojave River sands. Qof₁: older Pleistocene Mojave River silty sandstone.

Table 1. Measurements (in mm) and morphology of m/1 for selected species of *Microtus* arranged in order of increasing M/1 length, and for *Microtus* specimen from Victorville (LACM 154595).

<i>Microtus</i> species	Mean wength	Observed range	Anterior width	Posterior width	Curved/straight
<i>M. canicaudus</i>	2.50	2.5			
<i>M. oeconomus</i>	2.55	2.5-2.6	0.80	1.20	straight
<i>M. chrotorrhinus</i>	2.60	2.6			straight
<i>M. ochrogaster</i>	2.60	2.6	0.95	1.00	curved
<i>M. oregoni</i>	2.60	2.6			curved
<i>M. mexicanus</i>	2.70	2.6-2.8	0.70	0.90	straight
<i>M. miurus</i>	2.80	2.8			curved
<i>M. pinetorum</i>	2.90	2.8-3.0	0.90	1.00	straight
<i>M. montanus</i>	2.95	2.9-3.0			straight
<i>M. longicaudus</i>	3.15	3.1-3.2			
<i>M. californicus</i>	3.7	3.7	0.90	1.30	curved
<i>M. townsendii</i>	3.45	3.4-3.5			
<i>M. pennsylvanicus</i>	3.15	3.0-3.3			
<i>M. richardsoni</i>	4.00	4.0			
LACM 154595	2.80		0.70	0.90	straight

compared with the lower molars of thirteen other *Microtus* species in the comparative collection of the San Diego Natural History Museum (SDNHM). Comparisons considered the size (length and width) and shape of a tooth (straight versus curved or convex lingually), as well as the morphologies of the cusps and re-entrant angles forming the anteroconid. The species of *Microtus* studied and their measurements are listed in Table 1. The specimens were divided into two size groupings (small versus large), based on their M/1 lengths. Larger specimens were not considered further. For smaller specimens (i.e., M/1 lengths <3.0 mm), tooth morphologies fell into two groups, in which the teeth were either anteroposteriorly straight or lingually curved (convex).

The *Microtus* M/1 from Victorville (LACM 154595), with an anteroposterior length of only 2.8 mm and a straight anteroposterior axis, falls into a category comprising individuals with smaller teeth having straight anteroposterior axes. Such a tooth fits the size and shape of M/1s of *M. mexicanus*. Additionally, the anterior cap of the anteroconid complex has a very shallow lingual reentrant 4 opposing a deep buccal reentrant 5, which is comparable to that in *M. mexicanus*. Only *M. chrotorrhinus* has a lingual reentrant 4 that is almost flat and clearly shallower than in *M. mexicanus*. Each of the other small species of *Microtus* considered herein has a relatively deep lingual reentrant 4 opposing the deep buccal reentrant 5. Therefore, the size, comparatively straight anteroposterior axis of the tooth, and the morphology of the anterior cap of the anteroconid complex compare most favorably with those of *M. mexicanus*.

Biostratigraphic constraints

Fossil dire wolf, mammoth, and horse were recovered in coarse Mojave River sands near Village Drive and Amargosa Road in Victorville (Jefferson, 1989, 1991; Reynolds, 2006; Romero and Hilburn, 2006). Following

Bell and others (2004), the presence of dire wolf (*Canis dirus*) suggests that the fossil-bearing sediments in the uppermost ancestral Mojave River strata immediately below the Victorville Fan sequence are 0.2 Ma old and Rancho-labrean or late Pleistocene in age.

Fossil mammals from north of Air Expressway (formerly Air Base Road) include giant ground sloth (*Paramylodon*), short-faced bear (*Arctodus*), mammoth, horse, camel, llama, giant camel (*Titanotylopus*), and a meadow vole (*Microtus* sp.) (Scott and others, 1997). These taxa, particularly *Titanotylopus* and *Microtus*, suggest an early Pleistocene age of less than 1.4 Ma for ancestral Mojave River sediments.

The occurrence of the fossil cotton rat (*Sigmodon minor*) at a locality along Hesperia Road (Reynolds and Reynolds, 1994) suggests a Pliocene age for sediments south of where Interstate 15 crosses the Mojave River.

Associated stratigraphic and geographic locality data indicate that the fossil microtine specimen from Victorville was found 10-11 feet below the surface in a greenish-gray clayey to silty fine-grained sandstone containing calichified root casts (Figure 1; Lander, 2010). This fine grained sequence occurs stratigraphically below a 10-foot-deep channel of sandstone with east-southeasterly trending, prograded foreset bedding (Figure 1; Lander, 2010). The sand, in turn, is capped by white caliche with very well indurated sandstone that fills rodent burrows penetrating underlying units (Figure 1; Lander, 2010). The thickness and density of the caliche suggests the duration of carbonate development over tens of thousands of years, making the period of sand deposition even earlier. The entire sedimentary sequence is unconformably overlain by reddish-brown sediments (Figure 1; Lander, 2010) that might be a distal lateral equivalent of the Victorville conglomerate (Meisling and Weldon, 1989).

Summary

The fossil *Microtus* tooth from the Southern California Logistics Centre comes from ancestral Mojave River sediments that include both coarse fluvial sands and sandy silts. The temporal range of *Microtus* has been constrained by magnetostratigraphic studies in the Anza Borrego Desert and in the San Timoteo Formation in The Badlands of northwestern Riverside County (Repenning, 1992; Albright, 1999), suggesting that this genus did not appear in California until 1.4 Ma ago. Based on magnetostratigraphic constraints developed elsewhere and on stratigraphic analyses (Cox and Tinsley, 1999; Cox and others, 2003), the sand and sandy silt that produced the fossil tooth were deposited by the developing Mojave River system less than 1.4 Ma ago.

The modern Mexican vole (*Microtus mexicanus*) is not recorded in California (Ingles, 1965). It has a geographic range in the southwestern United States that extends from Arizona to southern Mexico (Nowak, 1991). Today, Mexican voles are found in dry habitats and grassy areas, such as those characterizing mountain meadows and yellow pine forests. However, the present distribution of the Mexican vole is not entirely the product of post-Pleistocene forest fragmentation and extinction. Its reduced modern range might be partly the result of the northward range expansion of many other southwestern mammalian species (Davis and Callahan, 1992).

Mexican voles are herbivorous. In the summer, they feed on herbs and green grasses, whereas they eat bark, bulbs, and roots during the winter. They forage all year round and neither stock up on food for the winter nor hibernate. The Mexican vole is endangered because of drought and because the tall grasses in which they live have been eliminated as a result of grazing by cattle and of human development (Encarta Encyclopedia, 2000). The need for specific types of vegetation and particular habitats suggests that their extirpation from the Mojave Desert occurred at the end of the last Pleistocene glaciation. A fossil occurrence of the Mexican vole from Victorville is a previously unrecognized geographic and temporal range extension for the species. Its presence in sediments of the ancestral Mojave River near Victorville provides insights to the early Pleistocene environment and habitats of the area.

Acknowledgements

I thank Dr. E. Bruce Lander of Paleo Environmental Associates, Inc. (PEAI), for reviewing this manuscript. The fossil *Microtus* specimen described herein was recovered by Mr. Patrick W. Riseley of PEA. He also constructed the stratigraphic column modified herein as Figure 1. I also thank Mr. Philip Unitt, Curator of Birds and Mammals at

the SDNHM, for access to the comparative specimens of *Microtus* used in this study. The mitigation program that was conducted by PEA and resulted in the recovery of the fossil specimen was supported by Benham DB Inc. dba Benham Constructors, LLC, on behalf Dr Pepper Snapple Group, Inc.

References cited

- Albright, L. B. 1999. Biostratigraphy and vertebrate paleontology of the San Timoteo Badlands, southern California. University of California Publications in Geological Sciences 144:1-121.
- Bell, C.H., E.L. Lundelius Jr., A.D. Barnosky, R.W. Graham, E.H. Lindsay, D.R. Ruez Jr., H.A. Semken Jr., S.D. Webb, and R.J. Zakrzewski. 2004. The Blancan, Irvingtonian and Rancholabrean Mammal Ages. Pp. 232-314 in M. O. Woodburne (ed.). Late Cretaceous and Cenozoic Mammals of North America. Columbia University Press, New York.
- Bortugno, E.J., and T.E. Spittler (compilers). 1986. Geologic map of the San Bernardino Quadrangle. California Division of Mines and Geology Regional Geologic Map Series Map 3A (Geology).
- Cox, B.F., and J.C. Tinsley III. 1999. The origin of the late Pliocene and Pleistocene Mojave River between Cajon Pass and Barstow, California. San Bernardino County Museum Association Quarterly 46(3):49-54.
- Cox, B.F., J.W. Hillhouse, and L.A. Owen. 2003. Pliocene and Pleistocene evolution of the Mojave River, and associated tectonic development of the Transverse Ranges and Mojave Desert, based on borehole stratigraphy studies and mapping of landforms and sediments near Victorville, California, in Paleoenvironments and paleohydrology of the Mojave and southern Great Basin Deserts, Y. Enzel, S.G. Wells, and N. Lancaster (eds.). Geological Society of America Special Paper 368:1-42.
- Davis, R., and J.R. Callahan. 1992. Post-Pleistocene dispersal in the Mexican vole (*Microtus mexicanus*): an example of an apparent trend in the distribution of southwestern mammals. Great Basin Naturalist 52(3):262-268.
- Encarta Encyclopedia. © 2000. Mammal Image Library. MSL Order Rodentia, accessed on 08/05/04 at <http://www.emporia.edu/biosci/msl/rodent.htm>. The Mammals of Texas, Online Edition. Mexican Vole, accessed on 08/05/04 at <http://www.nsrll.ttu.edu/tmot1/micrmexi.htm>
- Hernandez, J.L., H.J. Brown, and B.F. Cox. 2008. Geologic map of the Victorville 7.5' quadrangle, San Bernardino County, California: a digital database, version 1.0. California Geological Survey.
- Ingles, L.G. 1965. Mammals of the Pacific states, Stanford University Press, CA. 506 pp.
- Izbicki, J.A. 1999. Transition Probability/Markov Chain analysis of the subsurface geology of the Victorville Fan in the western part of the Mojave Desert, southern California. San Bernardino County Museum Quarterly 46(3):55-64.
- Jefferson, G.T. 1989. Late Pleistocene and earliest Holocene fossil localities and vertebrate taxa from the western Mojave Desert. San Bernardino County Museum Association Special Publication 89-1:27-40.
- Jefferson, G.T. 1991. A catalog of late Quaternary vertebrates from California: Part two, mammals. Natural History Museum of Los Angeles County Technical reports No. 7:1-129.
- Lander, E.B. 2010. Paleontologic resource impact mitigation program final technical report of results and finding prepared in support of Dr Pepper Snapple Group, Inc., West Coast beverage production and distribution facility, Southern California Logistics Centre, Vic-

- torville, San Bernardino County, California. Paleo Environmental Associates, Inc., project number 2009-5. Prepared for Benham DB Inc. dba Benham Constructors, LLC, on behalf Dr Pepper Snapple Group, Inc.
- Meisling, K.E., and R.J. Weldon II. 1989. Late Cenozoic tectonics of the northwestern San Bernardino Mountains, southern California. *Geological Society of America Bulletin* 101:106-128.
- Morton, D.M., and F.K. Miller. 2003. Preliminary geologic map of the San Bernardino 30' x 60' Quadrangle, California: United States Geological Survey Open-File Report 03-293, <http://geopubs.wr.usgs.gov/open-file/of03-293/>.
- Nowak, R.M. 1991. Walker's mammals of the world, Volume II, Johns Hopkins University Press, Baltimore, 1659 pp.
- Repenning, C. A. 1992. *Allophaiomys* and the age of the Olyor Suite, Krestovka sections, Yakutia. *United States Survey Bulletin* 2037:1-98.
- Reynolds, R.E. 2006. Paleontological resources monitoring program, KB Homes Saratoga Project (Tract 13908), Victorville, California. LSA Associates, Inc. Prepared for KB Homes. 16 pp.
- Reynolds, R. E. and B.F. Cox. 1999. Tracks along the Mojave: A field guide from Cajon Pass to the Manix Basin and Coyote Lake. *San Bernardino County Museum Association Quarterly* 46(3):1-26.
- Reynolds, R.E., and R.L. Reynolds. 1994. The Victorville Fan and the occurrence of *Sigmodon*. *San Bernardino County Museum Association Special Publication* 94-1:31-33.
- Reynolds, R.E. and D.M. Miller. 2010. Overboard in the Mojave: a field guide, *in* Overboard in the Mojave: 20 million years of lakes and wetlands, R.E. Reynolds and D.M. Miller (eds). California State University Fullerton, Desert Studies Consortium, this volume.
- Romero, D. and R.F. Hilburn. 2006. Come look: Mojave River mammoths, *in* The 2006 Desert Symposium Abstracts and Proceedings. California State University, Fullerton, Desert Studies Consortium, p. 78.
- Scott, E., K. Springer, and L. K. Murray. 1997. New records of Early Pleistocene vertebrates from the west-central Mojave Desert, San Bernardino County, CA. *Jour. Vert. Paleo.* Vol. 17(3), p. 75A.
- Sibbett, B. S. 1999. Pleistocene channels of the Mojave River near Victorville, California. *San Bernardino County Museum Association Quarterly* 46(3):p. 65-68.

The first tungsten-mining boom at Atolia, as recorded in the Randsburg Miner, 1904–1907

Alan Hensher, compiler

462 East Alexander Avenue, Merced, California 95340

Prologue

Although tungsten was discovered in the late 18th century, not until the 1890s were all of its properties known.

Tungsten has the ability to toughen steel and resist heat; in fact, it has the highest melting point of any metal. Even when red hot, cutting tools made of tungsten steel hold their shape. After the introduction of the alloy, in 1900, German factories began using tungsten-steel tools to cut shell casings and armor plating for warships.

Colorado emerged as the leading center of tungsten mining in the United States, but that soon changed. Under the headline “Scheelite,” the Redlands Citigraph of October 31, 1903, reported: “Randsburg Miner: A ledge or deposit of very high grade tungsten ore, what is known as scheelite, has been discovered in the Baltic mine. It assays 69.5 per cent., which is the highest grade. About seventy-five pounds of this ore has been sent to the state Mining Bureau. The superintendent, W. W. Wynn, informs us that he has a sale for all of the ore he can produce delivered in New York. The scheelite lies between walls of schist in the shape of a ledge in the same formation that the gold is found, the Baltic being a gold mine equipped with a 10-stamp mill. It has been a good gold producer. Scheelite is the calcium tungstate which is used in the Roentgen ray [X-ray] apparatus. The chief use, however, of tungsten is in the hardening of tungsten steel. The price of tungsten ores from 65 to 75 per cent, is \$2.50 per unit, or about \$150 per ton.” (The Baltic was in the Stringer district, about a mile from Randsburg, as Marcia Rittenhouse Wynn, the granddaughter of W. W. Wynn, recalled: *Desert Bonanza: the story of Early Randsburg Mojave Desert Mining Camp*

(2nd ed., Glendale: Arthur H. Clark Company, 1963): Then in the spring of 1904, W. A. Wickard and several others found scheelite south of the Black Hawk Mine, a gold producer. Other important mine owners included Charles Taylor, a miner; Thomas McCarthy, the owner of a bookstore and newsstand in Randsburg; and C. Grant Illingworth, who owned a general store.

Several mine owners sold out to Edwin De Golia, already a successful mine owner; David Atkins; and Clifton H. Kroll. One stockholder was Frederick W. Bradley, a mining engineer, who became the company’s manager in 1923. (Sources: Articles of incorporation, Atolia Mining Company, California State Archives, Sacramento, and Clark C. Spence, *Mining Engineers and the American West*[:] *The Lace-Boot Brigade, 1849-1933*, New Haven: Yale University Press, 1970; the University of Idaho Press reprinted this volume in 1993.)

[NOTE: Because of an inferior method of microfilming, the prints of several articles in the *Randsburg Miner* are unreadable. Also, I have summarized several articles that contain technical information.]



Atolia mine shown in a postcard from 1908.

1904

“Deposit of Tungsten Ore,” October 15, 1904, p. 2:
Mr. Gideon Boericke, secretary and treasurer of the Primros [Primrose?] Chemical Co., near Philadelphia, Pa., came into town Thursday evening with the intention of sampling and buying tungsten ore. Last spring, Messrs. Wickard & Hanson, whilst working their claim near the Black Hawk, came upon some deposits of tungsten which seemed to be of [good?] quality. They sent samples to the company which Boericke represents but did not hear anything of the [matter] until . . . [The rest of this article is missing.]

“Tungsten,” December 24, 1904, p. 1:
In a letter to the editor of the Mining Journal of London, a writer calls attention to the more extended uses of tungsten, saying:

‘I wish to point out that the great results obtained by the use of wolfram or tungsten in steel for cutting tools of all kinds is small compared to the revolution its use in causing in the manufacture of armor plates for battleships, armored cruisers, and fast torpedo boats; a very much thinner plate can be used with equal resisting power, thus lessening the weight and enabling the vessels to attain a higher speed (a very important factor). Further, the use of tungsten in the manufacture of guns by the Japanese Government, and now being successfully used in the present war, is an object to the whole world, as it gives greater strength and tenacity to the gun metal and enables a powder of much higher explosive power to be used, and thus obtains a longer range.

‘Wolfram steel for railway and train rails, bridge building, etc., will soon be largely used. I am told that many engineers in the United States now specify for its use for the framing of their lofty buildings. In fact, its use in Germany, Japan, and the United State is much more general than in Great Britain at the present time; and what is most important it has come to stay, and is becoming one of the family of commercial metals, like copper and tin, and will have its daily quotation with the other metals.’

1905

“Great Finds of Tungsten,” April 8, 1905, p. 1:
There is considerable excitement just now in Randsburg over tungsten. Many claims are being located and some very good ore is being taken out. Tungsten is a mineral used for hardening and

toughening steel and is not common. The principal mines are in South America, but near Boulder, Colorado, there are several deposits and two or three mills are in operation. About one half of the tungsten used in the United States is taken out there. It is a white quartz, no so clear and transparent as quartz is usually found and is much heavier.

The principal discoveries so far have been made by W. A. Wickard, south of the Black Hawk mine, who has a large vein, or deposit, of seven [unreadable section] which will go from twenty to eighty per cent. Sanderson has a location on farther south which shows up a large ledge or body of very high grade ore. These two claims together with the workings of Churchill, who has eight claims on down the valley near St. Elmo, has within the last week produced more [than] fifty tons of ore that assays values as high as \$450 a ton.

Tungsten is worth \$6 per unit a ton and eighty-four per cent is the highest, as it must have as much as sixteen per cent of lime for a body. Some of that found will run from sixty to eighty per cent which makes a ton of it very valuable. It can be shipped from here to Wilmington, Delaware, where it is treated, for \$27 per ton, but in the form of concentrates would be very rich. A ton of pure tungsten is worth about \$500.

Many prospectors are on the ground and are locating claims, mostly out towards St. Elmo, but miners say there is a showing of it all over the camp. It reminds one of old times to see men getting up early in the morning and rushing out with all sorts of rigs to locate tungsten claims, just as they used to do with gold claims.

“More Finds Of Tungsten,” April 15, 1905, p. 1:
Since last week people are waking up to the possibilities in Tungsten, A constant stream of people are going and returning St. Elmo way. During the week the editor of the MINER has visited nearly all of the camps and the prospect is good.

As you go south, Wickard’s camp is the first one to be encountered. There are a number of gentlemen associated with Mr. Wickard and the company is a strong one. W. A. Wickard, Pat Byrne, and George Gaylord one a one-half interest; the other half is owned by P. H. Myers, Harry C. Rambo, J. Albert Smith, W. H. Abbey, F. A. Gully, and H. A. Blodgett, of Bakersfield.

Mr. Wickard and P. H. Myers are at work. They have a big ledge and considerable ore on the dump.



Atolia in 1909. USGS Hess.

Their shaft is down something like thirty feet and they have worked the claim in several other places to the west and on top of the hill east of the Black Hawk mill and have found good ore there. They have shipped some ore but in small quantities, sending it to the Primas Company, Philadelphia.

The largest lot of ore shipped was by Val Schmidt, who sold a long ton, 2240 pounds, to a representative of the Philadelphia house, receiving therefore \$200. Several other small lots have been sold here but at the time that buyer was here it had just begun to attract attention and few had anything to sell, and fewer still knew the value of it.

Immediately south of the Wickard claim and parallel to it A. E. Sanderson has uncovered a well defined ledge of high grade ore. His ledge is about ten inches thick and in the past week he has taken out over a thousand dollars worth at a very moderate estimate of the ore. Sanderson has two claims.

Thomas McCarthy, Charles Taylor and Henry

Giandoni have several claims on the south and one east of Sanderson, as yet undeveloped but float can be found on any of them. Immediately south of Sanderson, Charley Adams has two claims and on one he had found tungsten with very little digging. James A. Gardner of the MINER office has one claim just east of and adjoining Adams. McCarthy and Roberts join Adams on the south with two claims. Mr. Greenwood and several others have locations south of these claims and nearly all of the country is staked down to Churchill's.

C. H. Churchill, just below the big rock on the Kramer road, has his camp and employs two men to help develop his claims. He has located eight claims west of the road and has found tungsten on all of them. He has one shaft in the edge of the draw down about twenty feet and has taken out some very rich ore, some that he thinks will go 80 per cent tungsten. Mr. Churchill is a man of wide experience as a miner and mineralogist and has great faith in the future of the camp. He is associ-

ated with some men back east and has found good float and located three claims east of the railroad and north of St. Elmo. Just south and east of these claims McCarthy, McDivitt and Gardner have three claims and John McCarthy one. These are all that we are sure of but there are a number of others.

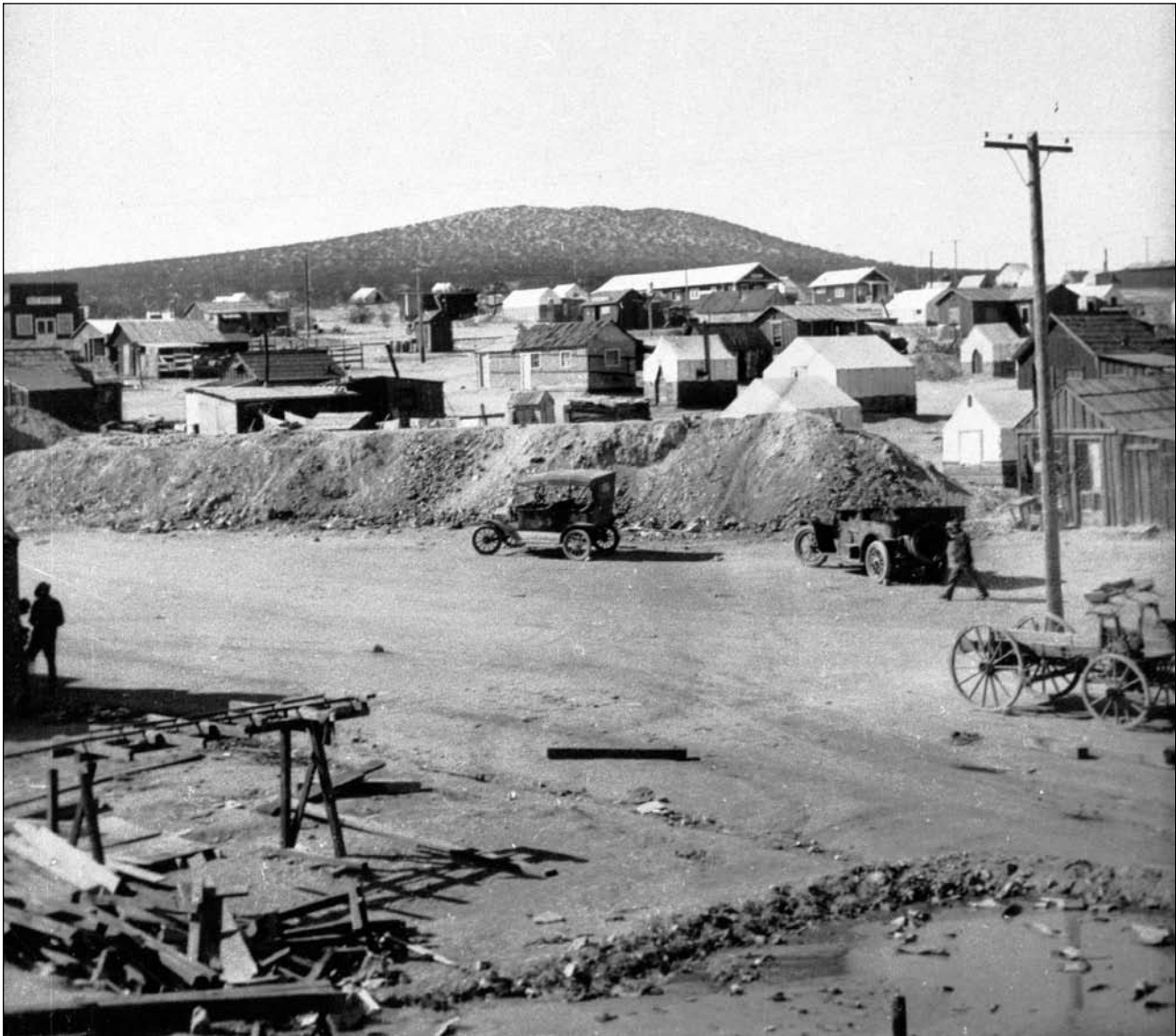
Tungsten is worth nearly \$500 per ton in a pure state and is used in the manufacture of steel for hardening and toughening; it is also used in the form of tungstic acid for dyeing fast colors. It has a specific gravity of 1-2 times the weight of water and is nearly as heavy as lead. Lead ores vary in specific gravity from 5.5 to 8.2. Galena ore is 86.50 lead and 13.40 sulphur. Tungsten is nearly three times as heavy as pure crystal quartz that having a specific gravity of 2.65.

G. W. Lloyd is fixing up his mill to reduce tungsten ore to concentrates and in that form it will be shipped east for treatment. We have no desire to create any undue excitement, or fictitious[?] situation but the prospect at the present time certainly looks encouraging and the undisputed fact remains that already a number of men have the real value before with very little effect or expense.

“Clipping on Tungsten,” April 15, 1905, p. 1:

The great necessity for tungsten as the product from which is derived an acid absolutely essential in hardening steel gives it the importance it holds to the iron manufacturer and its high market value.

Tungsten is a Swedish word meaning ‘heavy stone,’ referring to its great specific gravity which is 6 while quartz, the usual gangue, is only 2.5 to 3. This high specific gravity of all tungsten ores makes



Atolia, March 12, 1916. California Division of Mines and Geology.

them a fine concentrating proposition. The chemical symbol is W for wolfram, wolf's soot, which is a combination of iron and tungsten.

The principal ores of tungsten are wolframite, iron manganese and tungstate; wolfram, iron tungstate; hubrorite [huebnerite], manganese tungstate; scheelite, lime tungstate; cuposcheelite [cuproscheelite], lime and copper tungstate. Tungsten ochre is an earthy form, and an oxide being 80 per cent tungsten and 20 per cent oxygen, which by treatment with acid forms a rich yellow pigment.

Iron tungstate added in certain proportions to steel produces the hard and tough and tough material which is used in the finest cutlery. Another combination produces the steel used for steam lathes and planes and also for ordinance and armor plates. A mixture of iron and eighty per cent tungsten produces an iron almost infusible even in the intense heat of the modern blast furnace. A soda tungsten compound used with starch adds to the stiffening property and renders the material unflammable.

In Colorado the ores are usually found in the granite in veins accompanied by magnetic iron and quartz. In other sections they are found with tin ore.

The ores occur massive crystalline [crystalline] and in earthy, pulverulent [powdery] forms, as incrustations on the gangue. The usual color is brown and dark gray to black, with lustrous crystals intermixed. The streak is reddish brown. The neutral, tungsten, is a dark gray powder.

The only practical field tests are its infusibility before the blow pipe, its insolubility in nitric acid, and, when treated with aqua-regia, its change to a yellow color, and the absence of effervescence.

The prevailing opinion that the ores will not be found at depth is not sustained by development in Boulder county, Colorado, several mills are working steadily on the ores, but the leaders are the Wolf Tongue Company's twenty stamp mill at Nederland, the Union Iron Company's mill at Boulder, and the Boulder County Mines Company mill, all of which are doing custom work as well as working their own output.

Outside of the Nederland belt large and valuable veins of tungsten have been found on the slopes of Sugar Loaf mountain extending northwest and southeast. The mines in both camps are conve-

nient to railroad and in easily accessible situations for mining and surrounded by the necessary accessories, timber and water,

Boulder county mines are furnishing at least fifty per cent of the tungsten used in the United States, and as the output is steadily increasing, they will soon be in shape to satisfy the demand, which is also growing as new uses and enterprises come into the industrial field.—Telluride Journal.”

“**Local Notes,**” April 15, 1905, p. 3, c. 1:

[Excerpt:]

...

Thomas McCarthy, Charlie Taylor and Henry Giandoni have secured a lease and bond on the Lacrosse mine, adjoining the Sunshine in the Stringer district [near Randsburg]. This mine has prospected rich in places and considerable gold has been taken from it. These parties propose to work it and determine what there is in it.

“**Tungsten Notes,**” April 22, 1905, p. 1:

Chas. Churchill has uncovered two large ledges of ore on his claims east of the railroad.

The Hopcraft-Hallford contingency have a body of rich ore on a claim adjoining Ray's.

G. W. Lloyd has his mill arranged for concentrating and made a run on Randsburg tungsten this week.

The usual results of a new strike are manifest. Claim-jumping, quarrels, and threats of legal complications.

Wickard and Myers who have been running a cross-cut at a depth of 28 feet have encountered a very rich deposit.

Thomas McCarthy has bought of A. M. Ray a half interest in the Papoose, east of the railroad north of Churchill's for a consideration of \$450.

Charles Adams has a well-defined stringer on his claim immediately south of Sanderson's and is working like a Trojan to find the main ledge.

We are informed that some business men consider the outlook so promising that they contemplate putting a mercantile establishment at St. Elmo.

Shorty Oakley has sold two claims to the west and adjoining Wickard and Sanderson to Macdonald, Thompson and I. W. Rinaldi, consideration not stated.

Pete Jensen and W. A. Ruffeed have made some

very rich finds on their claims near the large boulders west of St. Elmo. Mr. Jensen is the first tungsten prospector of the camp and has other claims with good ledges of ore as a result of much persistent and systematic prospecting.

“Tungsten Stories Not Exaggerated,” April 22, 1905, p. 1:

James Curry, the Randsburg hay and grain merchant, was in Bakersfield yesterday. Mr. Curry says that the reported strikes of tungsten at Randsburg have not been at all exaggerated and that the discoveries of that mineral are really marking a new epoch in the history of the mining town. There have been a large number of new arrivals at Randsburg since the tungsten excitement, many people coming from Los Angeles and quite a number from Bakersfield.

Mr. Curry says that Wickard and Gaylor, who were among the first to begin prospecting for Tungsten, are now down about fifty feet on a ledge that runs about 10 inches thick. The tungsten on the surface runs anywhere from \$200 to \$400 a ton, and as it is almost as heavy as lead it does not require a very thick vein to make an enormously rich mine.

Some of the deposits of tungsten are in schist and granite, and some are in clay. New finds are being made every day and the indications are that the territory over which it is likely to be found is quite extensive. Deposits have been found all over the old bearing belt and some finds have been made outside the gold belt.

Mr. Curry says that considerable quantities of tungsten have been found on the dumps of old mines where it has been thrown away by gold seekers who did not know its value.

—[Morning] Echo, Bakersfield.

“New Tungsten Discoveries,” Randsburg Miner, April 27, 1905, p. 1:

The tungsten excitement keeps up and new discoveries are made everyday. A visit to the camps today found all the older prospectors developing their properties and new men looking over the ground.

The location of Ray, Taylor and McCarthy, ‘the Papoose,’ seems to be attracting the most attention just now. They are located east of the railroad about a half a mile from St. Elmo, and have developed a fine ledge of first-class ore. It is about twenty inches in width standing perfectly defined and clear of all other matter. They have only gone

down about eight feet but the ledge shows the same as both ends of [the] shaft.

Adjoin[ing] them on the east and south Churchill, Mertz, and Curran have located a fraction 600 by 700 by 300 and have developed a ledge of good high grade ore which with a little more work may prove equal to the other.

P. Mertz, a young man from Breckinridge, Colorado, who is with Churchill, has three claims west of the Papoose, and Mort Curran, also with Churchill, has two south of the Papoose.

Charles H. Churchill [Churchill], at his camp west of St. Elmo, is doing development work on his claims and has several shafts down twenty or more feet.

G. W. Lloyd started his mill today on ore from Wickard and Company’s claims. This is somewhat of an experiment as no tungsten ore has ever been worked here and everybody will anxiously await [the] results.

“Tungsten Developments,” May 4, 1905, p. 1:

Two new ledges have been discovered on the Papoose.

J. S. Roberts is developing the Par, and has found considerable good float.

Bernard Vogt is prospecting on his claims adjoining Churchill’s and east of the railroad.

Chas. Churchill has given a sixty-day option on his group of claims west of St. Elmo for the sum of \$20,000.

Mike Doland has a good claim near the Papoose, and has found good float. He says he can hear distinctly the ‘chug-te-chug’ of an auto-to-be-a-mobile coming over the sand dunes.

Pete Jensen and A. M. Ray are now in permanent camp on the Papoose, and with Charles Taylor and John McCarthy the four are taking out an average of 1000 pounds of ore per day. They have uncovered the ledge for a distance of over 400 feet and varies in width from 10 to 18 inches wide.

[Appended to the above article is the following letter:]

ALTOONA, PA., Apr. 28, 1905.
RANDSBURG MINER,
California.

Dear Sir:

I notice in the Los Angeles Mining Review of the 22nd inst. A quotation from your paper in regard to Tungsten ore being found at Wickard’s Camp,

St. Elmo Dist. etc. What I would like to learn is the proper address of miners of that ore, also other high grade ores, with a view to purchase for highest cash prices, soon as loaded on cars or as early as get bills of lading: [;?] could also arrange to make cash advance, if wanted, if can buy said ores at satisfactory figures to all concerned, therefore, will thank you much for any favorable attention given this by which can hear from the producers or miners or owners of such ores. Very greatly oblige.

Respectfully,
G. A. MCCORMICK

The above letter is entirely self-explanatory, however, we might add that Mr. McCormick deals extensively in steel rails and railroad equipment.

“**Local Notes,**” May 11, 1905, p. 3:

There is a specimen of Tungsten ore, no larger than a football, in McCarthy’s that weighs forty lbs. Call in and heft it.

...

Exploitation of tungsten still continues and the work of developing the mines already discovered is going on rapidly. Many inquiries are coming into the camp in regard to the number of the mines located and their development.

...

Returning to Randsburg a few nights ago, after a short absence hungry and tired we hurried down to our usual boarding houses for supper and found the place closed. A big sign tacked to the building told the story: ‘Tungsten, four miles south.’ This place closed. The unexpected often happens in a mining camp.

“**Local Notes,**” May 18, 1905, p. 3:

The tungsten men are feeling just as good as ever. Thomas McCarthy had a return from a sample sent to San Francisco which showed over forty per cent of pure tungsten. This surprised the company up there so much that they wrote down and expressed a doubt of any more to be found as good, but said they would send a man down to investigate.

“**Tungsten Concentrator,**” May 25, 1905, p. 2:

“H. C. Rambo, who has just returned from Randsburg, says that arrangements are being made to build plants for handling the lower grade of tungsten. The thing that is delaying this work is the uncertainty as to what particular kind of mill and concentrator will handle the product best. As Mr.

Rambo says they are all afraid that if they get one outfit installed another more economical or more effectual method will be discovered and the first investment will be lost. As far as the tungsten is concerned there is already enough in sight to warrant a plant being erected.

“Mr. Rambo and his associates have uncovered a good quantity of the ore on their claims. They have taken out quite a little quantity of a high enough grade to ship and have milled a little of the poorer grade, but their chief efforts have been toward getting a good quantity of the ore in sight so as to form some estimate as to the amount they will be able to produce.—Echo.

“**Tungsten Analysis,**” June 1, 1905, p. 1:

[Excerpt:]

RANDBURG, May [-?]

To the Editor of The [Los Angeles] Times:

I herewith hand you a specimen of tungsten ore. We know but little about it, but it is found in paying quantities, ledges, near our town[,] and is creating a great deal of interest and I would be pleased if in the Times you would kindly give us an analysis of it, its uses and value.

Yours very truly,
C. J. McD.

ANSWER:

This department of the Times gladly gives analysis on any mineral of benefit to California, at the same time our responsibility ends with sample forwarded.

The specimens received are gray white in color, of adamantine luster and about 4.4 in hardness. Of uneven fracture. Massive in structure, lenticular or pyramidal in habit. Brittle formula or of calcium, one of wolfram and four of oxygen. Density 5.8. Test: powder finely some of the mineral, place some of the pulverized ore in a small evaporating dish, over which ore pour some strong c. p. hydrochloric acid. Put evaporating dish and contents on top of stove, let contents boil slowly for twenty minutes, at the end of which time you will have a yellow solution, set dish and contents aside for an hour or so, when a canary yellow deposit separates out from the solution, this deposit is known as tungsten trioxide (tungstic acid). If aqueous ammonia be added to some of this powder it forms a solution, which solution, if set aside for three or

four hours[,] crystallizes into a white salt encircling on upper pan [pans?] and inside of evaporating dish. Further testing: If granulated tin be added to the boiling hydrochloric acid solution a blue color soon forms, which color changes to brown on further boiling. A chip of the mineral fuses B. B., to a partly transparent glass which is opaque when cold. If microscopic salt be added to a little of the powder and the flame applied, a bead is formed, green when hot, and blue when cold. Scheelite is found in brown, yellow, green white, pale red, or orange colors. Streak usually white. Your samples analyze as follows: Lime 22 per cent. Tungsten trioxide 78 per cent equals 100, but there is no guarantee that so high a per cent. will be obtained as a general average, from the mineral deposit. . . .

“New Tungsten Discoveries,” June 1, 1905, p. 3:

Prospectors all over the West now realize that the search for the rarer elements is a profitable pursuit, says the Denver Mining Reporter. Tungsten especially appeals to the prospector, not merely on account of its high value, but also because of the steady demand and the ease with which tungsten minerals can be concentrated. Almost simultaneously we receive news of two new localities for tungsten, one in Southern California and the other in Washington. The California field is in the desert country of San Bernardino, near St. Elmo. Prospectors are busy opening up very promising veins. We are informed that very clean mineral twenty inches in width has been shown up in the Papoose claim. Other claims also contain tungsten minerals in more or less quantity. The Washington district reported as containing tungsten in Stevens county, and the discovery was made in the Deer Trail district. A shipment of one ton of ore to Pittsburg is said to have realized a very high price, and prospectors are making a large number of locations. If these new deposits come up to expectations and the older producers maintain their output tungsten mining will become a very important branch of the industry.

“Tungsten,” July 6, 1905, p. 1:

The following article is clipped from the Engineering and Mining Journal of June 1, and will doubtless be interesting reading to many of our people.

‘TUNGSTEN.—1. I would like to ask whether the price for Tungsten (best), at present \$1.25 per lb., recorded in your paper under ‘minor metals,’ is the price for the ore, or whether this is the quotation on the pure metal (tungsten). What does ferro-

tungsten (37 per cent) mean? Is this tungsten ore.

‘2. Where is tungsten found in America, especially in the United States, and which are the chief producing mines of tungsten in this country?’

‘3. How large has the production of tungsten been in the United States, and is it increasing? What is the world’s production? How large in [is] the consumption of tungsten and is it increasing?’

‘4. Can you recommend a book on tungsten? Please give me such general information on the metal as is possible.

—F. A. B.

ANSWER 1. The price of tungsten given in the Journal is for refined metal. Ferro-tungsten is an alloy of iron and tungsten, which is used in making steel. Tungsten ore is usually sold by assay. It is essential that the ore be free, or nearly so, from phosphorus and sulphur, but the presence of carbon and silica will not be considered injurious. At present, an ore averaging 60 per cent WO₃, and containing not more than 0.25 per cent phosphorus, and 0.01 per cent sulphur can be sold in New York at \$7 per unit of tungstic acid, equivalent to \$420 per long ton. For higher-grade ore, up to \$7.50 per unit is quoted, because just now the demand exceeds the supply. It is customary to contract for ore on a basis of 90 per cent cash on delivery f. o. b. New York, the balance of 10 per cent being retained for a month to allow time for comparison of assays and adjustment of possible differences between buyer and seller.

‘2. Tungsten is found in Arizona, Colorado and Connecticut; in small quantities in other places. The largest production has been from Colorado.

‘3. The production in the United States for several years past has been between 3000 and 5000 short tons of crude ore, which will yield from 200 to 350 tons of 50 to 60 per cent concentrate. The production of metallic tungsten has been from 80,000 to 90,000 lbs. yearly in the same period. The increase is slow. The production has a tendency to increase faster than the consumption. There are no figures accessible for the world’s production; but in Great Britain and Germany the output and consumption are larger than in the United States.

‘4. There is no book on tungsten, and only short references to the metal can be found in most of the metallurgical works. To sum up the information regarding the metal, we reproduce here part of an

article which was published in the Journal, Nov. 10, 1904: 'Metallic tungsten has a sp. gr of 18.7, is practically free from carbon, can be welded and filed like iron, and when used in tool steel the alloy preserves its hardness, even when heated to temperatures that would rapidly draw the temper from ordinary high-carbon steel. In the manufacture of permanent magnets, used in the construction of electrical meters, the steel employed contains approximately 5 to 6 per cent tungsten. An alloy of 35 per cent tungsten and 65 per cent steel will make a shell for lead bullets that has much higher penetrating power than ordinary steel. Tungsten steel may also be used for armor plate. Metallic aluminum can be hardened advantageously with tungsten, its resistance to oxidation making it much superior to copper. A small percentage of tungsten will also greatly increase the carrying power of spring steel. Tungsten steel is employed in the sounding-plates of pianos.

'Vegetable tissue are rendered incombustible by the use of tungsten compounds. Usually the material to be treated is coated with tungstic acid and glue, or with tungstate of soda and muriatic acid in a thick solution of glue. A tungsten compound is also employed in the preparation of fluoroscope used in X-ray observations.'

"The Tungsten Mines," July 13, 1905, p. 3:

"The development of the tungsten properties is steadily going forward notwithstanding the hot weather. On the Papoose the [illegible section] if possible better defined as they go down. They will ship a car of ore in a few days and the result of that shipment will be looked forward to with a great deal of interest by all concerned. In the meantime, everybody feels good and confident of the future."

"The Papoose," July 20, 1905, p. 1:

On Sunday we visited the tungsten district and while it was too warm for much investigation we took a good look at the 'Papoose.' On this mine the owners have found the ledge for more than 400 feet on the surface. Only in one place have they attempted to sink on the ledge and there the shaft is down about eighteen feet, showing one of the best and most clearly defined ledges in the camp. The ledge is about two feet and seems to be pure tungsten. The appearance of the ledge at this depth gives the best evidence of its permanence. It is a clear cut distinction between it and the granite on either side.

The boys have had no money to hire help and the weather has been very warm to work in the sun, but every day finds them at it. They are carefully assorting and sacking a car load of ore and will have it ready to ship now in a short time. The first shipment they will send to Wilmington, Delaware, and the freight on a thirty ton car is expected to be in the neighborhood of \$400.

When the returns from that shipment come there will be either great rejoicing or great depression. The owners feel confident and believe they will not be disappointed. Very few people know much about tungsten ore but from all accounts, the best information obtainable, the shipment should bring several hundred dollars per ton. If the ledge should pinch out as the present depth, and there is not the slightest indication of it, they still have enough to bring them many thousands of dollars.

"First Shipment of Tungsten," July 27, 1905, p. 2:

McCarthy, Taylor & Co. expect to ship a thirty ton car of tungsten ore to Wilmington, Delaware on Monday next. They expect the car in Saturday evening. They have had correspondence with many parties, even from the Krupps in Germany and have all kinds of prices offered, but for the first venture they have decided to try Wilmington.

The Papoose is looking better and showing a larger body of ore with every day['s] work.

"That Car of Sheelite [Scheelite]," August 10, 1905, p. 3:

Two weeks ago the Miner stated that a carload of tungsten ore would be shipped from the Papoose mine the next week. The car is not yet shipped but is mined, carefully assorted and sacked and only awaits satisfactory terms of shipment and sale, when it will be started.

Such is the scarcity of tungsten—Scheelite, that probably never before has there been a carload of such ore shipped in the United States, certainly not from the State of California. The owners found this difficulty when they began negotiations that only a few dealers knew anything about it, or its value and they could get no definite or satisfactory offers.

They have communications with dealers everywhere, at Wilmington[,] Delaware, at Philadelphia, Pa[,] and finally to Germany, where it is nearly all treated. They have had several offers running from three to eight dollars and have finally decided to await definite proposals from Germany. They want to make no mistake in getting full value. In the

mean time they are sinking the shaft and taking out ore. The ledge at thirty feet looks better than ever, the ore being of a high grade.

“Greenwood & Thompson,” August 17, 1905, p. 1: Greenwood and Thompson are doing work on their tungsten claims near St. Elmo. They are getting out considerable milling ore but not rich enough to ship. They find it in an old shaft nearly a hundred feet deep, which had been sunk years ago for gold. They have a lease on the claim, and hope to find paying gold ore before they quit. They have found some molybdenum, a rare and valuable metal which in its pure state is quoted at \$2.75 per lb. This is the first time this metal has ever been found on the desert. Mrs. Greenwood has developed a capacity for mining that some men might imitate to advantage. Unaided by others she has sunk a shaft eight feet in depth, taken out considerable and run several prospecting cuts on the surface which amount in labor to far more than the shaft. She can and does drill holes, load them with dynamite and shoot them, the first woman to do this in this camp. Mr. and Mrs. Greenwood have a comfortable little frame house with a tent adjoining and a view outlook to the south and east that would be hard to match anywhere. They can see the train almost from the time it leaves Kramer, twenty-five miles away, and until it passes their door. James Thompson and wife live a short distance away with similar accommodations [accommodations].

[**Untitled**], September 7, 1905, p. 2:

The shipment of a car-load of tungsten this week, by McCarthy Taylor & Co., is an event of great interest all over the state and means much to Randsburg.

“Two Cars Tungsten[:] One To Germany,” September 14, 1905, p. 5:

Two carloads of tungsten were shipped from here last Monday night, instead of one as stated in the Miner of last week, one going to Philadelphia and the other to Germany by way of San Francisco.

The cars were left on the track Saturday. A regular picnic was made of the loading, about forty of our people going down to assist.

Both cars were loaded with high grade ore running 72 per cent tungsten. This is the first shipment of tungsten ever made from this state in carload lots and marks the beginning of a new industry in Southern California.

The following gentlemen are interested in the shipment: McCarthy Taylor & Co., Pat Byrne, W. A. Wickard, C. G. Illingworth, F. Mertz, and J. G. Porter. There is an abundance of ore which goes about 40 per cent which can be handled to a great advantage when they have erected a mill on the ground, which they intend to do within a year.

When the returns come in we will celebrate.

“Tungsten,” October 5, 1905, p. 1:

[SUMMARY: According to the Mining Reporter of Colorado, the Beaver Metallurgical Company has completed the installation of a large metallurgical and chemical plants at Beaver Falls, Pennsylvania. So far, the company will specialize in the production of metallic tungsten, ferro-tungsten, and other alloys and compounds of tungsten.]

“A Sale of Sheelite [Scheelite],” October 19, 1905, p. 1:

Messrs. Baverstock & Staples assayers of 222 West First St., report they have sold a trial shipment of sheelite [scheelite] for the Minnehaha Mining Co.

The shipment was about three tons and will net the company about \$185 per ton. The mine is situated about seventeen miles from Caliente, Kern county, California and has besides the sheelite [scheelite] values, a good gold vein averaging about \$20 ton, upon which a 3-stamp mill is now being erected, furnished by the Baker Iron Works of this city.

If this shipment of sheelite [scheelite] proves successful, Messrs. Baverstock and Staples say they will contract for the whole output of the mine, as they have European as well as Eastern orders to fill.—L. A. Mining Review.

“Local Notes,” October 19, 1905, p. 5:

It is rumored that Chas. Schwab the steel magnate, paid Randsburg a flying visit last week, under the name of French. French, accompanied by an expert, came in on Friday[']s train, examined the tungsten properties of the McCarthy, Taylor Co., and left the same night. The expert said French was a man of means and greatly interested in steel production.

“More Tungsten,” November 9, 1905, p. 3:

“Wm. Houser, A. Castro and Mike Schmalak have located four Tungsten claims about 7 miles from Randsburg, just off the old Borax road near Garden Station. The ore according to samples assayed run 20 per cent tungsten, the ledge averages 10 feet in width and in some places is much wider. Mr. Houser says there are tons upon tons of ore

in sight and that after a little further investigation they expect to erect a mill on the property and can then work low grade ore to advantage.”

1906

“**Receives First Payment,**” January 11, 1906, p. 3:

Thomas McCarthy has gone to San Francisco in response to a summons from the parties who have negotiated for the tungsten property to receive the first payment. It is reported that the price is \$114,000 and includes all the tungsten claims in the immediate vicinity of the Papoose mine.

Before the parties would conclude the deal they required McCarthy, or some on[one] else, to secure options on all adjoining claims, which he did, and then the deal went through. The parties have until July to investigate and develop at which time the entire payment is to be made. This is a good thing for the boys interested. There are several of them and this sale will make them all independent. It is still a queer thing to think about that tungsten should be developed here, in a purer state and better defined ledges, that [than] elsewhere in the United States.

“**Shipment Of Tungsten,**” January 18, 1906, p. 3:

The DeGolia Atkins Co. who recently purchased the tungsten properties belonging to the McCarthy Taylor Co., will ship a car-load of about 25 tons Monday next. The shipment goes to Germany by way of New York and is of a higher grade than the last shipment made by McCarthy Taylor and Co.

...

Thos. McCarthy, who went to San Francisco to receive the first payment made by the purchasers of the tungsten properties, is expected to return next Saturday evening with his coin. We predict ‘a hot time in the old town’ Saturday night.

...

Mr. [Edwin] DeGolia, who represents the new Tungsten Co., left last evening for San Bernardino on a short business trip.

“**Another Car Of Tungsten,**” January 25, 1906, p. 2:

Another car of Tungsten was shipped on Monday last. This car went by express via the Santa Fe and Erie railroads to Liverpool, England. This shows some enterprise on the part of Wells Fargo Agent here as it was thought impossible to make such a

shipment by express in such bulk. The car contained nearly twenty for [four] tons of high grade ore.

The deal has gone through and the car went out in the name of DeGolia and Atkins who represent the present owners.

Thomas McCarthy returned Saturday night from San Francisco with the first payment, ten per cent of the purchase price. Under the terms of sale fifteen per cent of purchase price must be paid April 3d, 1906. In the meantime all returns from ore shipments are to apply on the purchase price. C. S. Taylor is for the present superintendent [superintendent] of the works.

“**Notice of Non-Responsibility**” (ad.), January 25, 1906, p. 2:

From and after this date we will not be responsible for labor or material employed or furnished on the Papoose, tungsten mine.

A. D. Ray,
Thomas McCarthy,
Chas. S. Taylor,
Henry Giondoni,
H. P. Jensen.
January 4, 1908 [1906].

“**Tungsten Owners Here,**” February 15, 1906, p. 3:

“On Monday evening four men, Messrs. E. B. DeGolia, Mr. Bradley a mining engineer and president of the Bunker Hill and Sullivan Co., of Idaho, Senator Voorhees, of Sutter Creek and Mr. Street, representing large mining interests, came in on the train and stopped off at the tungsten mines. They came to Randsburg and stayed all night going the mines next morning, taking the train from there Tuesday evening. Vrey [Very] little of their plans could be learned other than that a gang of men will be kept at work taking out ore and the low grade stuff will be shipped to Barstow for concentration in the Barstow Mill, the first car of ore going out Wednesday of this week.”

“**Local Notes,**” March 1, 1906, p. 3:

[Excerpts:]

...

Chas. Taylor and Henry Giondoni have leased the 600 feet of the east end of the Little Butte mine to Jan. 1, 1907[,] including the machinery shaft and buildings.

The first car of Tungsten was shipped last week to

the sampling works at Denver from the American Tungsten Company at Boulder, Colorado.

“Developments At Tungsten,” March 22, 1906, p. 2: Charles Taylor, Superintendent at the tungsten mines, says there are now working 22 men and taking out a high grade of ore. The railroad company has put in a spur and they now load cars at the mine. The shipping to the Barstow mill has proved a success and they concentrate about five cars of ore into one of high grade stuff.

About the first week in April they expect to ship two cars of high grade, selected ore direct from the mines to Europe, Liverpool, England[,] or to Germany. They have “quite a village out there and they are putting on airs. Just plain Tungsten was not good enough for them they now call the new camp and station ‘Atolia.’ Just what that word means we don[‘]t know, but if you hear anyone saying something of that kind you will know that it is tungsten.

A new Fairbanks and Morse gasoline hoist has been ordered and is now on the way and efforts are being made to have a postoffice established with the soft, smooth-founding name of ‘Atolia.’ Our genial friend Charlie Taylor, Superintendent, is despondent and says he is fourteen kinds of a d—fool for ever selling, or consenting to sell, a property in which he had a one sixth interest for \$114,000 when it is now so soon after, easily worth a million, and of course we are all sorry with him.

[NOTE: Atolia is the contraction of the surnames De Golia and Atkins.]

“New Telephone Line,” March 29, 1906, p. 3: A new telephone line was strung last week to the tungsten mines. Work was begun Friday afternoon and communication was opened up Sunday evening. Pretty quick work. Harry Wilson was the man in charge. Tungsten is in open communication with the world.

“Another Car Of Tungsten,” April 5, 1906, p. 2: A car of high grade tungsten ore was shipped last evening by Superintendent Charles Taylor, for the Tungsten company. This car, of 27 tons, goes by Wells Fargo Express to Roberts and Bense, Hamburg, Germany, and will go to Chicago over the Santa Fe, thence to New York and across the ocean. The car is high grade and is probably worth \$10,000.

All the men interested in the 42 claims which were included in the sale of the Papoose mine were made happy on Tuesday of this week by the payment of the second installment of the sale price of \$170,000, which was fifteen per cent. Or 25,500 dollars.

“New Firm,” Randsburg Miner, April 5, 1906, p. 2: Daniel Gunderson, principal of our schools, has bought the business of Thomas McCarthy and will hereafter have the entire newspaper distribution of the camp, as well as the book, stationery, cigars and notions of the business.

Mr. Gunderson has been a teacher in our schools for two years, is very well known and liked and will doubtless succeed in his new venture. His brother, Robert Gunderson, will have charge of the business until school is out.

“Local Notes,” April 5, 1906, p. 2: Thomas McCarthy, the man longest in business in Randsburg, coming here in September, 1896, and going through all the varied ups and downs, strikes and fires, but still the same genial, good natured Thomas, has sold his business, books, news and papers to Daniel Gunderson and retired on the first of this month. Mr. McCarthy has not fully decided what he will do, but rather thinks he will buy a farm somewhere and settle down to rural life. His interest in the Tungsten mines will give him enough money to establish himself very comfortably wherever he may choose to locate, where he and his little family may enjoy the future, without regret at the long years of privation spent on the Mohave desert. The Miner wishes him an abundant success wherever he may go, or whatever he may engage in.

“Local Notes,” April 12, 1906, p. 3: Mining & Scit. Press [San Francisco]. Charles Taylor, superintendent at the Tungsten mines, says they are working 22 men and taking out high grade ore. The railroad has put in a spur and they load cars at the mine. Shipping to the Barstow mill has proved a success and they concentrate five cars into one. They now call the new camp and station Atolia. A new Fairbanks-Morse gasoline hoist has been ordered for the mine and is now on the way.—George and Harry Swartout and B. B. Summers are taking rich ore from the Minnehaha, and expect to have a milling soon. The Minnehaha lies just west of the Yellow Aster mines and some of the richest ore ever taken out in the camp came

from this mine.—In drilling for water near the wells in Goler Wash much difficulty has been experienced in going through the boulders which compose all of the drift after getting a little below the surface and the work is slow. They have one well down over 400 ft. and struck the water 30 ft. before reaching that point. They expect to continue at least another 100 ft. The new deep well east of Johannesburg has never been thoroughly tested, as it is a difficult job to pump that depth, but the company has nearly completed arrangements to test its capacity, which will be done soon.

[NOTE: I have summarized much of this article.]

R. H. Van Wagenen, “**Tungsten in Colorado**,” April 26, 1906, p. 1:

Under the above heading the Mining Magazine publishes a long article on tungsten in the April number, and purports to give a list of countries where they ore is found, but among them fails to name our own State of California. Right here in Randsburg large quantities of purer ore are now being mined and milled than in any other location in the United States and in no other section do they get ore from the mine, without concentration, pure enough to ship[.] The article deals largely in scientific and geological terms and is not therefore understandable by the average miner, but below we give portions of it:”

[SUMMARY: According to the Colorado School of Mines Bulletin, published in January, 1906, tungsten is found in Australia, Cornwall, England, Saxony, Germany, Spain, Portugal, Sweden, Austria-Hungary, Bohemia, Canada, Brazil, Peru, and Japan. Inside the United States, tungsten is found in Arizona, South Dakota, Connecticut, Nevada, Washington, New Mexico, Oregon, Idaho, Montana, North Carolina, and Colorado. The tungsten ores in Colorado occur in Boulder, San Juan, Gilpin, Lake Ouray, Teller, and Dolores counties. In San Juan County, the mineral of tungsten consists of hubnerite [huebnerite], and in Boulder County, the mineral is wolframite, which has been found at Salina, Wall Street, Gordon Gulch, Caribou and Nederland. The mills are adapted to processing either huebnerite or wolframite.

[At the Great Western Exploration mill, the ore is shoveled onto a 2-inch grizzly, the oversize of which goes to a 7 by 10 Blake crusher. The ore is then moved to a ten-stamp battery. The stamps

which weigh 1,000 pounds, drop eight inches 80 times per minute. The battery crushes through a 20-mesh screen, and the product is laundered to classifiers, which makes two products. The larger size goes to a Wilfley concentrating table No. 1, while the slimes pass to another classifier, a Wilfley concentrating table No. 2.]

“**Local News**,” April 26, 1906, p. 3:

[Excerpt:]

...

Thomas McCarthy has bought a farm near Ukiah, in Mendocino county, and left Randsburg with his family Tuesday for their future home, having closed up all his business here after a continuous residence since September, 1896, just when the camp was starting. He is of the opinion that he can do better, especially with regard to raising [raising] his children, on a farm than in the town. From the sale of the Tungsten properties, his business and his houses, Mr. McCarthy is possessed of a sufficiency of this world's goods to make him and his family comfortable for the future, if rightly managed, and there is little doubt he will see to that, having had experience with the other side. The best wishes for himself and family, of a lot of friends, go with him.

“**Local Notes**,” Randsburg Miner, May 10, 1906, p. 3:

[Excerpts:]

There is quite a little camp at Tungsten and the roofs of all the house have been painted red.

W. Greenwood was in from Tungsten Tuesday of this week, the first visit he has paid Randsburg for nearly a year and he only five miles away.

...

William Bouchard has moved his handsome little house out to Tungsten and contemplates building a large one on the same lot on Butte avenue in the near future.

...

The Tungsten people have installed two large galvanized iron water tanks and who now [will] get their supply from Hinkley by the carload at a very much less cost. The two tanks will hold a full tank car of water.

J. C. Robinson and E. H. Spencer arrived in town on Monday of this week on business and will remain for a few days. Mr. Robinson is an old resident of the camp, but has not been here for five

years until now. He finds many changes.

“Local Notes,” May 17, 1906, p. 3:

[Excerpts:]

...

Miners' tents and cabins dot the hills about Tungsten, all of whom are busy about that locality prospecting for the mineral more valuable than gold—tungsten.

...

Tungsten can rightfully claim being the neatest mining camp on the desert. With its painted houses with red roofs it presents a pleasing picture of color in the clear air of the desert.

...

A large bunker, at a convenient elevation, has been erected at the Tungsten mine. The ore will hereafter be dumped inside instead of on the ground as heretofore, making it easier to load and haul to the cars. The engine and hoist, which formerly stood some distance off, has been moved close to the shaft.

“Another Car of Tungsten Shipped,” May 24, 1906, p. 3:

Superintendent Charles Taylor of the Tungsten mines shipped another car of tungsten last Saturday to Germany. The car was billed out at 20 tons of high grade ore and went by Wells Fargo express. This is the fourth car of high grade ore shipped direct from the mines, without other process than that of careful selection, in the last few months.

A car was shipped out to Barstow for concentration about a week ago, as the greater part taken out is handled in that way, three or four tons being reduced to one to avoid cost in shipping. The tungsten mining industry of this place is making remarkable progress and is remarkable, as nowhere else in the United States is there found such quantities of it, nor of such high grade ore. To show how little knowledge men had of tungsten, when the Randsburg railroad was built eight years ago, the graders cut right through the Papoose, which is the principal mine, and did not recognize the ore as valuable.

“Local Notes,” May 24, 1906, p. 3:

...

C. G. Illingworth is erecting buildings and sheds in the corral next to his store to shelter his horses and wagons from the sun and wind.

“Local Notes,” Randsburg Miner, May 31, 1906, p. 3:

[Excerpts:]

Charlie Taylor, superintendent at the tungsten mines, spent several days in Los Angeles last week, returning last night.

...

Leo McCarthy has accepted a position in the Press office and is taking lessons in the art preservative. He bears the ear marks of a good newspaper man and may some day make his mark in the newspaper world.—Ukiah Press. That is good news from Leo, but in after years when he becomes learned and famous he must not forget that he took his first lessons in the Randsburg Miner office. We have always taken a lively interest in that boy and wish him all kinds of success.

“Local Notes,” June 28, 1906, p. 3:

[Excerpt:]

Superintendent Charles Taylor will ship another car of high grade tungsten ore to-morrow. This is the fifth car sent out.

“Tungsten Claims Sold,” June 28, 1906, p. 3:

Mr. J. E. Pagh last Saturday sold the South Side Group of quartz claims, consisting of the Iowa, X-Ray, Tom Reed, Robert S and Michigan, to the Stauffer Chemical Co. of San Francisco. The claims are about one and one-half miles above the St. Elmo station in a northerly direction, and about four miles east of Randsburg. The prospects indicate valuable tungsten deposits.

“Local Notes,” July 5, 1906, p. 3:

[Excerpt:]

A new postoffice at the tungsten mines, just over the line in San Bernardino county, has been established with the name of Atolia. Charley Taylor is the new postmaster and the office is about ready for business.

[NOTE: The post office was officially established on May 10, 1906, according to H. E. Salley, *History of California Post Offices*[:] 1849-1990, edited by Edward L. Patera, 2nd edition. [Lake Grove, Oregon:] The Depot, 1991, p. 11.]

“Another Tungsten Shipment,” July 5, 1906, p. 3:

Charley Taylor, Superintendent at the Tungsten mines, shipped another car, the fifth, of high grade ore direct to Germany by Wells Fargo express. This car contained 20 tons and went out Friday evening last. The deepest shaft on the Papoose mine is a

little over 150 feet and runs about the same.

“Local Notes,” Randsburg Miner, July 12, 1906, p. 3: The tungsten mine near Randsburg is operating and shipping the ore. At first the tungsten mineral (scheelite) was found only in detached boulders; now the vein from which these fragments come has been found. It is reported here that \$27,000 has been paid on the purchase price, which is \$100,000, and after paying all expenses, the owners have over \$25,000 in the treasury, all from the proceeds of the operation.—Mining and Scientific Press [San Francisco].

“Local Notes,” August 2, 1906, p. 3:

[Excerpts:]

Charlie Taylor, superintendent of the tungsten mines and postmaster of the new town of Atolia, is spending a few days at Los Angeles.

...

Thos. McCarthy, A. N. Evans and Marvin Luce have organized the Yokayo Lumber company and purchased the planning mill and saw mill property belong to the Evans & Orr estate. They are now operating the planning mill at this city [Ukiah] under the superintendency of A. N. Evans, one of the best mill men in the county and are working a full crew of men. They will start the saw mills at Low Gap in a short time, or as soon as things can be put in readiness for running and will furnish their patrons a grade of lumber equal to that to be found anywhere in the market. They have secured a number of acres of fine redwood time from George McCowen in addition to that held by the Evans & Ore estate from which lumber will be cut. Martin Luce will be in charge of the saw mills and the teaming. The business management of the plant will be in the hands of Thomas McCarthy, a well-known capitalist and business man who recently moved here from Randsburg.—Ukiah Republican Press.

“Local Notes,” August 16, 1906, p. 3:

[Excerpt:]

Charlie Taylor, the genial Charles, has returned to the strenuous life at Atolia, after an absence of three weeks, enjoying six plunges a day in the briny deep, while stopping at Long Beach, afterwards visiting Hanford [his former home] and the San Joaquin valley.

“Local Notes,” October 4, 1906, p. 2:

[Excerpt:]

October 3d was the day on which the last payment on the tungsten mines was to be paid. We have not heard definitely, but have no doubt the money was paid into the bank at San Francisco by [Edwin] De Golia and [David] Atkins.

“Mill Run On Screenings,” October 4, 1906, p. 3:

Charlies Koehn has just finished a mill run of 50 tons of screenings from the Kinyon and Wedge mines which went \$9 to the ton. This was on the two-stamp mill owned by Parker and the run was made by Jack Nossor and J. M. Rice, who have a lease for a year. The ore from the Wedge and Kinyon mines was of very high grade and nothing was run but the best. So there are several thousand tons of screenings that will pay to mill. The first run was simply a test. Koehn said that if it went \$4 per ton it would pay to run. Nossor & Rice had an option for \$100 in case the first run paid. They paid the money and secured it.

“Local Notes,” December 6, 1906, p. 3:

[Excerpt:]

The new mill is rapidly going up at Atolia and 26 men are now on the payroll of that prosperous company.

“Mill At Tungsten,” December 13, 1906, p. 3:

The machinery for the new mill at Atolia (Tungsten) is on the ground and five men are now at work putting it up. There will be five concentrators with a capacity of 15 tons daily. They will have six gas engines installed when the mill is completed, and Atolia begins to look like a town.

1907

“Local Notes,” February 7, 1907, p. 3:

[Excerpt:]

Charles Taylor, superintendent of the tungsten mines at Atolia, reports everything progressing finely. The mill is nearly completed, and will be running shortly. An automobile will will [sic] soon make regular trips between Atolia and Randsburg.

“The Tungsten Mines And the New Mill,” February 21, 1907, p. 3:

Atolia is now one of the most active little mining camps in this part of the desert. Situated about four miles south of Johannesburg on the Randsburg railroad the buildings are located on a street, over 100 feet wide, running east from the railroad and immediately adjoining the principal mine, the ‘Papoose.’

This mine has been worked less than a year and

grows better all the time. The principal working shaft is down between two and three hundred feet. Heretofore they have either shipped their ore direct to Germany, or have had it concentrated at Barstow and then shipped. Now they are erecting a mill of their own and have it near completion.

The new mill building is 80x36 feet and is very substantially constructed. The timbers are heavy and the foundations are concrete. The main ore bin will hold 100 tons and the bin from which the ore feeds to the rollers is 71 feet above the foundations and holds 50 tons. The outside is covered with galvanized corrugated iron and the whole is surmounted with a flag pole of different sizes gas pipe running up over 100 feet from which will float a fine American flag, now on hand and simply awaiting the completion of the mill to be unfurled to the desert breezes.

In the concentrating room, which is 40x40 feet, will be located six 6-foot Fruevanner concentrators and the pulp belt elevator to carry the water and tailings to the settling tanks. These have a capacity of 18,000 gallons for pumping back purposes. It will require about 2000 gallons per hour and they have a reserve tankage capacity of 60,000 gallons. The water will be brought to the mill on cars from Hinckley on the Santa Fe road.

The mill is a Huntington roller mill with a capacity of 12 to 15 tons and is run by a Fairbanks & Morse gasoline engine of 15 H. P. A second engine of the same size and make will run the rock breaker, the concentrators, the centrifugal pump and belt elevator, located 14 feet below discharge of mill. All the machinery is new and of the most modern build and make.

The mill has been constructed under the charge and supervision of B. S. Garbarini of Jackson, Cal., a man of wide, practical experience as a millwright. He will remain until the mill is in full operation.

Charles Taylor is the manager of the tungsten mines and employs at the present time over 30 men. The principal development so far has been on the Papoose claim, but De Golia and Atkins, the principal owners, have a great number of adjoining claims.

“Concentrate Tungsten Ore,” March 28, 1907, p. 3: Charles Wetherbee and W. W. Wickard have been experimenting on concentrating tungsten ore at the Phoenix mill at Johannesburg with entire success. They have used the Woodbury Concentrator

and find that they can take low grade tungsten ore and concentrate it into values running as high as sixty per cent; values that pay well for shipment. This will be of immense benefit to many holders of tungsten claims that do not produce ore rich enough to ship without concentrating, as is the case with the Papoose.

“Shipment of Tungsten Ore,” April 25, 1907, p. 4: Tomorrow Chas. Wetherbee ships four tons of tungsten ore. This ore comes from several parties. Some of it is from the Winnie mine owned by Charlie Koehn and is in a pure state, just as it comes from the mine. Some of it comes from the Gold Stone mine owned by Wetherbee just west of the Papoose and the balance from Wickard mine. Some of this ore has been concentrated and is of a high degree of purity. The shipment goes to Philadelphia and is paid for at the rate of \$8.00 per unit F. O. B. Johannesburg. The Gold Stone mine is showing up in fine shape. They were down 100 feet and now drifting.

“Mine Sold,” June 13, 1907, p. 4: Charlie Koehn has sold the Winnie mine to Vald Schmidt and J. W. Barter, from Los Angeles. It is now considered more valuable for tungsten [tungsten] than gold. Charlie has taken out \$1,350 worth of tungsten in the past two months, and a leaser took out \$200 in two weeks. The payments are \$3,000 cash and the balance in payments.

Charlie says the sale, although a good one, will not swell his head and he still expects to retain his official position of mayor of Kane Springs, as the new railroad will run through his back yard.

Wm. Greenwood, **“A Letter to the Miner,”** June 20, 1907, p. 2:

Editor of the Miner:

Before the muck rakers start for Atolia I wish to set at rest what promises to grow into another mystery of the great desert if not nipped in time.

It is true as reported, that Mr. Charles Wetherbee and myself shipped a car load of tungsten ore to Colorado last week to be milled, but neither of us have any designs against the cherished [cherished] institutions of our country.

Because there is a mill here that could do what we have sent over a thousand miles a way to have done, the belief is gaining ground that the sending of that car is nothing less than another sinister move in the game for a third term, all of which I

here and now brand as a deliberate and unqualified falsehood.

The home plant should certainly have the first call, and I was always ready to do business with them, but, incredible as it may seem, I just can't get it out of my wool but what our friends are trying to build up a monopoly in our midst.

I never ran to them to have my ore milled; never at any time have I asked them to do it. They do the running. Time and again did they come to me to secure the handling of my output, which I agreed to give them, and they promised over and over again that they would install the ore testing and sampling adjuncts to their mill that custom work calls for, and this was to have been done over three months ago, but they have yet to make the first move towards putting in that equipment. Why they didn't is of course plain enough. It doesn't take the seventh son of seventh wife to smoke the coon out of that wood pile.

But greed makes us all dull or they would have seen that it was impossible for monopoly to get control of this tungsten camp against the will of the independent claim owners. There is no freezing out if the right men are behind the guns, and if I have been always supremely indifferent whether my ore was treated here or not, there have been mighty good reasons for this display of the spirit of old '76 and the Declaration. Among them are these two smashing facts which our embryo monopoly should study. The first is that milling charges are much less in Denver than here, and the other is that market pays from \$100 to \$240 more per ton for the same grade of concentrates. Nor does it take a high grade of ore to make the trip. Rock that is too poor to ship is too poor to mill in this camp, for it should be known that the Papoose people gave us poor devils to understand that they wouldn't do our milling anyway unless we agreed to sell them our concentrates, and at their price.

Sure, greed makes us dull. But it is true now as always that to cook our rabbit you must first catch him.

WM. GREENWOOD.

“Mining Notes,” June 27, 1907, p. 4:

The mining news of Randsburg vicinity is about the same as usual—a constant producing and milling, which long ago has passed the experimental

stage and settled down to the bed rock which has made this the substantial mining camp it is.

This week Wilhite & Jeffarard had a milling of \$1500 worth of tungsten ore from the Santa Ana mine in the Stringer district.

Barney Ostick [Osdick] has just had a milling from the Merced mine of \$800 worth of tungsten.

McCormick & O'Leary have also had a recent milling from the G. B. mine, but the exact amount we did not learn.

Chas. Weatherbee is taking out a considerable quantity of tungsten every day.

Considerable development is also being done over in the tungsten field, and rich strikes may be expected at any time.

C. G. Illingworth has a force of men at work in his mines there, from which he recently took out about \$2000 worth of the ore. He now has a shaft down about 80 feet, with favorable indications.

Wm. Greenwood, H. S. Roberts and others are making extensive developments in the tungsten line.

The Atolia Mining Co. are constantly enlarging their force, and the general mining outlook for that section is very bright.

[**Untitled**], Randsburg Miner, July 18, 1907, p. 3: Mining and Scientific Press: The chief use of tungsten is in the manufacture of tungsten steel, and this again is mostly used for high speed tools, magnet steels, and in European countries also for armor plates and projectiles. The percentage of tungsten in tool steels varies greatly, according to the special uses for which they are intended, but tungsten always imparts to the steel the property of self-hardening in the air after forging without necessitating any of the usual methods of tempering.

“**Criminal Negligence**,” Randsburg Miner, July 18, 1907, p. 3:

From the American Mining Review [Los Angeles] [:]

For the first time in many years a California coroner's jury on July 4th, rendered a verdict of criminal negligence on part of a mining company, making it responsible for the death of a miner killed by a cave. The accident happened in the Bartlett mine of the Wheeler Mining Company, near the village of Dagget[t], in San Bernardino county.

The mine is opened by means of a flat incline which, 100 feet in, takes a steeper pitch. A good-sized excavation had been made there, said to be 12 by 15 feet, and no timber had been used in supporting the back. Every desert miner knows the phenomenal standing qualities of ground in those dry mines, which probably accounts for the absence of timbers, but after blasting, ground in desert mines requires picking and baring down, the same as elsewhere. It seems that it was while engaged in cleaning up after a round of holes that the miner was killed by a fall of rock from the roof.

“Farewell Entertainment At Atolia,” August 8, 1907, p. 4:

To inaugurate the first social function given at Atolia, Mr. and Mrs. William Bouchard entertained with a very charming musicale on Wednesday evening at their residence on Grand Ave., in honor of Miss Lois Price, of Bakersfield. The vivacious hostess, Mrs. Bouchard, was assisted in receiving her numerous guests by Mrs. C. Nebeker and Mrs. W. Atkinson.

Among those bidden welcome to the evening’s amusement were Mrs. Thos. McCarthy, Mrs. C. Taylor, Mrs. R. McEwin, Mr. C. Nebeker, Mr. W. Bouchard, Mr. W. Atkinson and Mr. P. Jensen.

A bevy of pretty debutantes and young girls enlivened the time very much by their graceful appearance and winsome ways. Miss Price certainly is honored in having as her friends and companions the Misses Theresa, Sarah, Marie and Clara McCarthy, Ada, Albertina and Gertrude Schoonmaker, Crecent Nebeker, Beatrice Atkinson, Minnie and Winfried Stanbury, Lena Giawanda and Helen Moore.

The Messieurs [Messrs.] Leo and William McCarthy, John, Edward and William Bartino and Morris Atkinson contributed not a little by their singing and acting a piece from ‘The Merchant of Venice.’

Sandwiches, cake and ice cream were served to the guests on the spacious lawns beneath heaven’s starry dome, and all enjoyed the occasion immensely.

Many were the well-wishes for Miss Lois at the hour of parting, and a hearty welcome was assured her should she again visit our town in the future.”

“Local Notes,” August 29, 1907, p. 4:

[Excerpts:]

It is reported that Thomas McCarthy has struck a good body of tungsten again in his lease from the Papoose at Atolia.

Chas. Weatherbee, of the tungsten mines, has just returned from a long trip by wagon through Owens valley and up to Bodie.

...

There is still a large quantity of pipe to be hauled out to complete the water system for Skidoo. It will take several months yet to get it on the ground, as every team is busy that can be engaged.

“Accidentally Killed at Atolia,” September 5, 1907, p. 4:

On Monday of this week shortly after going to work in the morning, George Anderson, a young man who had been working in the Papoose mine for several months, was instantly killed by the falling up a bucket, weighing perhaps a hundred pounds, striking him on the head.

No one saw the accident, but from appearances, in moving buckets from a lower to a higher level he had tied on bucket to the bottom of another and then gave the signal to blast. He seemed to have followed the buckets up the ladder, and from some cause the lower bucket became detached and falling struck Anderson on the head, breaking his skull and killing him instantly.

His body fell something less than fifty feet, lighting on the platform which covered the shaft at the 250 foot level.

The coroner of San Bernardino county was notified and he came up the same day and held an inquest, with a verdict of accidental death, in accordance with the above circumstantial evidence.

Rev. Amis[?] conducted funeral services Wednesday evening in the miners’ hall, after which the remains were taken to Johannesburg to be shipped to the grief stricken mother at Salinas, Calif.

The Atolia and Stringer miners came in to attend the funeral, after which a parade of 112 Union miners followed the remains to the train at Johannesburg.

“McCarthy’s Tungsten,” September 12, 1907, p. 4:

Thos. McCarthy’s tungsten strike at Atolia is proving better with age, so it is reported. He struck the ledge about 20 feet from the top of the ground and has now followed it down about 20 feet and it is opening up larger the deeper he goes.

“Local Notes,” September 26, 1907, p. 3:

[Excerpts:]

Thomas McCarthy, the Randsburg Tungsten miner, was in Bakersfield last Sunday. Mr. McCarthy recently made a very rich strike on some of his tungsten property and is working a good force of men getting out the valuable mineral.—Echo.

...

S. P. MacKnight and H. R. Bacon, a mining man from Los Angeles, have taken a lease and bond on the Tungsten Mines owned by Wickard, Byrne and others and have set some men at work. There is a strong probability of selling this property outright, soon as some interested parties are expected in next week.

“Local Notices,” October 17, 1907, p. 4:

[Excerpt:]

Mr. Curtiss, in the employ of C. G. Illingworth, took a wild ride in the rain Tuesday night, not down on the program of a groceryman delivering goods. In the first place our enterprising merchant, Illingworth, owns good stock and has two especially fine teams. Tuesday evening Mr. Curtiss drove one team to Atolia, five miles out, and while delivering the team got away from him, starting out at a 2:40 gait for Kramer. Curtiss procured a saddle horse and followed after in a wild John Gilpin ride. Soon the rain began, he had been thoughtful enough to take a rubber coat but it was in the wagon, so he had to take the rain. He followed until within a couple of miles of Kramer and there met the team coming back. They had the wagon all right, the rubber coat and other things, and had lost nothing but a sack of barley. Mr. Curtiss was dripping wet and cold, but he got back with the team about 5 o'clock Wednesday morning. The team was uninjured, but they looked pretty serious, and so did Curtiss.

“More Tungsten Mines,” November 7, 1907, p. 1:

Another rich find of tungsten in San Bernardino county is reported by the Long Beach correspondent of the Los Angeles Express, who says:

‘Carl Firhard and Peter Harrison, two local mining men, claim to have struck a veritable bonanza in the form of a tungsten mine on their claims on the Mojave desert.

‘They returned to Long Beach to complete arrangements for testing the value and extent of the deposit, which they believe will make them immensely wealthy.

‘Tungsten is an extremely rare metal used in toughening steel for armor-plate and similar purposes and is worth \$15 a pound.

‘Both gentlemen claim that preliminary examinations and tests of the deposit showed 75 per cent tungsten, and that there is a great deposit of it, from all indications. The metal was struck at a depth of but 15 feet.’

“The Value of Tungsten,” November 14, 1907, p. 2:

There is always something fascinating about minerals, especially those that command a high price in the market. Among these there is no doubt that tungsten is the most interesting, because it is one of the youngest minerals known to commerce. Already it has been the victim of many a fake story and perverted report. The fabricator being in little danger of contradiction, feels free to give his imagination full rein. Only a short time ago an otherwise reputable mining journal published an article concerning this new metal, in course of which the statement was made that tungsten was now being used for making frogs, especially for street car tracks where the switch had to be thrown very frequently, and that a frog of tungsten would outwear ten frogs of ordinary steel.

In truth a frog of tungsten alloy—it is never employed for any other purpose than alloy—would be so brittle that it would go to pieces in a few hours. The most common error concerning the use of tungsten is that it is employed in the making of armor plate and heavy projectiles. It is not difficult to trace this mistake to its source. Tungsten alloy is used in the manufacture of tools for cutting steel, especially such tools as are run at high speed. The later work on the steel jackets of projects and the paring and trimming of the sections of armor plate for battleships is all done with tools of tungsten alloy. Some such statement as this was probably read and misconstrued by the public.

The peculiar black minerals was first used in laboratory experiments by Robert Muchet, a German chemist, who knew absolutely nothing of its peculiar properties. He had made a variety [variety] of alloys and was comparing results when he observed that some of the blocks of steel, which he had neglected to plunge into water, had become very hard. The idea of hardening steel without water was an entirely new one, and the chemist pursued his line of investigation until he succeeded in making a much better grade of steel than any

that had hitherto been known.

That was in 1872 and it was not until 1898 that a practical use for the tungsten alloy was found. Tungsten was to be had in such small quantities that its sphere of activity was practically limited to the laboratory.

Before the discovery of the properties of tungsten, there was a definite limit to the rate at which tools for cutting and sawing steel could be run, because of the heat developed at the point of contact. The old-fashioned chisel, even when operated slowly, was worthless after two large projectiles had been finished. The edge was gone entirely. The chisel made of tungsten alloys works better when it is operated at such a speed as to develop enormous heat than it does when operated slowly. Whereas the ordinary tool cuts a steel chip at the rate of twenty-five lineal feet a minute, the tungsten tool has been known to cut a chip two hundred feet long in the same time, without injury to the tool. With this basis of comparison it may readily be seen that although the common alloys costs two cents a pound and the tungsten alloy costs sixty-five cents a pound, the latter is the cheaper in the long run.

Wood working tools are also now made of air-hardened tungsten steel, the metal so hard that it cannot be cut but must be shaped by grinding on the emery wheel. In some instances it is so brittle that it must be re-heated to prevent it from crumbling. Because of this brittleness it is easy to perceive that tungsten alloy would be worthless for armor plate or railroad frogs. It is of inestimable value to the machinist, the locomotive manufacturer and the workers in refractory metals and wood. Scientific men are now at work seeking other uses to which it may be put.—L. A. Herald Weekly Magazine.

Epilogue

During the next few years, Atolia grew into a well-developed camp. In February, 1908, the San Bernardino County Board of Supervisors board created the Atolia School District. The mining company gradually expanded its operations. When World War I began, in the summer of 1914, Atolia boomed. It came to enjoy water and electrical service and a wide variety of businesses, including two short-lived newspapers. The end of the war and the influx of tungsten from a newly discovered deposit in China nearly killed the town. The post office closed on August 15, 1922.

After the United States Congress raised the tariffs on tungsten, a long revival followed. The Atolia Mining Company reopened its mines and began leasing out other properties. The post office was re-established on November 10, 1927. Mining remained active throughout the Depression and increased during the early years of World War II. On July 31, 1944, soon after the shutdown of the mines, the post office closed for the last time.

Thermal infrared airborne hyperspectral detection of ammonia venting on the Calipatria fault in the Salton Sea Geothermal Field, Imperial County

David K. Lynch, David M. Tratt, Kerry N. Buckland, Jeffrey L. Hall, Brian P. Kasper, Michael G. Martino, Luis J. Ortega, Karl Westberg, Stephen J. Young, and Patrick D. Johnson

The Aerospace Corporation, 2310 E. El Segundo Blvd., P.O. Box 92957, Los Angeles, CA 90009-2957. dave@caltech.edu

Abstract

An airborne hyperspectral imaging survey was conducted along the Calipatria Fault in the vicinity of the Salton Sea in Southern California. In addition to strong thermal hotspots associated with active fumaroles along the fault, a number of discrete and distributed sources of ammonia were detected. Mullet Island, some recently exposed areas of sea floor, and a shallow-water fumarolic geothermal vent all indicated ammonia emissions, presumed to originate from the eutrophic reduction of nitrate fertilizer in agricultural runoff and the decay (oxidation) of organic matter, probably algae. All emission sources detected lay along the putative Calipatria Fault, one of a number of en echelon faults in the Brawley Seismic Zone that is part of the northern-most spreading center of the East Pacific Rise. The techniques developed during this field experiment suggest a potential methodology for monitoring certain of the toxic episodes that are a known source of mass aquatic fauna kills within the Salton Sea ecosystem. The imagery was acquired at $\sim 0.05 \mu\text{m}$ spectral resolution across the $7.6\text{--}13.5 \mu\text{m}$ thermal-infrared spectral region with a ground sample distance of approximately 1 m using the SEBASS (Spatially Enhanced Broadband Array Spectrograph System) sensor.

Introduction

Hyperspectral imaging, consisting of moderate-spectral resolution ($\Delta\lambda = 100 - 1000$), is an especially powerful diagnostic tool for environmental remote sensing. Hyperspectral imagery is in the form of "data cubes", two spatial dimensions and one spectral dimension (Figure 1). Airborne hyperspectral imagers such as AVIRIS (Green et al., 1988), SEBASS (Hackwell et al., 1996) and Hyperion (Pearlman et al., 2003) have demonstrated their value in such areas as geology, biology, agriculture, surveillance and hydrology, principally because of their ability to remotely identify airborne molecules and determine surface composition.

Virtually all molecules have vibrational and rotational spectral features that are well-resolved by hyperspectral techniques. Thus hyperspectral imagery can be used for exploration by identifying a wide variety of molecular species with digital filters applied to hyperspectral images

that show the location and concentrations of the molecules (Figure 2).

The work reported here was conducted with the SEBASS (Spatially Enhanced Broadband Array Spectrograph System; Hackwell et al., 1996) airborne hyperspectral imager. SEBASS data cubes are first corrected for effects of the intervening atmosphere using the Inscene Atmospheric Compensation process (Young et al., 2002) and rendered into time-stamped, georeferenced brightness temperature maps before being scanned for the presence of gaseous emissions. Molecular emissions are then detected and identified using principal-component projection and spectral matched filtering (Young, 2002).

Crustal thinning from the divergent plate boundary and associated upward magma migration within the Salton Trough have resulted in local hotspots whose surface expression is characterized by low-grade volcanic features

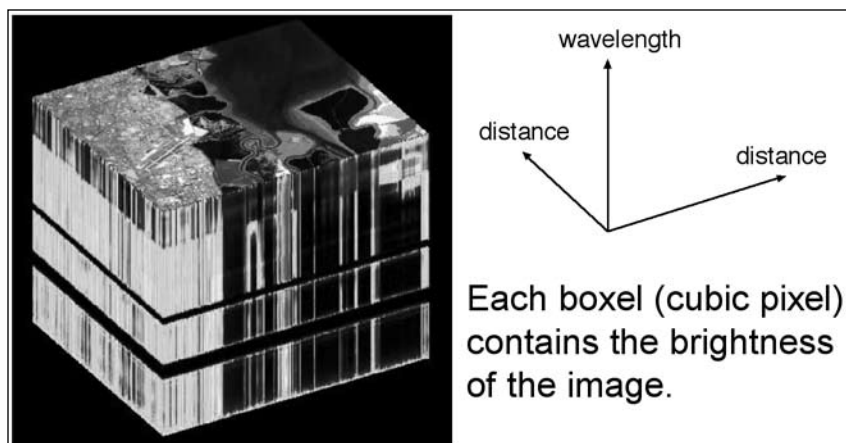


Figure 1. Hyperspectral imagery in the form of "data cubes."

such as fumaroles and mud pots. The Salton Sea lies at the bottom of the trough (Kaiser 1999; Holdren and Montaña 2002; Barnum et al. 2002; Hulbert 2008). It was created accidentally in 1905 by a levee failure (Larkin 1907; Kennan 1917), though it has naturally existed in historic times as ephemeral fresh water Lake Cahuilla. The Mullet Island/Calipatria Fault complex is associated with the Mullet Island Thermal Anomaly (Sturz, 1989) and was chosen as a principal target for a March 2009 collection campaign, having been previously identified and mapped in detail (Sturz, 1989; Svensen et al., 2007, 2009; Lynch and Hudnut, 2008). Mullet Island is one of five rhyolitic volcanic necks thought to define the axis of a small spreading center.

In this paper we report SEBASS imagery of one such target—the Calipatria fault (CP)—and the detection of gaseous ammonia (NH_3) emission from Mullet Island and from a fumarolic vent, both currently in the Salton Sea (Figure 3).

Observations

The Aerospace Corporation's hyperspectral sensor SEBASS (Spatially Enhanced Broadband Array Spectrograph System) operates in the thermal infrared part of the

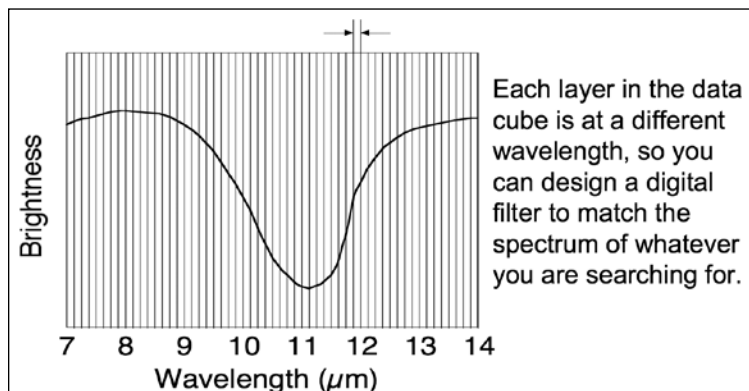


Figure 2. Hyperspectral image showing location and concentrations of molecules.

spectrum (7.6 to 13.5 μm , Hackwell et al., 1996; Figure 4). For the last few years, it has been used for mineralogical (Kirkland et al., 2002; Hackwell et al., 2008), agricultural and urban studies including the World Trade Center collapse (Hall 2009). SEBASS is calibrated using ground measurements and in-scene methods to produce radiometrically-calibrated imagery, i.e. scene radiance in $\text{W cm}^{-2} \mu\text{m}^{-1}$ (or equivalently brightness temperature derived by inverting the Planck function) for each boxel. In March 2009 SEBASS was flown over a variety of

southern California targets.

On March 26, 2009 (19:43–20:21 UTC) the CP was overflowed using SEBASS mounted in a deHavilland Twin Otter. From an elevation above ground level (AGL) of 3 km, the pixel size on the ground was ~ 1 m and the image swath width was 128 pixels (~ 128 m).

Figure 5 shows part of the flight swath recorded by SEBASS that has been color coded using the 8.8, 11.7 and 12.7 μm infrared channels. The background picture is a true color optical image acquired by the Quickbird satellite on 1 May 2005. A fumarolic vent (circle, center) and Mullet Island (upper left) are evident in the picture. Figure 6 shows the vent in 2007.

Figure 7a shows a close-up broadband thermal image (8–13 μm) of the vent and figure 7b shows the results of the ammonia image obtained of the same area processed with the digital ammonia filter. Our procedure entails applying filters for several hundred compounds, but NH_3 was the only molecule identified in the imagery that met the detection criteria of the filter system.

In these images, dark represents ammonia seen in absorption against the colder background. Light reveals thermal emission from ammonia gas that is hotter than the background. Ammonia is seen drifting to the NE, driven by local winds (7b), while warm water is seen drifting N and W by currents in the sea (7a). Figure 7a also reveals a number of smaller, hot CO_2 gas seeps in the sea east of the main vent.

Figure 8 shows a brightness temperature image of the vent area. Temperatures as high as 76.7 C were measured remotely using SEBASS. A visit to the fumarole on a US Fish and Wildlife Service boat revealed that the emitted gas and water were boiling (>100 C), and that the fumarole itself was covered by a few cm of water (Figure 6). The lower temperatures retrieved

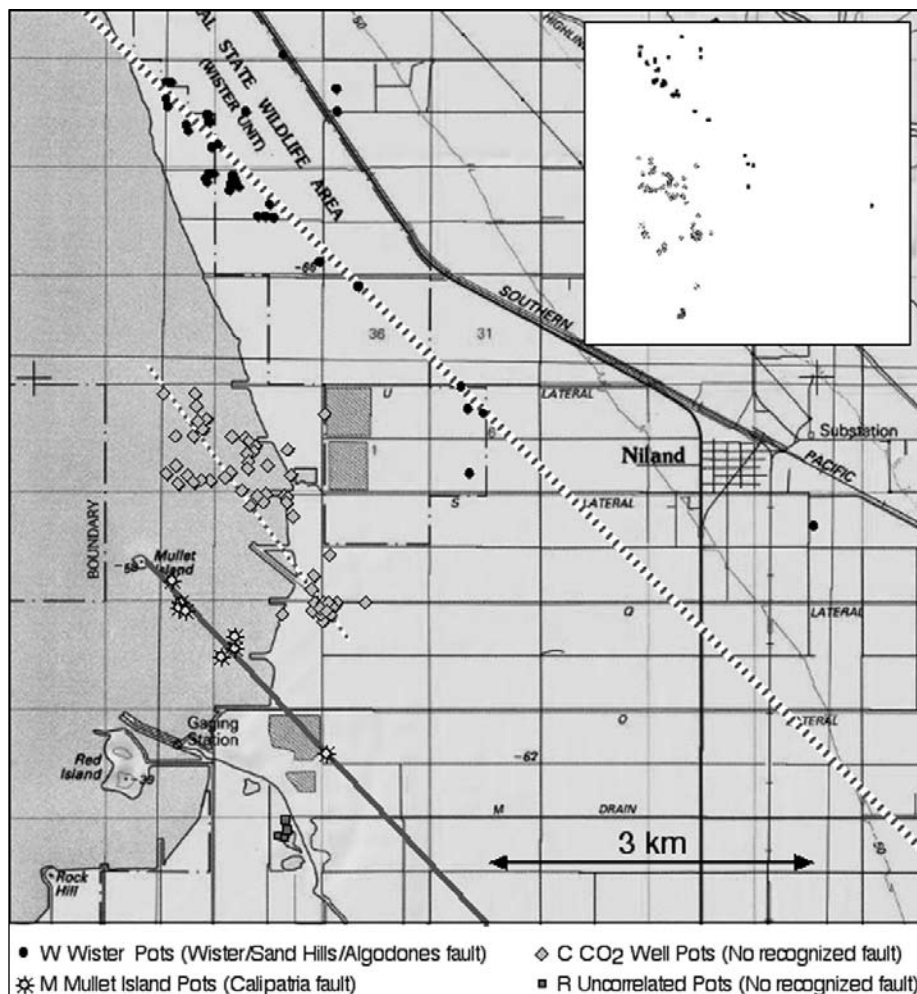
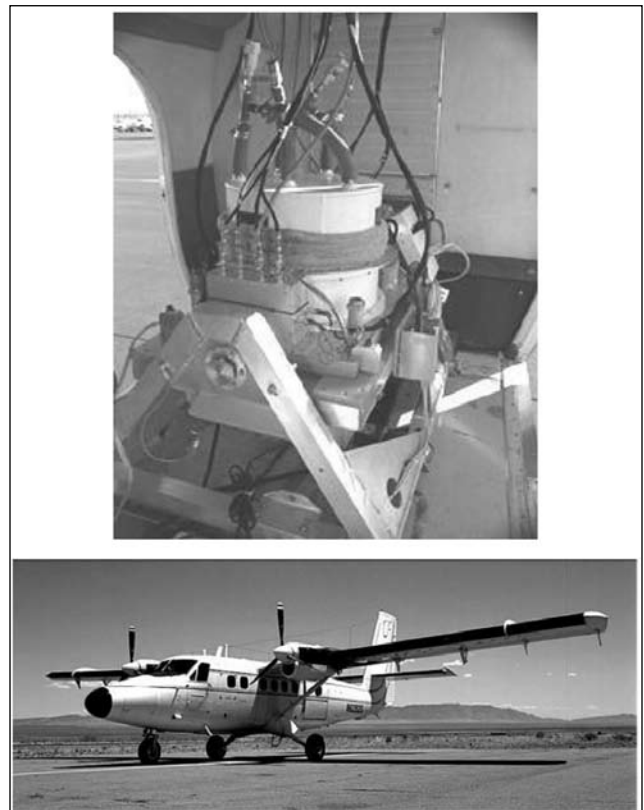


Figure 3. The Calipatria fault (CP) and the detection of gaseous ammonia (NH₃) emission from Mullet Island and from a fumarolic vent, both currently in the Salton Sea.

from the hyperspectral imagery are probably due to two effects: evaporative cooling of the gas, and 2) signal averaging of different temperature zones within the ~1 meter square pixel.

Like Figures 7a and 7b, Figures 9a and 9b show thermal radiance and ammonia filter images of Mullet Island. The sources of ammonia are distributed over the island. Spatial variations in surface emission rates and turbulent/convective mixing probably serve to render ammonia detectable in both emission (bright) and absorption (dark). We are currently modeling an ammonia-air mixture to attempt retrieval of the ammonia column density.

Figure 4. The Aerospace Corporation's hyperspectral sensor SEBASS (Spatially Enhanced Broadband Array Spectrograph System) operates in the thermal infrared part of the spectrum.



Discussion

The Calipatria fault (CP) is one of many in a complex region of en echelon and cross faults in the northern-most spreading center of the East Pacific Rise. It is in the Brawley Seismic Zone (BSZ) which contains the Salton Sea Geothermal Field (SSGF, Younker, Kasameyer and Tewhey 1982). The region represents a spatial transition in crustal dynamics between a transform fault (San Andreas fault to the north) and a divergent plate boundary (spreading center) to the south. Owing to agricultural modification of the surface, the CP has little surface manifestation and its trace is not well located. Its existence was originally suggested by Kelley and Soske (1936) who noted that mud volcanoes near the corner of Davis and Schimpf roads and those that were a popular tourist destinations in the 1920s (Lafin, 1995) - before

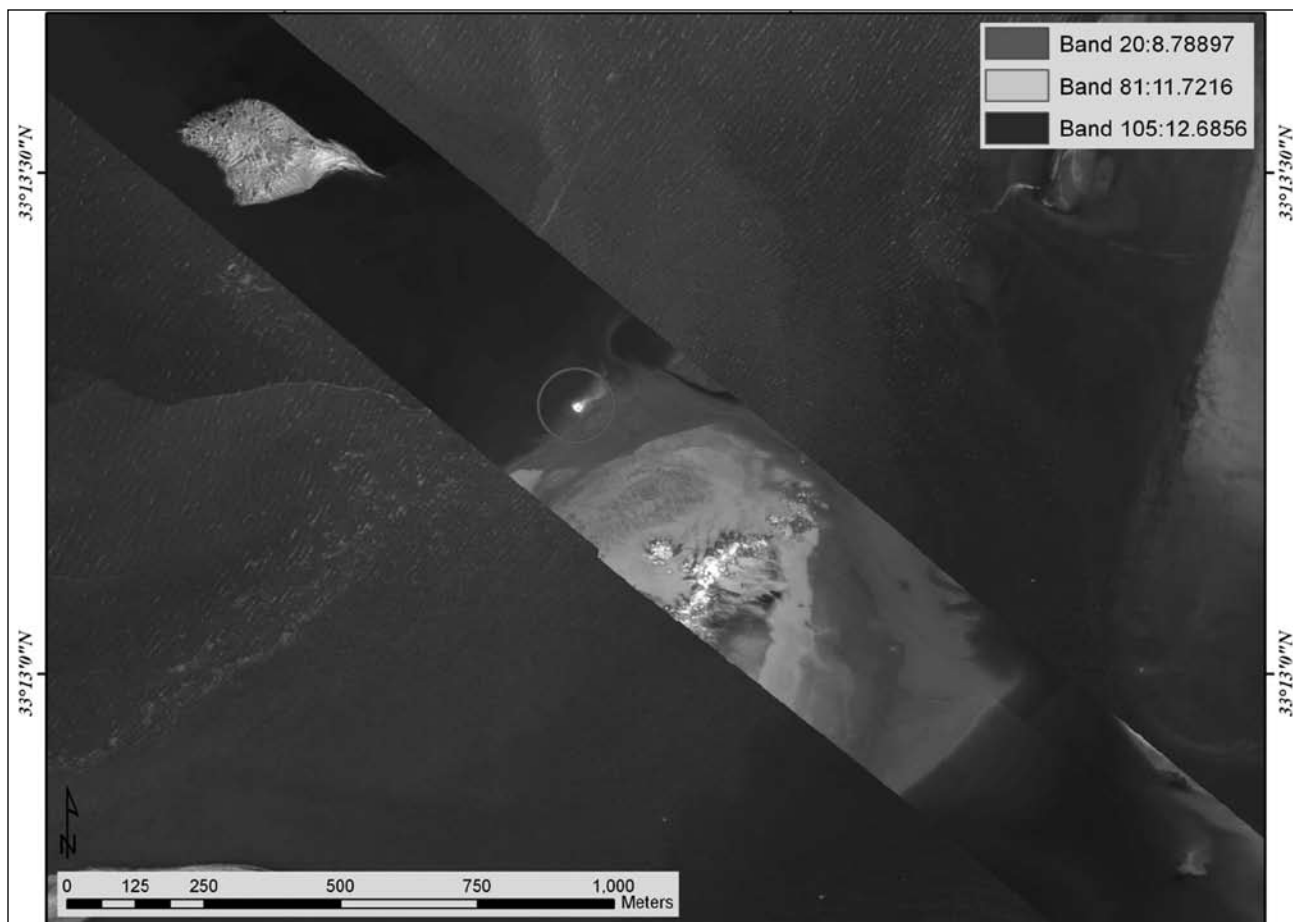


Figure 5. Part of the flight swath recorded by SEBASS. The background picture is a true color optical image acquired by the Quickbird satellite on 1 May 2005. A fumarolic vent (circle, center) and Mullet Island (upper left) are evident in the picture.

being submerged as a result of a rise in the Salton Sea - fell along a line striking more or less N45°W that passed through Mullet Island. The hot vent observed here is probably the former tourist attraction.

In 1972, Meidav and Furgerson suggested that this line might indicate the location of a fault that they called the Calipatria fault. According to Schroeder (1976), Meidav and Furgerson posited several other northwest-trending faults in the area that they called (from north to south) the Wister, Calipatria, Red Hill, Brawley, Fondo, and Westmorland faults. Their map suggests that the fault locations are only conceptual because all six indicated faults are exactly parallel to each other and strike N45°W. Based on electrical resistivity, magnetic and gravimetric



Figure 6. The fumarolic vent (Fig. 5) in 2007.

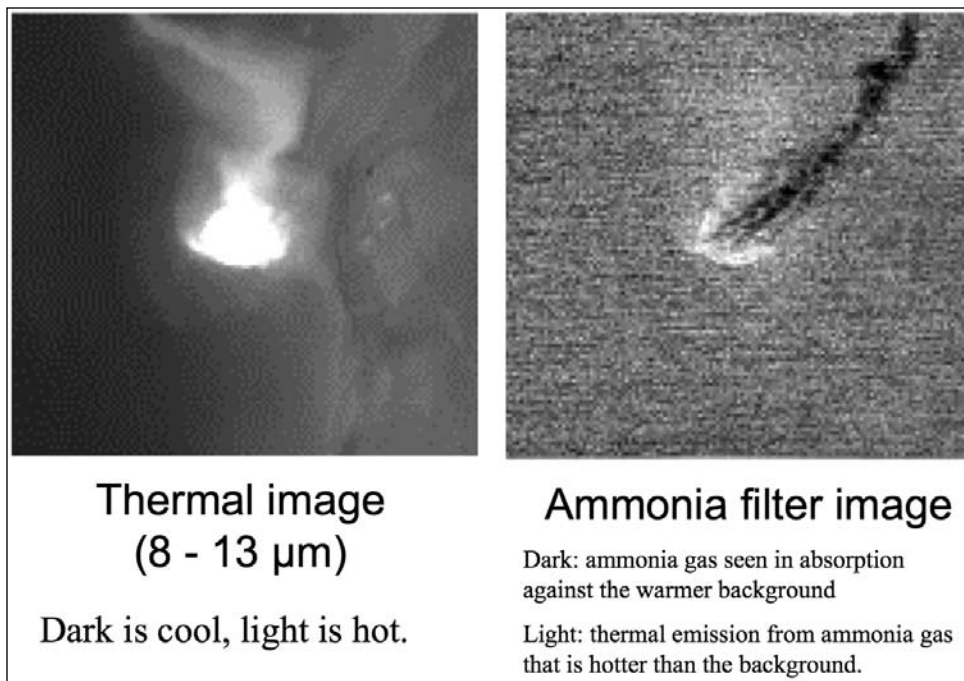


Figure 7a. Close-up broadband thermal image (8-13 μm) of the vent. Figure 7b. Results of the ammonia image obtained of the same area processed with the digital ammonia filter.

studies (Kelley and Soske 1936; Meidav and Furgerson 1972), and on the obvious lineament of geothermal features, the CF is thought to extend on a NW strike (roughly) from SE of the well-known mud volcanoes to Mullet Island and beyond an unknown distance.

In the hot vent the ammonia from the sea bed is probably entrained by upwelling CO_2 in the water and

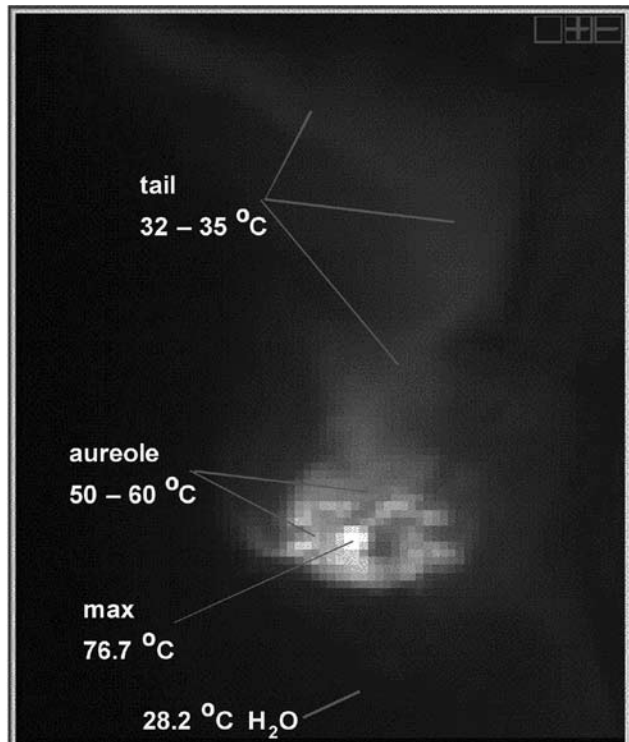


Figure 8. Brightness temperature image of the vent area.

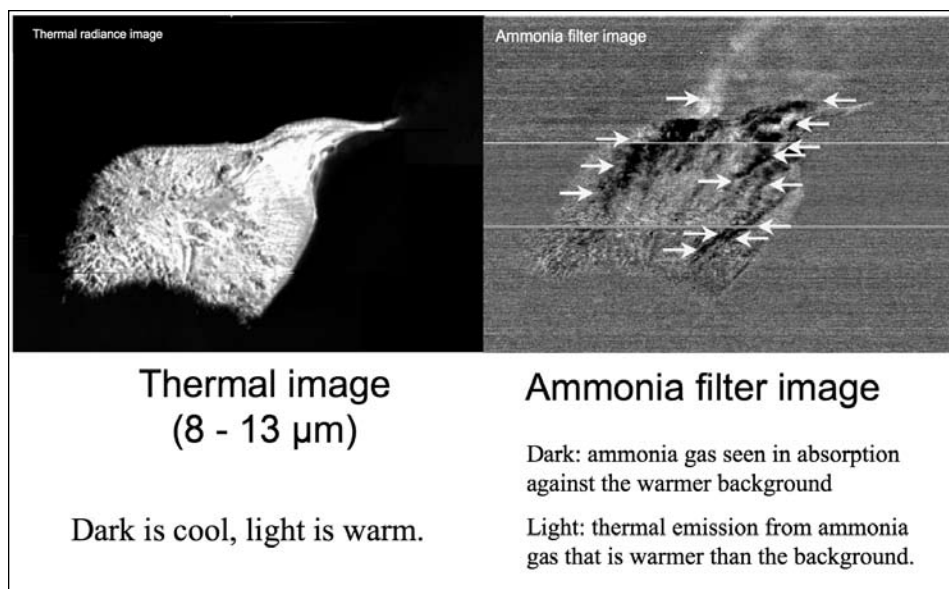
carried aloft. CO_2 is the main vent gas (>95%). As a penetrating surface rupture, the CP subjects the Salton Sea sediments to local disruption. It seems likely that ammonia rising from Mullet Island is from decomposing bird guano. Guano consists of ammonia, uric, phosphoric, oxalic, and carbonic acids, as well as some earth salts and impurities.

The ammonia originates in two main ways: 1) eutrophic reduction of nitrate fertilizer in agricultural and livestock runoff and 2) the decay (oxidation) of organic matter. Both processes are involved with the nitrogen cycle in water,

in which atmospheric nitrogen (N_2), ammonia (NH_3), ammonium ion (NH_4^+), nitrite ion (NO_2^-), and nitrate ion (NO_3^-) undergo various chemical transformations, some through the agency of biologic organisms. There is a large nitrogen and phosphate component to fertilizers used in the Salton Trough, and much of it drains into the Salton Sea as agricultural runoff or by seeping into the water table of the area. Nutrient levels rise and algae and microbes grow prodigiously, extracting dissolved oxygen (DO) from the water. Foraminifera are the most abundant microorganisms in the sea, and the average productivity of the Salton Sea is greater than that of the oceans.

Ammonia concentrations in the Salton Sea rose by about a factor of five between 1969 and 1999 (Setmire et al., 2001), and are expected to continue rising. Broadly speaking, ammonia levels in 1999 were between 0.22 and 2.5 mg/L, far in excess of the EPA standards (EPA 1989) of 0.035 mg/L, though the exact value of the standard depends on water temperature, salinity and pH.

In the Salton Sea, there have been occasional mass fish kills, primarily Tilapia (*Oreochromis mossambicus*). According to Setmire (2000) "Major shifts in the direction and velocity in the southern end appear to be associated with fish kills. Strong winds from the south agitate the anaerobic sediments creating an immediate oxygen demand in an area that already has [dissolved oxygen] DO depleted water in the lower part of the water column. As this water is turned over and mixed with the upper water column, the entire water column of these shallow water



Figures 9a and 9b. Thermal radiance and ammonia filter images of Mullet Island.

areas is depleted of DO, killing the fish. As prevailing winds reestablish, windrows of dead fish often are found off of the New and Alamo River deltas. Although fish kills have been common at the Salton Sea for a long time, during the past decade die-offs are more numerous and many of the die-offs number several million fish. Tilapia, the most abundant fish in the Salton Sea, also is the species most often involved in kills.”

Summary and conclusions

Demonstrated here is the ability of thermal infrared hyperspectral imagery to detect relatively small amounts of gaseous ammonia in the atmosphere; in this experiment ammonia was detected that was emanating from a fumarole and Mullet Island, both in the Salton Sea. The techniques developed during this field experiment suggest a potential methodology for remote or stand-off detection of gases, and the technique shows the viability of monitoring certain biologic events such as toxic episodes that are a known source of mass aquatic fauna kills within the Salton Sea ecosystem.

Acknowledgments

This work reported here was supported in part by The Aerospace Corporation’s Independent Research and Development program. D. K. Lynch acknowledges support from Southern California Earthquake Center (SCEC) research award 09084.

References

- Barnum, J. F. Elder, D. Stephens & M Friend (eds) 2002 in “The Salton Sea”, *Developments in Hydrobiology*, *Hydrobiologia* 473, 1-21, Kluwer
- EPA 1989 Ambient Water Quality Criteria for Ammonia (Saltwater), EPA 440/5-88-004 <http://www.epa.gov/waterscience/criteria/ammonia/>
- Green, R.O., M.L. Eastwood, C.M. Sarture, T.G. Chrien, M. Aronsson, B.J. Chippendale, J.A. Faust, B.E. Pavri, C.J. Chovit, M. Solis, M.R. Olah, and O. Williams (1998), “Imaging Spectroscopy and the Airborne Visible/Infrared Imaging Spectrometer (AVIRIS),” *Remote Sensing of Environment*, 65, 227–248
- Hackwell, J.A., D.W. Warren, R.P. Bongiovio, S.J. Hansel, T.L. Hayhurst, D.J. Mabry, M.G. Sivjee, and J.W. Skinner (1996), “LWIR/MWIR imaging hyperspectral sensor for airborne and ground-based remote sensing,” *Proc. SPIE*, 2819, 102-107
- Hall, J. 2009 “Mineral and Gas Identification using High-Performance Thermal Infrared Imaging Spectrometers” Naval Postgraduate School Thermal Imaging Workshop 14 Sept 2009, Monterey, California
- Hall, J.L., J.A. Hackwell, D.M. Tratt, D.W. Warren, and S.J. Young (2008), “Space-based mineral and gas identification using a high-performance thermal infrared imaging spectrometer,” *Proceedings of SPIE* 7082, 70820M
- Holdren, G.C., and A. Montaño 2002 “Chemical and physical characteristics of the Salton Sea, California,” *Hydrobiologia* 473(1-3), 1-21
- Hulbert, S. H. (ed.) 2008 *The Salton Sea Centennial Symposium Developments in Hydrobiology* 201 *Hydrobiologia* 604, Springer (2008) Dordrecht
- Kaiser, J. 1999 “Battle over a dying sea” *Science* 284:28-30
- Kelley, V. C., and J. L. Soske 1936 “Origin of the Salton volcanic domes, Salton Sea, California” *J. Geol.* 44(4) 496–509
- Kennan, G. 1917 *The Salton Sea: An Account of Harriman’s Fight With The Colorado River* The Macmillan Company, New York
- Kirkland, L. E., K. C. Herr, E. R. Keim, P. M. Adams, J. W. Salisbury, J. A. Hackwell, and A. Treiman 2002 “First use of an airborne thermal infrared hyperspectral scanner for compositional mapping” *Remote Sensing of Environment* 80, 447-459

- Kirkland, L. E., K. C. Herr, and J. W. Salisbury 2001 "Thermal Infrared Spectral Band Detection Limits for Unidentified Surface Materials" *Applied Optics*, 40, 4852-4862
- Lafin, P. (1995). "The Salton Sea: California's Overlooked Treasure", *The Periscope*, Coachella Valley Historical Society, Indio, California, 61 pp. (reprinted in 1999): available at <http://www.sci.sdsu.edu/salton/PeriscopeSaltonSeaCh7%29.html> (last accessed January 2010)
- Larkin, Edgar L. (March 1907). "A Thousand Men Against A River: The Engineering Victory Over The Colorado River And The Salton Sea", *The World's Work: A History of Our Time XIII*: 8606-8610 <http://books.google.com/books?id=3IfNAAAAMAAJ&pg=PA8606#v=onepage&q=&f=false>
- Lynch, D.K. and K.W. Hudnut 2008 "The Wister Mud Pot Lineament: Southeastward extension or abandoned strand of the San Andreas Fault?" *Bulletin of the Seismological Society of America*, 98(4), 1720-1729
- Meidav, T., and R. Furgerson (1972). Resistivity studies of the Imperial Valley geothermal area, California, *Geothermics* 1, 47-62
- Pearlman, J.S., P.S. Barry, C.C. Segal, J. Shepanski, D. Beiso, and S.L. Carman (2003), "Hyperion, a Space-Based Imaging Spectrometer," *IEEE Transactions on Geoscience and Remote Sensing*, 41(6), 1160-1173
- Schroeder, R. 1976 Reservoir engineering report for the MAGMASDG&E geothermal experimental site near the Salton Sea, CA, Report UCRL-52094, Lawrence Livermore Laboratory
- Setmire, J. G., Holdren, C., Robertson, D., Amrhein, C., Elder, J., Schroeder, R., Schladow, G., McKellar, H., and Gersberg, R. 2001 "Eutrophic conditions at the Salton Sea". A topical paper from the eutrophication workshop convened at the University of California at Riverside. September 7-8, 2000
- Sturz, A. (1989), "Low-temperature hydrothermal alteration in near-surface sediments, Salton Sea geothermal area," *J. Geophys. Res.*, 94(B4), 4015-4024
- Svensen, H., D.A. Karlsen, A. Sturz, K. Backer-Owe, D.A. Banks, and S. Planke 2007 "Processes controlling water and hydrocarbon composition in seeps from the Salton Sea geothermal system, California, USA," *Geology*, 35(1), 85-88
- Svensen, H., Ø. Hammer, A. Mazzini, N. Onderdonk, S. Polteau, S. Planke, and Y. Y. Podladchikov 2009 "Dynamics of hydrothermal seeps from the Salton Sea geothermal system (California, USA) constrained by temperature monitoring and time series analysis", *J. Geophys. Res.*, 114, B09201
- Young, S.J. 2002 "Detection and Quantification of Gases in Industrial-Stack Plumes Using Thermal Infrared Hyperspectral Imaging", Aerospace Report No. ATR-2002(8407)-1, The Aerospace Corporation, El Segundo, Calif.
- Young, S.J., B.R. Johnson, and J.A. Hackwell (2002), "An in-scene method for atmospheric compensation of thermal hyperspectral data," *Journal of Geophysical Research*, 107(D24), 4774
- Yunker, L. W., P.W. Kasameyer and J. D. Tewhey 1982 "Geological, geophysical, and thermal characteristics of the Salton Sea Geothermal Field, California", *J. Volcanology and Geothermal Res.* 12, 221-258

Abstracts of proceedings: the 2010 Desert Symposium

Robert E. Reynolds, compiler

Desert preservation: when is enough enough?

Richard Crowe

Beaumont, CA

Over the 29 years I worked for the Bureau of Land Management (BLM) in the California Desert (Desert) many environmental groups—the Sierra Club and their ilk—successfully eliminated many perceived inappropriate Desert uses via Congressional preservation advocacy. I believe the aggregate of such advocacies, and the resulting Congressional actions, to be arrogant and bad public policy – i.e., land-grabbing. Many public lands users were unfairly and unnecessarily hurt. Certainly science-based management was needed 40 years ago, but the cumulative effects of years of Congressional designations, BLM management actions, law suits, and species listings have all but decimated most desert uses (livestock grazing, mining in particular) and gutted BLM’s multiple use management mandate. The economic values lost are sorely missed in these tough times. Today, we read about green-energy proposals and the environmentalists are in opposition again with another proposed restrictions Congressional bill. How much protection is enough?

The last 40 years in the Desert has produced a continuous and monumental (pun intended) clash among a wide variety of public interests as BLM worked—and reworked—resource allocations and restrictions as required by its Congressional multiple use management mandate (i.e., to provide for a variety of public uses and conservation measures guided by collaboration among public interests and science-based planning). The aggregate of environmental factions subscribes to a sort of environmental religion of naturalness and views many public uses as evil desecrations, not to be tolerated to any degree, regardless of consistency with said Congressional mandate. This is not to say that I am an advocate of energy farms or any other particular use, but I am an advocate of multiple uses and fairness on the “playground”. There should be enough land to go around.

Perspective is often lost in discussion so I offer the following “Fourths” analysis, the geographic starting point of which is the 1976 Congressionally designated California Desert Conservation Area (CDCA), 25 million acres in size—essentially the entire Desert less Owens Valley:

1. By 1930 about 25% of the Desert had become private land, most of it concentrated in Antelope, Apple, Coachella, Mojave River, Palo Verde, and Imperial valleys. On the remaining 75%, federal public lands, there was little to no management or use restriction, and BLM did not even exist.
2. Between 1930 and 1945 several national monuments and military reserves were created, along with Anza-Borrego State Park. Together they covered another 25% of the Desert and had significant preservation/restrictive effects on other public uses. BLM came into existence in 1946.
3. In 1994 the California Desert Protection Act (CDPA) designated millions of acres of national parks and BLM wilderness areas, affecting public access to a third 25% of Desert. Unfortunately, the focus of the CDPA was primarily on parks and higher elevations of the Desert and did little or nothing to address species conservation issues which lie primarily in lower elevations.
4. By 2007, the remaining 25% is largely in BLM multiple use management, but about half of this “portion of the pie” is dedicated to sensitive species/habitats, most notably the desert tortoise. Disturbances on desert tortoise habitats are severely restricted to 1% of the total area of dedicated habitat.

Today, 2010, the remaining 25% is under further restrictions stress. The combination of the still-churning environmental advocacy combined with new green-energy proposals has produced yet another Congressional bill. It would carve the following out BLM lands: nearly 1,000,000 acres of BLM national monument, 250,000 acres of BLM wilderness, 74,000 acres of national parks add-ons, and several miles of wild and scenic rivers. After this bill passes, the remaining CDCA “Fourth” will be reduced to 15% of classical multiple use management—for the variety of public uses that are not allowed on military, park, and wilderness lands.

Back to the introductory question: will there ever be enough land in preservation management? As long as intensive use projects are proposed—probably not.

Where does the BLM 1980 CDCA Plan for managing resources and public uses fit in to this story? The CDCA Plan was an exhaustive government-public effort to con-

serve the full array of sensitive natural and cultural values and allow for current and future human uses such as cattle and sheep grazing, mining, utilities, and recreation – grew out of the motorcycle-use controversies of the 1960s and 1970s. The Plan has been amended many times, especially to keep current with species and habitats issues. Today, a good many of the originally allowed uses are gone or severely restricted, some through amendments that considered better data and science and needs of species listed under the Federal Endangered Species Act. However, most use restrictions grew from Congressional designations and lawsuits, political actions (end runs) to void the difficult, King Solomon administrative decisions contained in BLM's CDCA Plan. Political action has emasculated science-collaborative land management.

One can even argue that the remaining 15% is still under severe management difficulty:

1. **Buffers.** Environmentalists do not like looking out from park/wilderness areas and seeing mines, utilities, and other disagreeable uses on BLM lands (e.g., the proposed Eagle Mountain Landfill near Desert Center, sited adjacent to Joshua Tree NP). They advocate that such uses be buffered away from boundaries. BLM does not subscribe to “buffer” management. Given the large array of parks and wilderness areas the “edge effect” would be enormous and could effectively double the size of parks and wilderness areas.
2. **Acquired Private Lands.** The BLM CDCA Plan defines and allocates management of resources and uses on the basis of large management zones. The recent Kennedy v Feinstein article described the The Wildlands Conservancy demand that BLM manage acquired former Catellus (private) lands to their preservation liking. As conservation groups are often instrumental in facilitating such acquisitions, they have a management expectation and agenda for their effort. However, from a practical point of view acquired lands would best be managed according to the management zone into which they fall: i.e., in wilderness areas they are wilderness; in dedicated desert tortoise areas they are so dedicated; and in other managed use zones the particular set of management prescriptions would apply. Wildlands' demand creates nearly unworkable special “postage stamp” management areas inside these large management zones.

About the 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.

Geochronology and paleoenvironment of pluvial Harper Lake, Mojave Desert, California

Anna L. Garcia¹, Jeffrey R. Knott¹, Jordon Bright², Shannon Mahan³

¹CSU Fullerton, Geological Sciences, P.O. Box 6850, Fullerton, CA 92834

²Northern Arizona University, Department of Geology, Flagstaff, AZ 86011

³USGS, Box 25046, MS 974, Denver, Co 80225

The Mojave River is a well-known recorder of Southern California paleoclimate with a complex paleohydrology and past terminations in the pluvial (upstream to downstream) Harper (Harper basin), Manix (Afton, Coyote and Troy basins), and Mojave (Soda and Silver basins) lakes over the last 40,000 years. Previous studies at Harper Lake 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), stratigraphy section as well as radiocarbon and optically stimulated luminescence (OSL) ages from the Harper Lake basin. The 2.1-m-thick continuous stratigraphic section is near the highstand elevation, rests nonconformably on quartz monzonite, and is comprised of interbedded sand, silt and silty sand capped by a 0.6-m-thick sequence of carbonate mud. Lacustrine sediments contain *Anodonta californiensis* and ostracode (genera *Limnocythere*, *Candona*, and *Heterocypris*) fossils. Ostracode faunal data suggest lake level fluctuations, consistent with a transgressive – regressive cycle at the highstand. OSL data from the measured stratigraphic section constrain the highstand of Harper Lake between 45.6 ± 3.06 ka and 27.9 ± 1.73 ka. Calibrated radiocarbon ages from the measured section range from $33,645 \pm 343$ to $40,155 \pm 924$ cal yrs B.P. Our OSL and radiocarbon ages along with the continuity of the measured section support a

single Harper Lake highstand between ~45 and ~27 ka with no subsequent hiatus or second highstand at 25 ka as previously hypothesized. The OSL and radiocarbon data suggests that Harper Lake overlaps phases of Lake Manix.

Coalescent analysis of fifteen nuclear loci reveals low genetic diversity and Pleistocene speciation in the Mojave Fringe-toed Lizard, *Uma scoparia*

Andrew Gottscho

Humboldt State University, 4990 Lapis Lane, Pleasanton CA 94566, adg30@humboldt.edu

The Mojave Fringe-toed Lizard, *Uma scoparia*, is adapted for windblown sand habitats in the Mojave and Colorado Deserts of southern California and western Arizona. Due to the isolated nature of sand dunes, the geographic distribution of this species is fragmented and complex. A recent study using mitochondrial DNA suggested the existence of a distinct population segment in the lower Amargosa River drainage at the northern extent of the species' range. Here, I analyze fifteen nuclear loci (621,694 total base pairs) in a coalescent Isolation-with-Migration (IM) model to estimate the speciation time of *U. scoparia*, and effective population sizes for *U. scoparia* and the *U. notata* species complex. I also examined heterozygous single nucleotide polymorphisms (SNPs) and indels, as well as nucleotide diversity (π) by population to search for evidence of Pleistocene refugia. The role of Pleistocene glaciations, the San Andreas Fault, and the Colorado River in shaping genetic diversity within *U. scoparia* will be discussed.

Water on the Moon: how we know and what it means

David E. Harker

University of California San Diego/Center for Astrophysics and Space Sciences

If water at the poles of the Moon can be accessed by humans, then water can be used for life support resources at future lunar bases; specifically, water can be broken down into oxygen for humans and hydrogen for fuel. The early evidence for water existing at the lunar poles in permanently shadowed craters included the detection of excess H atoms by the Lunar Prospector neutron spectrometer and anomalous bistatic radar returns from the Clementine lunar orbiter. However, this evidence for water is indirect and the ice content controvertible (Schoughofer and Taylor 2007, Andreas 2007). Direct sampling by drilling is a future concept to unequivocally determine regolith water content; yet, such experiments currently are not on any NASA Lunar mission manifest. Therefore,

as a prelude to in situ searches, assessment of the polar regolith water content using remote sensing techniques was executed by the NASA Lunar Crater Observations and Sensing Satellite (LCROSS) Mission (Colaprete et al. 2008) impacting a permanently shadowed crater near the south pole of the Moon (Cabeus) at 11:31 UTC on 2009 October 09. LCROSS piggy-backed on the launch of the Lunar Reconnaissance Orbiter and used the spent second stage as an impactor. The mission objective of LCROSS was to impact and excavate permanently shadowed lunar regolith, ejecting this material above the crater rim into sunlight where measurements of water ice and water vapor entrained in the ejecta curtain using remote sensing methods from both the shepherding spacecraft and from ground-based observatories would enable extraction of the regolith water content.

Whether deposited in large quantities in discrete events (comets) or accumulating slowly and steadily for eons (solar wind implantation), volatiles that are delivered to the lunar poles are thought to be preserved in the permanently shadowed craters. The floors of such craters are so cold (<100 K) that water molecules could be cold-trapped there for billions of years (Vasavada 1999, Butler 1997). Over the history of the Moon, spanning back 3.9 Gyr to the late bombardment period, water molecules were delivered to the Moon by comet impacts and by impacts of meteorites containing aqueously altered minerals. Water molecules released into the lunar exosphere migrate on ballistic trajectories and spread eventually (~1 yr) to the lunar poles (Goldstein et al. 2001, Larignon et al. 2007). Indirect evidence for trapped water ice in the floors of lunar polar craters comes from (a) the detection of excess hydrogen atoms by the Lunar Prospector neutron spectrometer (Feldman et al. 1998, Feldman et al. 2000a, Feldman et al. 2000b, Feldman et al. 2001, Lawrence et al. 2006), which implies 1.5 ± 0.8 wt% H₂O in South polar craters (Feldman 2000a) and more specifically ≥ 20 wt% ice in ~10% of the floor area (Elphic et al. 2007), and (b) the anomalous bistatic radar returns from the Clementine lunar orbiter mission (Nozette et al. 1996, Nozette et al. 2001). However, Earth-based radar imaging does not reveal terrain properties characteristic of thick deposits of ice (Campbell et al. 2006). The radar imaging is consistent with the ice residing in spaces between regolith grains (Stacy et al. 1997, Campbell et al. 2003, Campbell et al. 2006, Crider and Vondrak 2003a, 2003b). Even though the floors of these craters do not see direct sunlight, surface volatiles can be removed by UV starlight, scattered UV zodiacal light, or by sputtering caused by interactions with the solar wind and the plasma that surrounds the airless body. In permanently shadowed craters, water ice best survives when hidden from surface

exposure. The LCROSS experiment was designed to address these latter issues, providing insight into how the ice domains evolve and are retained within the regolith, and ultimately enabling us to assess the quantity of water delivered to early Moon and early Earth by comet impacts during the late heavy bombardment period 3.9 Gyr ago.

The impact created by LCROSS produced a two-part plume from the bottom of the Cabeus crater: a high angle plume which is thought to contain mostly vapor and fine dust; and a low angle plume consisting mostly of heavier material. All of this material was exposed to direct sunlight for the first time in millions of years. In the quest for the search for water, the LCROSS team released a press report on 2009 November 13 showing the first evidence for the existence of water from the lunar regolith. Specifically, the onboard near-infrared spectrometer measured a spectrum with dips that can only be matched by the addition of water vapor and ice. Furthermore, the onboard ultraviolet/visible spectrometer measured emission lines indicative of other elements being released from the regolith during impact.

The ground-based observations are currently less conclusive. Keck near-infrared observations (using +NIRSPEC) reported a possible brightening of their spectrum post-impact. However, this has not been confirmed by other near-infrared observations including the NASA IRTF (+SpeX) and Gemini-North (+NIFS). I will address the current finding of both the spacecraft and ground-based observations of the impact.

Sources

- Andreas, E. L. 2007, *Icarus*, 186, 24
 Butler, B. J. 1997, *J. Geophys. Res.*, 102, 19283
 Campbell, B. A., et al. 2003, *Nature*, 426, 137
 Campbell, D. B., et al. 2006, *Nature*, 443, 835
 Colaprete, A. et al. 2008, [http://www.nasa.gov/centers/ames/multi-media/podcasting/2008/LCROSS First Step.html](http://www.nasa.gov/centers/ames/multi-media/podcasting/2008/LCROSS%20First%20Step.html)
 Crider, D. H. & Vondrak, R. R. 2003a, *Adv. Space Res.*, 31, 2293-2298
 Crider, D. H. & Vondrak, R. R. 2003b, *J. Geophys. Res.*, 108, 3845
 Elphic, R. C., et al. 2007, *J. Geophys. Res. Lett.*, 34,
 Feldman, W. C., et al. 1998, *Science*, 281, 1496
 Feldman, W. C., et al. 2000a, *J. Geophys. Res.*, 105, 4175
 Feldman, W. C., et al. 2000b, *J. Geophys. Res.*, 105, 20347
 Feldman, W. C., et al. 2001, *J. Geophys. Res.*, 106, 23231
 Goldstein, D. B., et al. 2001, *J. Geophys. Res.*, 106, 32841
 Lawrence, D. J., et al. 2006, *J. Geophys. Res.*, 111, E08001
 Nozette, S., et al. 1996, *Science*, 274, 1495
 Nozette, S., et al. 2001, *J. Geophys. Res.*, 106, 23253
 Schoughofer, N. & Taylor, G.J. 2007, *J. Geophys. Res. E*, 112, 02010
 Stacy, N., et al. 1997, *Science*, 276, 1527-1530
 Vasavada, A. R., Paige, D. A., & Wood, S. E. 1999, *Icarus*, 141, 179

Expanding the late Oligocene/early Miocene tectonic, magmatic and sedimentary history in the South Bristol Mountains

Janet Harvey¹, Joann Stock¹, and David Miller²

¹ California Institute of Technology, jch@gps.caltech.edu

² U.S. Geological Survey, 345 Middlefield Road, MS-973, Menlo Park, CA 94025

The southern Bristol Mountains lie near the eastern borders of the Mojave block and eastern California shear zone. The rock units composing the range are dissected by at least two generations of faulting: early Miocene normal faults associated with the Mojave block extensional terrane, and high angle northwest striking faults of the South Bristol Mountains fault (SBMF), including fault splays which cut Quaternary units. Total offset on the SBFM is estimated to be at least 6 km based on correlation of plutonic rocks across the fault zone. The range is composed of remnants of Proterozoic plutons, Jurassic hypabyssal and plutonic rocks, and metamorphosed Paleozoic passive margin sedimentary rocks. Neogene continental sedimentation is first preserved with latest Oligocene to middle Miocene sedimentary and volcanic rocks, unconformably overlain by thick sequences of coarse boulder conglomerates, and Quaternary alluvial, terrace and fan conglomerate deposits.

Our mapping shows that the basal nonconformity between Mesozoic basement and Miocene volcanic and sedimentary rocks is overlain in places by buff to dark orange, arkosic, massive, non-volcanic conglomeratic sandstone, and elsewhere a fluvial to lacustrine sequence of rhythmic sandstone beds with increasing volcanic input; only locally does the coarse nonvolcanic sandstone underlie the epiclastic volcanic sandstones. We have obtained a U-Pb zircon age of 23.8 ± 0.4 Ma (1 sigma) on an ash bed overlying the rhythmic beds; this is one of the older measured dates recording the onset of Neogene basin formation and deposition within the Mojave block. Sandstones inter-bedded within the volcanic rocks preserve rare rhizoliths and burrow structures, and a short camelid trackway has been identified in the fluvial sequences below the rhythmic lake beds. The lake bed sequence has been dismembered by Neogene and Quaternary faulting and crops out discontinuously at different elevations throughout the southern Bristol Mountains, but must have once had a significant areal extent in one or several sub-basins. Detailed mapping of the Neogene sedimentary and volcanic rocks should allow us to determine the amount of early Neogene extension within this region, and produce a more detailed total slip budget for the Quaternary strike slip faulting.

The effects of drought on host plant canopy condition and survival of the endangered *Astragalus jaegerianus* (Fabaceae)

T. R. Huggins¹, B. A. Prigge², M. R. Sharifi¹, and P. W. Rundel¹

¹Dept. of Ecology and Evolutionary Biology, University of California, Los Angeles, 90095-1606

²Mildred E. Mathias Botanical Garden, University of California, Los Angeles, 90095-1606

Astragalus jaegerianus Munz (the Lane Mountain milkvetch) is a federally endangered species that exists in only four fragmented populations within and adjacent to the U.S. Army's National Training Center, Fort Irwin, CA. Since 1999, our monitored *A. jaegerianus* populations have consistently declined, and are now 12% of their previous size. A number of subpopulations are in danger of local extinction. The decline of *A. jaegerianus* has occurred simultaneously with severe drought in the Mojave Desert. These drought conditions began in 1999 and are predicted to continue for decades, or may continue indefinitely under warmer temperature conditions projected by global climate change-type drought. Our results suggest that drought has direct and indirect effects on *A. jaegerianus* by killing or degrading its host shrubs. *Astragalus jaegerianus* host shrubs have decreased in shrub volume and cover by roughly 50 percent since the onset of drought, and shrub mortality has been high. Our results show that canopy condition has a profound effect on the microclimate within host shrubs. Furthermore, our results show a significant increase in survival of *A. jaegerianus* among host plants with more intact canopies. These results support our study hypothesis that drought-related changes to host plant canopies affect *A. jaegerianus* survival, and represent an indirect negative effect of long-term drought on *A. jaegerianus* populations.

Pliocene and Pleistocene lakes of Death Valley, California

Jeffrey R. Knott¹, Joseph C. Liddicoat², Elmira Wan³, David Wahl³, and Andrei M. Sarna-Wojcicki³

¹Department of Geological Sciences, California State University Fullerton, Fullerton, CA 92834, jknott@fullerton.edu

²Department of Environmental Sciences, Barnard College, Columbia University, New York, NY 10027, jliddico@barnard.edu

³Tephrochronology Project, U.S. Geological Survey, 345 Middlefield Road, MS 975, Menlo Park, CA 94025; ewan@usgs.gov; dwahl@usgs.gov; asarna@usgs.gov

In the modern, hyper-arid climate of Death Valley, California, lake deposits indicate extraordinary wet conditions in the past that are an extreme departure from today's

norm. In this study, we used geologic mapping, correlation of key tephra beds, and paleomagnetism along with dating methods to determine the timing and extent of Pliocene to middle Pleistocene lakes in Death Valley. The ~3.35 Ma tuffs of Zabriskie Wash and Mesquite Springs interbedded with fine-grained sediments are evidence that from about 3.3 to 3.2 Ma, a narrow lake was confined to the ancient Furnace Creek basin that stretched from present-day Ryan to Scotty's Castle. The ~2.06 Ma Huckleberry Ridge ash layer is found interbedded with evaporite and alluvial-fan deposits in both the Confidence Hills of southern Death Valley and the Kit Fox Hills to the north, respectively. The ~1.98–1.78 Ma tuffs of lower Glass Mountain are interbedded with alluvial-fan deposits at Artists Drive and east of the Confidence Hills and support a lake limited to the lower elevations of the valley. The evidence for Pleistocene lakes is relatively sparse and mainly at two locations: Mormon Point and the Kit Fox Hills. At both locations, fine-grained deposits interpreted as lake facies contain reverse-polarity ~1.13 – ~0.87 Ma Upper Glass Mountain ash layers and the Bruhnes/Matuyama paleomagnetic boundary. The ~0.76 Ma Bishop ash bed is found above these lake deposits interbedded with alluvial fans indicating a relatively dry interval. In contrast, the overlying ~0.64 Ma Lava Creek B ash layer is found interbedded with lacustrine deposits at both Mormon Point and Kit Fox Hills. The most extensive lake deposits are related to the 0.18–0.12 Ma deposits found throughout Death Valley. Core studies indicate that the 0.03–0.01 Ma lake deposits are confined to the lowest elevations of the valley.

Death Valley lakes are generally synchronous with lakes found to the west in Panamint, Searles, and Owens valleys. However, based on the limited exposures, with the exception of the 0.18–0.12 Ma lake deposits, Death Valley lakes are spatially limited. Searles Lake was full and overflowing at ~3.0 Ma, 1.0 Ma, 0.18 Ma, and 0.03 Ma. Unlike other pluvial basins, Death Valley was a terminal, closed basin with no overflow sill. Yet, at this time, the Death Valley record suggests that only the 0.18 Ma lake extended beyond the modern salt pan.

Paleomagnetic and radiocarbon record of the Searles Lake Formation (subunits A and ab) at Poison Canyon, San Bernardino County, California

Jeffrey R. Knott¹, Joseph C. Liddicoat², and Robert S. Coe³

¹Department of Geological Sciences, California State University Fullerton, Fullerton, CA 92834; jknott@fullerton.edu

²Department of Environmental Sciences, Barnard College, Columbia University, New York, NY 10027; jliddico@barnard.edu

³Earth Sciences Department, University of California, Santa Cruz, Santa Cruz, CA 95064

The Pleistocene–Holocene Searles Lake Formation of Smith (2009) records the last 150,000 years of pluvial Searles Lake in Searles Valley. The Pleistocene inflow from pluvial China Lake into Searles Lake was from the west via Poison Canyon in the southwestern portion of Searles Valley. The type locality for the Searles Lake Formation is along the south side of Poison Canyon where G.I. Smith described 20 lithologic units. In spite of the extraordinary efforts by Smith, the ages of the older subunits (A, ab1 and ab2) remained elusive with the only reliable age being >29 ka from the overlying ab3. In this study, we collected fossils for radiocarbon accelerator mass spectrometer (AMS) analysis and rock samples for paleomagnetic analysis. The improved AMS technique has a greater age range than the conventional analyses done by Smith. In addition, the older subunits should span two paleomagnetic excursions: Mono Lake Excursion (28–32 ka; MLE) and Laschamp Excursion (40.7 ka; LE). Our hypothesis is that the improved radiocarbon combined with the paleomagnetic data would determine the age of the older subunits at the type locality and allow correlation to similar deposits three kilometers west at the Tire Farm (informal name). At Poison Canyon, we obtained AMS radiocarbon ages of $42,900 \pm 100$ ¹⁴C yrs B.P. (all ages uncalibrated radiocarbon years) for subunit ab1, $31,830 \pm 170$, $35,540 \pm 420$ and $36,850 \pm 500$ for ab3, $28,280 \pm 110$ and $35,530 \pm 190$ for ab5 and $31,410 \pm 160$ for b. The Tire Farm sediments yielded ages of $40,210 \pm 400$ at the base of subunit A and $28,180 \pm 180$ and $29,150 \pm 200$ at the top of A. The overlying subunit ab3 yielded ages of $30,840 \pm 190$ and $47,150 \pm 190$. The radiocarbon ages are not consistently in stratigraphic order; however, these differences may be attributable to interferences from hard water and the changing paleomagnetic field. Reverse paleomagnetic polarity was recorded in sediments at both locations that were thermally demagnetized to 600 C. The combined mean paleomagnetic directions are $I=-37.5$, $D=180.2$, $\alpha-95=19.5$ ($n=12$) with a mean Virtual Geomagnetic Pole of 73.6 S, 231.8 E, $\alpha-95=20.6$

($n=12$). At Poison Canyon, the reverse polarity is in subunit ab2 of Smith whereas at the Tire Farm the reverse polarity is in subunit A of Smith. The reverse polarity and underlying radiocarbon ages, ~42 ka and 40 ka at Poison Canyon and Tire Farm, respectively are consistent with the LE and not the MLE. According to Smith (2009), the alternating mud and sand deposition of the A and ab subunits record fluctuating lake levels. If the correlation to the putative LE is correct, then Searles Lake was relatively deep with silt and mud deposits consistent with deposits below wave base. The paleo-shoreline fluctuated between Poison Canyon and Tire Farm between ~28 and 40 ka.

Stratigraphy, age, and depositional setting of the Miocene Barstow Formation at Harvard Hill, central Mojave Desert, California

Shannon R. Leslie¹, David M. Miller¹, Joseph L. Wooden², and Jorge A Vazquez¹

¹U.S. Geological Survey, 345 Middlefield Road, Menlo Park, CA 94025

²U.S.G.S.-Stanford Ion Microprobe Laboratory, Stanford University, Stanford, CA 94305

Our detailed geologic mapping and geochronology of the Barstow Formation at Harvard Hill, 30 km east of Barstow, CA, help to constrain Miocene paleogeography and tectonics of the central Mojave Desert. A northern strand of the Quaternary ENE-striking, sinistral Manix fault divides the Barstow Formation at Harvard Hill into two distinct lithologic assemblages. Strata north of the fault consist of: a green rhyolitic tuff, informally named the Shamrock tuff; lacustrine sandstone; partially silicified thin-bedded to massive limestone; and alluvial sandstone to pebble conglomerate. Strata south of the fault consist of: lacustrine siltstone and sandstone; a rhyolitic tuff dated at 19.1 Ma (U-Pb); rock-avalanche breccia deposits; partially silicified well-bedded to massive limestone; and alluvial sandstone and conglomerate.

Our U-Pb zircon dating of the Shamrock tuff yields a peak probability age of 18.7 ± 0.1 Ma. Distinctive outcrop characteristics, mineralogy, remanent magnetization, and zircon geochemistry suggest that the Shamrock tuff represents a lacustrine deposit of the regionally extensive Peach Spring Tuff (PST). When Shamrock tuff zircon age data are combined with zircon age analyses from three well-characterized PST samples, the peak probability age is 18.7 ± 0.1 Ma, thus providing new insight into the age of zircon crystallization in the PST rhyolite.

Results of our field studies show that Miocene strata at Harvard Hill mostly accumulated in a lacustrine environment, although depositional environments varied from a relatively deep lake to a very shallow lake or even onshore

setting. The rock-avalanche breccias and alluvial deposits near the base of the exposed section indicate proximity to a steep basin margin and provenance studies suggest a southern source for these deposits; therefore, we infer a southern basin margin setting at Harvard Hill during the early Miocene. Our geochronology demonstrates that deposition of the Barstow Formation at Harvard Hill extended from before ~19.1 Ma until well after ~18.7 Ma, similar to timing of Barstow Formation lake deposition in the Calico Mountains but at least 3 million years older than comparable facies in the Mud Hills type section. These observations are consistent with either of two paleogeographic models: westward transgression of lacustrine environments within a single large basin, or sequential development of geographically distinct eastern and western sub-basins.

A preliminary review of the effects of utility-scale renewable energy development on terrestrial wildlife in the desert southwest with emphasis on the desert tortoise

Jeff Lovich

*U.S. Geological Survey, Southwest Biological Science Center,
Colorado Plateau Research Station, 2255 North Gemini Drive,
MS-9394, Flagstaff, AZ 86001, jeffrey_lovich@usgs.gov*

The United States is poised to develop renewable energy sources at an unprecedented rate, including potentially large areas of public land in the west. This quantum leap is driven by escalating costs and demand for traditional energy sources based on fossil fuels. Attention is largely focused on renewable forms of energy especially wind and solar. Development potential for renewable energy is particularly high in the southwest United States (AZ, CA, NV, and UT), where wind and solar energy are abundant and already being harnessed (Wilshire and Prose, 1987). However, the potential for renewable energy conflicts with natural resources, especially wildlife, is also very high given the exceptional biodiversity and sensitivity of arid Southwest ecosystems (Lovich and Bainbridge, 1999), including the Mojave (Mittermeier et al., 2002) and Sonoran deserts.

A great deal of information is available on the effects of renewable energy development and operation on the environment (see reviews by Abbasi and Abbasi, 2000; Pimentel et al., 1994), although much of the information exists in environmental compliance documents and other “gray literature” (Kuvelsky et al., 2007). A substan-

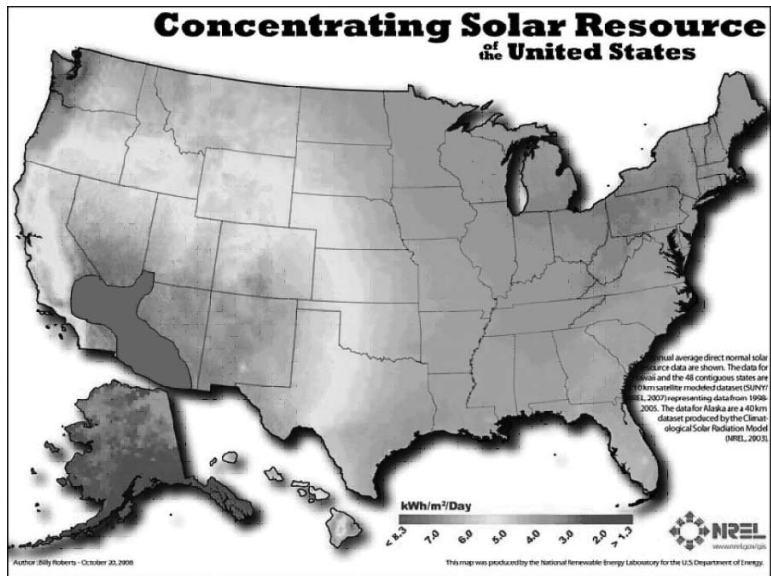


Figure 1. Solar energy potential of the United States. Superimposed is a dark shape defining the approximate outline of the range of the desert tortoise in the United States.

tion portion of this research has been conducted on the effects of wind energy development on birds (Drewitt and Langston, 2006) and bats (Arnett et al., 2007; Kunz et al., 2007) due to their sensitivity to aerial impacts. In

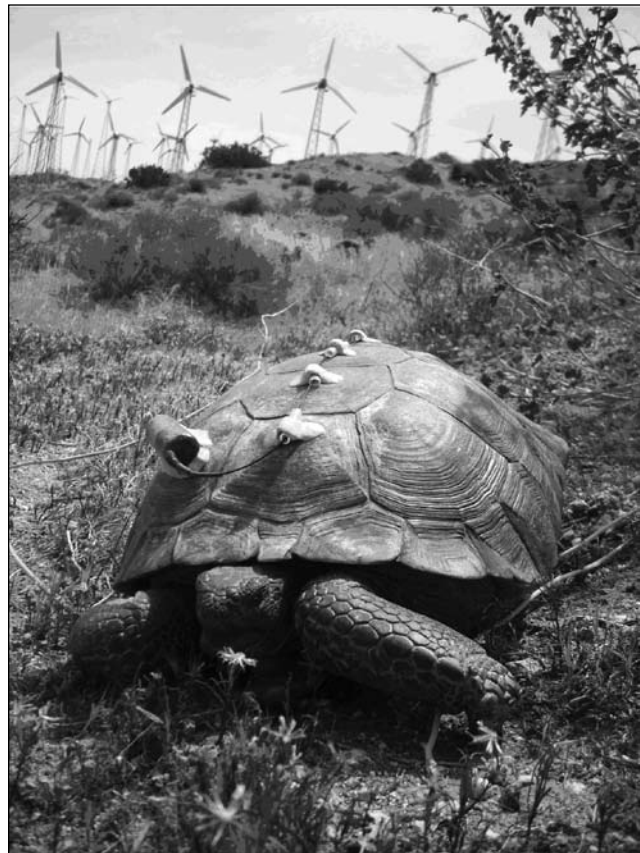


Figure 2. A radio-tagged desert tortoise at a wind energy farm in the San Geronio Pass near Palm Springs, California. Photo by the author.

contrast, very little information is available on either the effects of wind energy on non-volant wildlife (animals that don't fly), or the effects of solar energy development on wildlife in general, including sensitive species. From a conservation standpoint, one of the most important species in the desert Southwest is the desert tortoise, which is classified as federally threatened north and west of the Colorado River. Large areas of desert tortoise habitat are potentially developable for renewable energy extraction (Figure 1) and some areas occupied by tortoises (e.g., wind energy facilities in the San Geronio Pass of California) are already developed. Almost nothing is known about the effects of renewable energy development on this species (Pearson, 1986) although recent research on the effects of wind energy facilities (Figure 2) suggests that not all impacts are negative (Lovich and Daniels, 2002).

The construction of both utility-scale wind and solar energy facilities involves significant ground disturbance and associated impacts to terrestrial wildlife and habitat, especially for concentrating solar projects. Direct impacts also include accidental wildlife mortality associated with both construction and maintenance operations. For wind energy, largely unquantified impacts include the effects of industrial noise, vibrations and electromagnetic field generation on wildlife although some research is available on the effects of noise on marine mammals (Koschinski et al., 2003; Tougaard et al., 2008). Another study suggested that noise generated by wind energy turbines affected the behavior of California ground squirrels (Rabin et al., 2006). The utility-scale proposals for both wind and solar energy development in the desert also pose unquantified impacts related to habitat fragmentation. The desert Southwest is traditionally characterized by large blocks of continuous and interconnected habitat. Large-scale energy development is likely to fundamentally change this landscape through fragmentation, presenting potential barriers to movements and genetic exchange in wildlife populations including bighorn sheep, deer, desert tortoises and other species of concern.

Other potential effects of utility-scale energy development on wildlife and habitat are related to the extraction of large amounts of raw materials for construction (aggregate, cement, steel, and glass), the need for large amounts of water for cooling some installations, and the potential for production of toxic wastes including coolants, antifreeze, rust-inhibitors, and heavy metals (Abbasi and Abbasi, 2000). In addition, water used for steam production at a solar energy facility in the Mojave Desert of California contained selenium. The waste water pumped into evaporation ponds attracted birds that fed on invertebrates. Although selenium toxicity was not a

threat based on the results of one study (Herbst, 2006), the possibility exists for harmful bioaccumulation of this toxic micronutrient.

Many unanswered questions remain after reviewing the available literature of renewable energy development on non-volant wildlife. What are the cumulative effects of energy development? What density of development maximizes energy benefits while minimizing effects to terrestrial wildlife? Where are the best places to site energy development relative to the needs of wildlife? Is renewable energy development compatible with wildlife conservation? In addition, most available information is based on correlative data and "before and after" replicated studies are needed to assess impacts in a rigorous experimental framework (Kuvlesky et al., 2007). Detailed information on wildlife distribution and habitat needs are critically needed to assist in proper site location and design of renewable energy developments (Tsoutsos et al., 2005).

Literature cited

- Abbasi, S.A. and N. Abbasi. 2000. The likely adverse environmental impacts of renewable energy sources. *Applied Energy* 65:121-144.
- Arnett, E.B., W.K. Brown, W.P. Erickson, J.K. Fiedler, B.L. Hamilton, T.H. Henry, A. Jain, G.D. Johnson, J. Kerns, R.R. Koford, C.P. Nicholson, T.J. O'Connell, M.D. Piorkowski, R.D. Tankersley, Jr. 2007. Patterns of bat fatalities at wind energy facilities in North America. *J. Wildl. Mgmt.* 72:61-78.
- Drewitt, A.L. and R.H.W. Langston. 2006. Assessing the impacts of wind farms on birds. *Ibis* 148:29-42.
- Herbst, D.B. 2006. Salinity controls on trophic interactions among invertebrates and algae of solar evaporation ponds in the Mojave Desert and relation to shorebird foraging and selenium risk. *Wetlands* 26:475-485.
- Koschinski, S., B.M. Culik, O.D. Henriksen, N. Tregenza, G. Ellis, C. Jansen, and G. Kathe. 2003. Behavioral reactions of free-ranging porpoises and seals to the noise of a simulated 2 MW windpower generator. *Marine Ecology Progress Series* 265:263-273.
- Kunz, T.H., E.B. Arnett, W.P. Erickson, A.R. Hoar, G.D. Johnson, R.P. Larkin, M.D. Strickland, R.W. Thresher, M.D. Tuttle. 2007. Ecological impacts of wind energy development on bats: questions, research needs, and hypotheses. *Front. Ecol. Environ.* 5:315-324.
- Kuvlesky, W.P., Jr., L.A. Brennan, M.L. Morrison, K.K. Boydston, B.M. Ballard, F.C. Bryant. 2007. Wind energy development and wildlife conservation: challenges and opportunities. *J. Wildl. Mgmt.* 71:2487-2498.
- Lovich, J. E. and D. Bainbridge. 1999. Anthropogenic degradation of the southern California desert ecosystem and prospects for natural recovery and restoration. *Environmental Management* 24:309-326.
- Lovich, J. E. and R. Daniels. 2000. Environmental characteristics of desert tortoise (*Gopherus agassizii*) burrow locations in an altered industrial landscape. *Chelonian Conservation and Biology* 3(4): 714-721.
- Mittermeier, C. G., W. R. Konstant, R. E. Lovich, and J. E. Lovich. 2002. The Mojave Desert. In, pp.351-356, *Wilderness: Earth's Last Wild Places*. Mittermeier, R., C. G. Mittermeier, P. Robles Gil, G. Fonseca, T. Brooks, J. Pilgrim, and W. R. Konstant (eds.), CEMEX, Mexico.

- Pearson, D.C. 1986. The desert tortoise and energy development in southeastern California. *Herpetologica* 42:58-59.
- Pimentel, D., G. Rodrigues, T. Wang, R. Abrams, K. Goldberg, H. Staecker, E. Ma, L. Brueckner, L. Trovato, C. Chow, U. Govindarajulu, and S. Boerke. 1994. Renewable energy: economic and environmental issues. *BioScience* 44:536-547.
- Rabin, L.A., R.G. Coss, and D.H. Owings. The effects of wind turbines on antipredator behavior in California ground squirrels (*Spermophilus beecheyi*). *Biological Conservation* 131:410-420.
- Tougaard, J., P.T. Madsen, and M. Wahlberg. 2008. Underwater noise from construction and operation of offshore wind farms. *Bioacoustics* 17:143-146.
- Tsoutsos, T. N. Frantzeskaki, V. Gekas. 2005. Environmental impacts from solar energy technologies. *Energy Policy* 33:289-296.
- Wilshire, H. and D. Prose. 1987. Wind Energy Development in California, USA. *Environmental Management* 11:13-20.

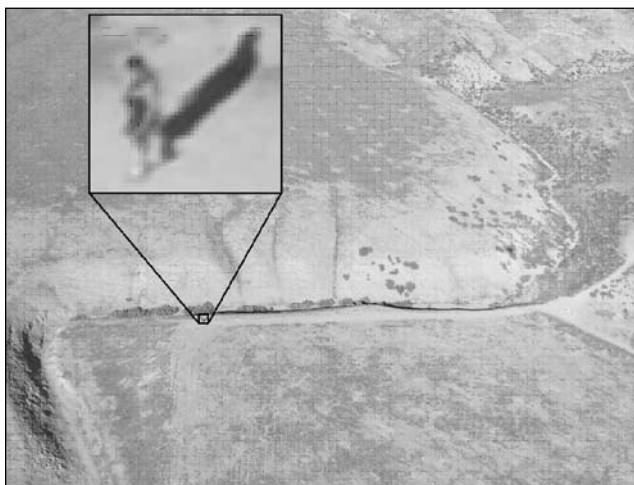
Aerial surveys using consumer electronics: fast, cheap and best of all: useful!

David K. Lynch¹, Kenneth W. Hudnut¹, and David S. P. Dearborn²

¹U.S. Geological Survey, 345 Middlefield Road, MS-973, Menlo Park, CA 94025

²Lawrence Livermore National Laboratory

We report results from two low-cost, low-altitude, aerial imaging surveys of the San Andreas Fault (SAF) carried out in late 2009. In total 541 km of the fault was imaged with a ground sample distance (pixel size) of a few cm. The two surveys covered the Carrizo Plain and points north to the Choice Valley on 24 Sep 2009, and the SAF between I-5 (Tejon Pass) and I-15 (Cajon Pass) on 29 Dec 2009. Each area was imaged twice, once on the first pass and a short time later on the return pass. The I-5 to I-15 flight included Lone Pine Canyon east of Wrightwood soon after the Sheep Fire of early Oct 2009. Ground that was normally covered by heavy brush was revealed for the first time in many years.



Wallace Creek – See <http://www.sanandreasfault.org/SAFMojave29Dec2009.html> and <http://www.sanandreasfault.org/SAFCarrizo24Sep2009.html>

The data set consists of 5216 ~6Mb jpg photographs (31 Gb total) which were posted on the internet within hours of their acquisition. Shortly thereafter they were placed into PICASA web albums for easy browsing. Total cost for both surveys (excluding camera) was about \$5000.

The pictures were taken with a Nikon D90 with an attached GP-1 receiver that wrote the aircraft's position into the EXIF file of each photograph. Organization, manipulation and geolocation of the images were done on a Macintosh laptop.

All photographs are freely available and carry no copyright. They are in the public domain.

Comparison of two Pliocene (Blancan) vertebrate fossil assemblages: Panaca Local Fauna (Lincoln County, Nevada) and Hagerman Local Fauna (Twin Falls County, Idaho)

Vicki Meyers and Stephen Rowland

University of Nevada Las Vegas, Department of Geosciences, 4505 Maryland Parkway, Las Vegas, NV 89014; meyersv2@unlvnevada.edu, steve.rowland@unlv.edu

The Pliocene Epoch was a significant time in North America for climatic fluctuations and intercontinental dispersal of mammals during a relatively short time interval of less than three million years. The Pliocene transitioned from a warmer to a cooler climate which influenced environmental changes from an open forest to grasslands in the continental mid-latitudes of North America.

The Panaca local fauna within the Panaca Formation of southeastern Nevada and the Hagerman local fauna within the Glens Ferry Formation of south central Idaho are inland mid-latitude terrestrial Pliocene fossil sites. The Hagerman fauna within the Hagerman Fossil Beds National Monument (HAFO) is well known and significant in its diversity, quantity, and quality. The Panaca local fauna has been studied sporadically beginning with Chester Stock, later by the Frick Laboratories, and more recently by Robert Reynolds, Everett Lindsay, and Yun Mou. The Panaca Formation in southeastern Nevada represents an excellent interval to compare Pliocene, (specifically Blancan), faunal assemblages in diversity and paleoecology.

The Glens Ferry Formation is a continuous section of fluvial and lacustrine strata spanning approximately 1 million years while the fluvial, lacustrine and eolian Panaca sediments are discontinuous and partially eroded. Well data suggest that as much as half the Panaca Formation may be buried. Seasonal climates of both sites are probable, but the Panaca Formation likely represents a drier climate. The Panaca fossils seem to be concentrated in

the fine-grained fluvial sands on the outer margins of the basin. Many HAFO fossils are concentrated in sand channels, but a variety of fossils are found stratigraphically throughout the ~4000 acre monument. A preliminary comparison shows the Panaca fauna of Meadow Valley to represent a less diverse ecosystem but with greater diversity of Cricetidae rodents. HAFO probably experienced a wetter climate than Panaca, providing more habitats along a meandering stream and lake ecosystem, which was able to support a more diverse fauna of birds, fish, amphibians, and mammals, along with freshwater mollusks and plants. Families common to both localities include Soricidae (shrews), Leporidae (rabbits), Heteromyidae (pocket mice), Geomyidae (pocket gophers), and Cricetidae (voles). Large mammals common to both localities include Equidae (horses), Camelidae (camels), Tayassuidae (peccaries), Canidae (dogs), and Felidae (cats). Mammal families found only at HAFO are Castoridae (beavers), Megalonychidae (ground sloths), Ursidae (bears), Mustelidae (weasels, skunks, and badgers), Mammutidae (mastodons), Antilocapridae (pronghorn) and Cervidae (deer), while families exclusive to Panaca area are Bovidae (sheep and goats) and Procyonidae (ring-tailed cat).

Reconnaissance geochronology of Miocene basins in the central Mojave Desert; implications for basin evolution and tectonics

D. M. Miller¹, S. R. Leslie¹, J. W. Hillhouse¹, J. L. Wooden¹, J. A. Vazquez¹, and R.E. Reynolds²

¹U.S. Geological Survey, 345 Middlefield Road, Menlo Park, CA 94025

²Redlands, California, 92373

The Barstow Formation, which at its type section in the Mud Hills is ~19.5 to ~13.3 Ma, is a primarily lacustrine and distal alluvial fan sequence that has been identified across a wide area of the central Mojave Desert. At the type section, near its northwest extent, the age of the Barstow is constrained by six dated tuffs and tuffaceous sandstones. In the Calico Mountains, the unit is bracketed by underlying 19.0 Ma Pickhandle Formation volcanic rocks and 17.1 to 16.8 Ma volcanic rocks that intrude it. Elsewhere, the Barstow Formation has yielded Hemingfordian and Barstovian vertebrate fossils but the unit is otherwise poorly dated. In an attempt to improve knowledge of the chronology and depositional environment for the Barstow Formation, we are dating zircons in tuff beds and testing ash-flow tuffs for the distinctive magnetic pole of the Peach Spring Tuff.

Thus far, we have learned that the Barstow beds in the Calico Mountains include tuff beds as old as 19.1 Ma. Eastward, in the Yermo Hills, a younger tuff, about 15.2

Ma, lies in Barstow beds. At Harvard Hill, a tuff between 20.3 and 19.1 Ma and an exceptionally thick tuff dated at ~18.8 Ma, and thought to represent a lacustrine deposition of the Peach Spring Tuff, occur in the Barstow Formation. Far to the south, zircon dating and paleomagnetic results demonstrate that the Peach Spring Tuff lies in or above Barstow Formation-type rocks both north and south of Daggett Ridge. The new dates indicate that Barstow lake deposition commenced well before 19 Ma in several areas and continued to 15 to 13 Ma, apparently initiating earlier to the south and southeast. The lake beds overlap the time assumed to represent maximum extension for the nearby central Mojave metamorphic core complex and in many places are partly or wholly coeval with the Pickhandle Formation, which is thought to represent syn-extensional deposition and magmatism. The new data indicate that the central Mojave Desert contained broad basins during extension (~21 to 17.5 Ma) and afterward, and that timing of volcanism, alluvial fan, and lacustrine deposition were not simply related to extensional tectonics.

Physiological responses of Mojave Desert shrubs to simulated summer stream channel flow

April R. Newlander¹, David R. Bedford², David M. Miller², and Darren R. Sandquist¹

¹Department of Biological Science, California State University, Fullerton, 800 N State College Blvd, Fullerton, CA 92834

²U.S. Geological Survey, 345 Middlefield Road, MS-973, Menlo Park, CA 94025

In Mojave Desert bajada ecosystems, water infiltrates to root-zones in greatest proportion via stream channels (washes). High intensity summer rainfall events exceed infiltration and generate runoff creating wash flow. Desert washes have a pronounced effect on plant dispersion and size across these landscapes, as larger and more plants are found adjacent to washes. As a winter-rainfall dominated ecosystem, climate changes in the Mojave that increase summer precipitation may play an important role in altering vegetation processes influenced by washes. The purpose of this project is to examine how increased summer precipitation may affect vegetation properties of a bajada in the Mojave National Preserve. Plant responses to summer precipitation were quantified by measuring ecophysiological responses of two dominant perennial shrubs, creosote bush (*Larrea tridentata*) and white bursage (*Ambrosia dumosa*), to a pulse of isotopically (δD) labeled water distributed directly into a wash. Responses included net photosynthesis (A_{net}), stomatal conductance (g_s), xylem water potential (Ψ_x) and stem-water stable isotope ratio

(δD). For all ecophysiological traits, both species showed pronounced responses to the summer wash pulse of water. These responses varied generally as a function of distance from wash for both species and showed responses up to 3 m from the wash. Water uptake responses for both species were similar with elevated water potentials persisting for over a month, not returning to pre-pulse values for at least 75 days on average. In spite of similar water uptake patterns between creosote bush and white bursage, there were temporal differences in the use of this water (based on stomatal conductance), which is likely due to differing leaf phenologies. Six days following the pulse, new growth on white bursage plants and the unfolding and expansion of existing creosote bush leaves was evident. Greenness of plants adjacent to wash was noticeably greater than those further from the wash for at least 6 months following the pulse event. These results contribute to improved understanding of wash dynamics in desert landscapes, which is needed to better understand how climate and landscape changes will alter ecosystem productivity of arid systems like the Mojave Desert.

Factors affecting recruitment of desert holly (*Atriplex hymenelytra*) in the Mojave National Preserve

Odara E. Pineda

Department of Biological Sciences, California State Polytechnic University, Pomona, odara_pineda@sbcglobal.net

Atriplex hymenelytra, commonly known as desert holly, is a salt tolerant perennial that grows in discrete stands on gravel fans of moderate salinity in deserts of the Southwest. At Soda Springs, the location of our investigation, the abundance of desert holly in study plots monitored since 1991 has slowly declined. Past studies also demonstrate that seeds are more likely to germinate in the vicinity of seed-producing plants, yet have a higher rate of survivorship away from them. As a dioecious plant, only female members of the population produce seeds enclosed by bracts and therefore seed production by desert holly will be limited by the abundance and size of females in a population. The primary objective of this study was to investigate several factors that could affect the recruitment of new individuals into the population including seed production (fecundity), soil conditions within various seed “bank” microsites, and seed predation. To assess fecundity, 10 female plants were enclosed by mesh nets in May 2008 prior to the release of bracts, and bracts were collected in early November of the same year. All bracts were counted and seeds from a percentage of these were examined for viability. In an effort to better understand germination and survivorship patterns

of seedlings, soil was collected from four microsites: in the vicinity of female plants, males plants, dead desert holly, and bare ground; and several physical features (pH, conductivity, moisture, temperature, and light availability) assessed. The seed predation study simply aimed to establish the presence of desert holly seed predators. Trays each containing 550 desert holly bracts were made available to vertebrates (rodents/lizards) only, ants only, both, or neither, and left under creosote bushes for one week after which all remaining bracts were counted. Bract production by female desert holly does not appear to be correlated to plant size. All physical variables measured during the collection of soil samples varied significantly among the four microsites in at least one of the sampling quarters, except for moisture. Bract loss among the various seed predation treatments was significant. Further investigations of plant fecundity and “banked” seeds within soil samples are ongoing.

A half-million years of paleoclimate and effects on landscape, Mojave Desert, California

Marith Reheis and David Miller

U.S. Geological Survey, 345 Middlefield Road, MS-973, Menlo Park, CA 94025

Paleolakes in the arid southwestern U.S. responded sensitively to past climate change, and their deposits contain paleoclimate information that can help refine forecasts of the effects of future climate change on water supplies and surface processes. During the middle and late Pleistocene, Lake Manix was the terminus for the Mojave River during periods of enhanced runoff from the Transverse Ranges, until the lake drained east to form Lake Mojave about 25,000 years ago. The sedimentary record reveals a complex history of fluctuations influenced by changes in climate, basin configuration (multiple sub-basins), and drainage integration events that must be evaluated to isolate the paleoclimate signal.

Studies of a 45-m core from the Manix basin, combined with outcrop data, span the interval 500-25 cal ka (calibrated thousand years). A well-dated lake-level curve shows multiple highstands during MIS (Marine Isotope Stage) 3 and early MIS 2 (~45-25 cal ka), at times when the Laurentide ice sheet was well below its maximum height and geographic extent and long before pluvial highstands in the Great Basin lakes to the north. These data suggest that some episodes of moderately deep water occurred during interglacials and interstadials, as well as glacial periods. However, the integrated record of lake levels in Lake Coyote, a subbasin of Lake Manix, and Lake Mojave were similar to those of other pluvial

lakes between 25 and ~9.9 cal ka, except that a dry period from 17.5 to 16 cal ka in Lake Estancia (east-central New Mexico) appears to have been relatively wet in the Mojave Desert. These data suggest that marine and atmospheric conditions in the late Pleistocene had non-uniform effects across the Great Basin.

Trackways of a gregarious, Early Jurassic therapsid, Aztec Sandstone, Valley of Fire State Park, southern Nevada

Stephen M. Rowland

Department of Geoscience, UNLV, Las Vegas, NV 89154-4010, steve.rowland@unlv.edu

The Lower Jurassic Aztec and Navajo sandstones of eastern California, southern Nevada, southern Utah, and northern Arizona represent a vast sandy desert that formed near the western margin of the supercontinent Pangaea. These desert sand-dune and inter-dune deposits have yielded very few fossil bones, but they contain a relatively diverse assemblage of trace fossils. One of the most common taxa of trace fossils is the mammaloid ichnogenus *Brasilichnium*, which was first described from the Jurassic of Brazil. The *Brasilichnium* trackmaker was most probably a therapsid (= mammal-like reptile), the synapsid group that evolved into mammals.

The focus of this study is one exposure of Aztec Sandstone in Valley of Fire State Park. This exposure contains approximately 100 *Brasilichnium* tracks, in eleven distinct trackways. The tracks are all subparallel to one another, heading up the slip face of a Jurassic sand dune. Track size varies from trackway to trackway, indicating that this site records a mixed-age group of animals, scampering up the face of a dune. Many of the tracks appear to have broken through a crust on the surface of the dune, leaving steep-sided tracks with an empty central shaft. The fact that these tracks retained their steep-sided morphology requires that the sand was very cohesive and must have been moist from rain or heavy dew. The tracks were apparently preserved when loose, dry sand buried the moist (or formerly moist) track-covered crust.

Gregarious behavior is well known in some groups of dinosaurs, but evidence of gregarious behavior in Mesozoic mammals and therapsids is scanty. Some previously described Navajo Formation trackway sites in Utah have multiple trackways of *Brasilichnium*, so it is not a complete surprise to find evidence that the *Brasilichnium* trackmaker lived in social groups. This discovery complements the recently reported occurrence of prairie-dog-town-like complexes of decimeter-scale burrows in inter-dune sediments of the Navajo Sandstone in Utah; these burrow complexes have also been interpreted as evidence of social behavior in Lower Jurassic dune-field-dwelling

therapsids. Together with the burrow complexes, the Valley of Fire *Brasilichnium* trackway site provides the most compelling evidence yet discovered of gregarious behavior among Mesozoic synapsids generally, not only in the southwestern U.S., but (as far as I know) anywhere in the world.

Late Holocene Lakes and Horticulture in the Mojave Sink

Claude N. Warren

Professor Emeritus, University of Nevada, Las Vegas

The late Holocene climate produced perennial lakes in the Mojave Sink. These lakes supported *Anodonta*, whose shells constitute a major component of late prehistoric middens in the area. These lakes would have supported aboriginal horticulture during two periods. Reports of prehistoric corn cobs from the area support this suggestion. During these same periods, associated ceramics and other artifacts reflect strong influence in the area from Four Corners area Anasazi during the earlier period, and from the Lower Colorado Hakataya during the later period. During both of these periods, the Turquoise Mountains were actively mined. The turquoise was transported east and southeast; none is reported from the California coast. It is suggested here that horticulture was practiced in the Mojave Sink, sustaining the miners and supporting Anasazi and Hakataya control of the turquoise mines.

Rock hounds and mining claims

Greg Wilkerson

Bureau of Land Management, 3801 Pegasus Drive, Bakersfield 93308-6837, gregg_wilkerson@ca.blm.gov

Traditionally, rock and mineral collectors ("rock hounds") have performed their activities on public lands under the regulatory definition of "casual use" that is described in Title 43 Code of Federal Regulations (CFR) 3622. For some rock hounds, collection is a business. Commercial mining activities are managed and U.S. Bureau of Land Management (BLM) lands under 43 CFR 3800 for locatable minerals and under 43 CFR 3600 for salable minerals. On U.S. Forest Service lands (FS), locatable minerals are managed under 36 CFR 228 Subpart A and salable minerals are managed under 36 CFR 228 Subpart C. Rock hounding clubs are almost always non-profit organizations, but the mineral shows they sponsor are commercial activities. Some rock hounds and some clubs own mining claims, and so have invested themselves in an exclusive property right for the severance of the locatable mineral estate from the public domain. These claims are sometimes in wilderness areas or other special management areas administered by federal land management

agencies. The special rights of mining claimants create a special category of land and resource management for the U.S. Park Service (PS), BLM, and FS.

Some materials that rock hounds collect, like gemstones and geodes, are locatable under the 1872 mining law (as amended). Common varieties of stone including common agate, common obsidian, and other materials used by rock hounds to create jewelry and art works are not locatable. These rocks and minerals can be sold by surface management agencies as a “mineral material” under the Mineral Materials Act of 1947 and the Common Varieties act of 1955. Petrified wood is a special class of stone that can be collected in quantities of “25 pounds plus one piece per day” without a permit as long as explosives or mechanized equipment is not used (43 CFR 3622). Fossils are another special category of material collected and sold by rock hounds. Only common non-vertebrate fossils can be collected under “casual use.” Collection of vertebrate or rare invertebrate fossils requires a permit under the provisions of 43 CFR 8270.

When a proposal for exploration or development is received by a federal surface management agency (PS, BLM or FS), that agency must conduct a common variety determination to classify the rocks and minerals of that proposal as locatable or salable. The outcome of the Common Variety Determination is the decision document that specifies whether the exploration or development will be managed under 43 CFR 3800 for locatable minerals and under 43 CFR 3600 for salable minerals.

Fast-tracking solar development in the desert

Howard Wilshire

Public Employees for Environmental Responsibility, howardgw@comcast.net

Grand plans for accelerated development of utility-scale solar power plants in southwestern U.S. are underway with blessings of the Department of the Interior, assorted environmental groups, and a rapidly growing industry. Public lands provide essentially free land to go along with the free sun energy. Any project that gets underway by the end of 2010 also will get free money from the stimulus funds. Of the 250,000 square miles of southwestern deserts deemed suitable, the grand planners call for 48,000 to be developed by 2050 and 173,000 (equal to the total area of California, Maryland, and Massachusetts) by 2100. The land desired has slopes not greater than 3%, and dominant technologies have very low tolerance for land surface irregularities (such as ephemeral washes). Hence, the land is typically graded to slopes of <1%, thus eradicating existing ecosystems. In addition, a new national high-

voltage direct current transmission system will have to be built. Alternatives include widespread use of commercial and residential rooftop solar, use of urban brownfields, conservation, and passive solar building.