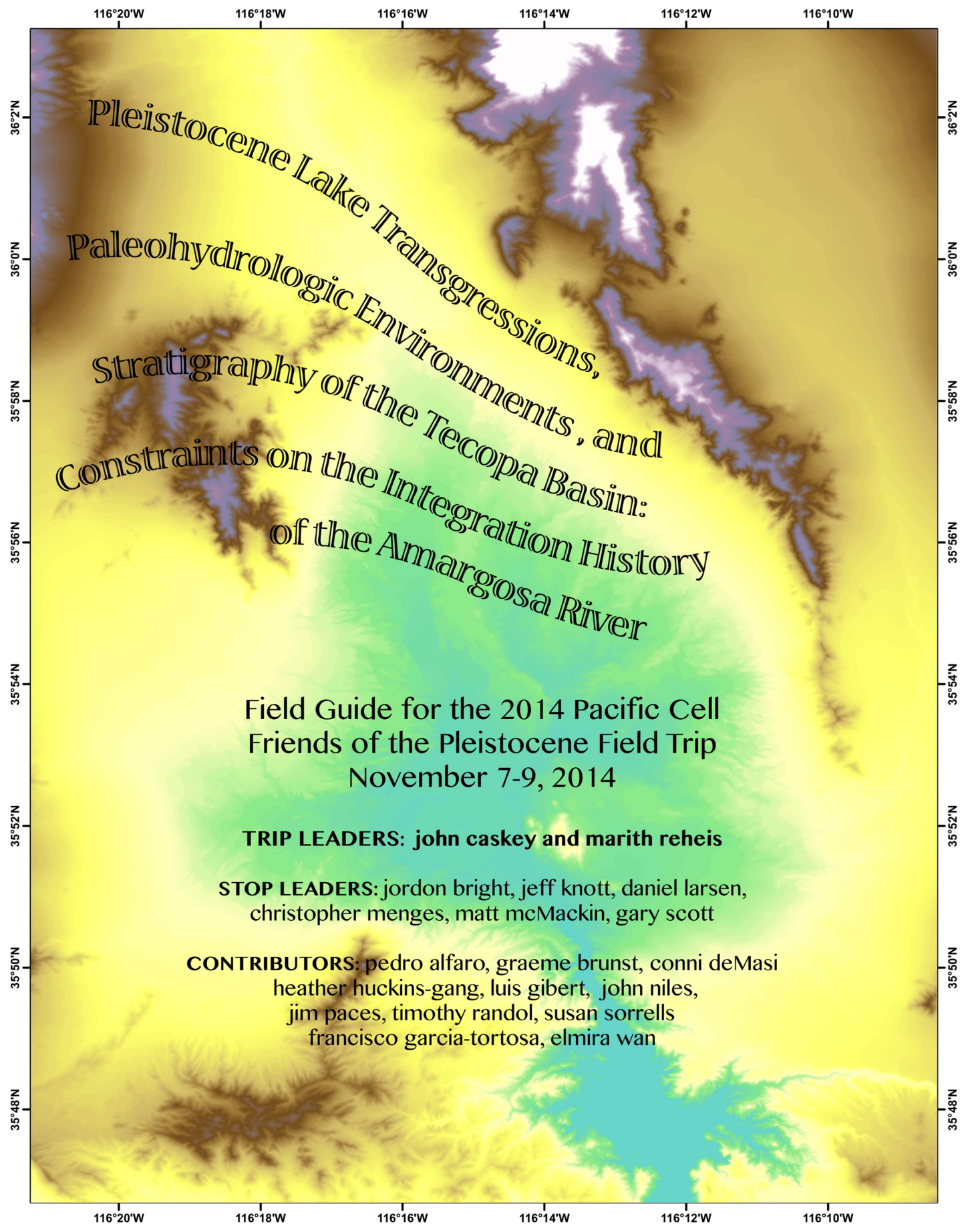

Pleistocene Lake Transgressions, Paleohydrologic Environments, and Stratigraphy of the Tecopa Basin: Constraints on the Integration History of the Amargosa River

Field Guide for the 2014 Pacific Cell
Friends of the Pleistocene Field Trip
November 7-9, 2014

TRIP LEADERS: john caskey and marith reheis

STOP LEADERS: jordon bright, jeff knott, daniel larsen, christopher menges, matt mcMackin, gary scott

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Photo taken on a GSA field trip to the Tecopa basin in 1999, the last field trip ever led by Roger Morrison. L-R: Blair Jones (USGS mineralogist and chemist, dec. 2014), Ike Winograd (USGS, groundwater hydrology, Devils Hole paleoclimate record), Rick Forester (USGS ostracodes and paleoclimate--dec. 2014), Alan Riggs (worked with Ike on Devils Hole), Marith Reheis, Roger Morrison (USGS, dec. 2006), Marty Mifflin (Nevada hydrogeologist, worked on Nevada pluvial lakes and with Roger in Tecopa), and Dwight Schmidt (USGS hydrogeologist and geologic mapper, dec.--2007)

Roger Barron Morrison (1914-2006)

Quaternary Geology Scientist, Colleague, and Close Friend.

By M.D. Mifflin

I feel exceptionally privileged to offer an overview of Roger Barron Morrison's contributions to the sciences of the Quaternary, and some idea of the man, from the perspective of a close friend and colleague for over forty years. Roger, normally an extremely private man, in relaxed moments over many years told me bits and pieces of his experiences. From memory with the help of his resume, I offer the following:

Roger was born in Madison, Wisconsin, the son of a Professor of Animal Science, one of two original editors that published what became, with many updated revisions, the bible on animal nutrition for the next 60 or so years. The Morrison family became rather wealthy from this independent source of income, and Roger's father rose through the ranks of academia to become the Dean of the College of Animal Science at Cornell University.

Roger earned a BS in Geology and an MA (1934) in Economic Geology by age 20 at Cornell. In the 1934-1935 academic year, he began work on a Ph.D. in Economic Geology, University of California, Berkeley, then transferred to Stanford (1935-1938) and completed all requirements, but never completed revising his dissertation (Structural and Mineralogical Evolution of Ore Deposits in the Bald MountainElkhorn Ridge area, Baker and Grant Counties, Oregon) because of illness and WWII duties. The illness was brucellosis, and he was pretty much incapacitated for more than a year. He had picked it up from drinking raw milk during field work. He initially chose economic geology one of the lead geoscience specialties of that era because of his father's interests. His father invested in mining ventures. Roger described his first paid professional assignment 10 (on his summer break) at Olinghouse, a gold property his father was considering near Wadsworth, Nevada. Roger spent the summer of 1934 reviewing the mine (located within sight of the Wadsworth Amphitheater stratigraphic section of the Lake Lahontan, the most complete exposure of the Eetza Alloformation known). He managed to uncover evidence that it had been "salted", and was essentially mined out. Roger would have been 20 years old that summer, and 60 years later when I heard the account he was still pleased that he saved his father from a carefully planned scam.

In 1935 Roger published his first professional paper based on his Master's thesis on the occurrence and origin of celestite and fluorite at Clay Center, Ohio, in the American Mineralogist. In 1940, as second author (with Joe Poland). An electrical resistivity apparatus for testing well waters was published in the American Geophysical Union Transactions, a basic design that is still marketed. Roger had begun his career with the U.S. Geological Survey in 1939, and his first open-file report was on groundwater resources, Big Sandy Valley, Mojave Co., Arizona. From 1942 to 1947, Roger was with the Military Geology Branch, USGS, and basically prepared military intelligence maps and reports dealing with geomorphology, materials, and water supply for areas in Europe, South America, and Asia. He prepared military training manuals on water-supply development. At times, he worked directly with operations commanders on key intelligence questions: river bank conditions for crossing the Rhine, and beach conditions for the invasion of the Philippines. In Roger's subsequent work, no matter the

topic, he was expert in establishing comprehensive literature searches and often incorporated earlier findings into his writings. This talent is well illustrated in the sections he authored in the DNAG volume K-2, Quaternary Nonglacial Geology: Conterminous U.S. I also admired his clear and concise writing style; he was always close to getting it right on the first draft when we worked together on reports.

Roger's interest in Quaternary geology was likely stimulated by the military intelligence work (European geologic literature) and the period of groundwater studies in the basins of Arizona and New Mexico before WWII. He got his opportunity in 1949 to pursue this interest when he was named Chief of the Fallon, Nevada, Project, a study of Tertiary and Quaternary stratigraphy in four 15-minute quadrangles in the Carson Desert of Nevada, the largest sub-basin of Lake Lahontan. The occurrence of several water wells yielding natural gas had caused much speculation and provided the incentive for the study. I never thought to ask Roger how he landed the project, but I suspect it might have been up for grabs because of its flat, desolate nature.

It did not take long after beginning this project for Roger, often with coauthors, to begin publishing on Quaternary topics. In 1952 he submitted two abstracts (GSA Bulletin) on the stratigraphy of Lake Lahontan, and the late Quaternary climatic history of the northern Great Basin. With Gerry Richmond and Howard Bissell as co-authors, another GSA Bulletin abstract was offered that year on correlations of Late Quaternary deposits in the La Sal Mountains and the Lahontan and Bonneville basins based on interglacial soils.

In 1957 two more publications came out with additional co-authors, James Gilluly and Charles Hunt. From that period on Roger was publishing frequently on Lake Lahontan, Lake Bonneville, and regional correlations of Quaternary deposits based on geosols (the topic of another Morrison Ph.D. dissertation, this one at the University of Nevada, Reno (UNR). This time around, he was granted the Ph.D. in 1964, and it was the first issued in any field by this institution. Roger's mentor at UNR was George Burke Maxey. Dr. Maxey (my boss at the Desert Research Institute, a research branch of UNR) formerly taught at the University of Illinois, and was a close friend of John Frye, widely recognized for his work on glacial tills and associated stratigraphy in the Midwest. Roger and John Frye published (1964) on a proposed correlation of Middle and Late Quaternary sequences, beginning in the Lake Lahontan basin, to Lake Bonneville, to the Wasatch Range, to the Southern Great Plains and western Midwest sequences. Roger was publishing with the elite in the US geologic profession, cooperative efforts triggered by his detailed stratigraphic work in the Lahontan basin and the use of geosols for long distant correlations of widely varying Quaternary deposits and climates.

It was during this period (1963-1965) that I met and worked with Roger for the first time. He was extremely busy with the combination of finishing up his dissertation, USGS work on Lake Bonneville's stratigraphy, preparing the paper co-authored with John Frye, and putting the finishing editorial touches on his U.S. Geological Survey Professional Paper 401 on Lake Lahontan (1964). In addition, he was a co-organizer, with Clyde Wahrhaftig, for the Northern Great Basin and California Field Excursion of the 1965 Congress of INQUA. He was more than spread thin, and during a GSA meeting in the Fall of 1963 at Reno, he asked me if I would help him by giving backup support to Margaret (Peg) Wheat while she measured and described some key Lake Lahontan

stratigraphic sections that winter and spring. The request came as a surprise he barely knew who I was, but I quickly agreed.

At the time, I didn't know any of the above, or question how it was that he came to ask me for help. In retrospect, it was likely Burke Maxey who recommended me to Roger. Burke knew that I was interested in Quaternary geology and that I had already attempted to break out and map some of Roger's stratigraphic units at the eastern margin of the Carson Desert (unsuccessfully). I was ready to learn something from the master as Burke referred to Roger's position in the Quaternary sciences.

Over the next year and a half, I worked with Roger and Peg Wheat on the two localities where the stratigraphic relationships of early lake cycles were exposed, plus saw many other localities of the younger lake cycles with key relationships, benefiting from the last 15 years of Roger's efforts. And so began two working relationships (and friendships) that would endure until their deaths decades later.

Roger and I got off to a good understanding at a major highway cut through the root zone of a Lake Lahontan pointbar. It offers a 3-D exposure of Sehoo gravels overlying the subaerial Wyemaha alluvium and eolian sand, with the well-developed Churchill Geosol preserved in some areas, and the underlying Eetza bar gravels also exposed. It was the last field day to finalize the 1965 INQUA fieldtrip road log and send it off to the printer. We had run out of daylight, and were making up the descriptions with the aid of headlights. In frustration with the poor, indirect light, I muttered aloud the question of what the B horizon colors might prove to be in natural daylight, as Roger stood by taking notes. Roger said nothing for a long pause then something to the effect, “You know, this highway is too dangerous for 50 people to be crossing the highway and wandering around. We will make it a slowdown locality. By this time, I was well aware of Rogers quest for accuracy and detail in anticipation of several European experts on geosols participating in the field trip. I began to laugh, as only a few minutes before this was to be a half hour stop. Roger also began to chuckle, as he knew that I knew he didn't want to be beat up over some poor guesses on colors, and he had just made an executive decision to avoid such problems. We selected the color matches together and headed for the closest town, Lovelock, and food at the only place still open, a small, rather seedy casino. Roger promptly ordered the finest wine on the well-worn wine list, and after the traditional taste, refused it as spoiled! The young waiter, trained to offer the first taste, had no idea of what to do if the most expensive bottle in the house was refused. Roger, suspecting that all the good wines on the list had been stored upright for years, then ordered the house wine. Roger, the master of the Quaternary, and Roger, the Character, all in one day!

Seemingly minor because it's old hat now, but perhaps one of the more important contributions that Roger's early work established was the 3-D mapping approach to facies changes and age relationships in the lacustral and interbedded subaerial bolson deposits, with lateral mapping of varying marker horizons, geosols and tephra, and also diagnostic lithologic characteristics would sometimes prove useful. What appears to be look-alike fine-grained deposits in disjunct exposures begin to have real character, with predictable facies changes. This detailed 3-D mapping strategy later helped Roger's disciples, including Jay Quade and myself, to sort out lacustrine deposits from lake beds' of a different origin. Experienced (and highly respected)

field geologists had mapped ”lake beds’ in nearly all southern basins of the Great Basin to that period; many of these beds are now understood to be paleo-groundwater discharge deposits.

After looking back at some of Rogers early publications, I decided to review the three chapters that he contributed to the DNAG Volume K-2 published in 1991. I became hooked and finished reading the entire volume (about a week later). Roger's fingerprints are present throughout the volume, and it's an exceptional compilation from both the topical and regional chapter perspectives. This may not have been Roger's last hurrah, but in my view, it's his best. It incorporates excellent summaries of his major contributions, and he is in excellent company with 100 other authors both detailing and summarizing regional relationships and providing comprehensive discussions of supporting sciences. As editor and contributor, he established his own memorial and his place in the Quaternary sciences. Each of us may have differing views of the contributions he has made, but this volume lays out the full spectrum for consideration. I found it fascinating how extensive his field work and knowledge was of many regions. Not only did Roger make the volume happen (see the Foreword), but he guided it into a major scientific contribution. The structuring of the chapters allowed all of the many authors to offer their best analyses and reviews.

One of the interesting aspects I never gave much thought to until this review was the apparent splash of Roger's Lake Lahontan-Lake Bonneville work and inter-regional correlations generated in the 1950's and 1960's. In the context of the stage of development of Quaternary sciences at this period, these detailed records were somewhat analogous to the impacts of subsequent detailed long Quaternary records such as the ocean core analyses and ice core records offering much greater detail and new insights into varying climates than had been documented before. Also, the inter-regional correlations using geosols came at a time when none of the currently adopted constraining dating techniques had been fully developed or applied. Suddenly, here were two records (Bonneville and Lahontan) with complete records of pluvial lake cycles and stages accompanied by major soil-forming interludes. Other regions had geosols and continental deposits, but not the essentially continuous sedimentation records to establish more confidence in the comprehensive nature of the records. The new records were a major breakthrough at the time.

Roger ranked the Lahontan, Bonneville, and Tecopa basin Quaternary studies as his most important contributions. He believed these basins offered the longest, most complete and most detailed paleoclimatic records in continental settings. He often referred to his efforts in the Tecopa basin as his 'last hurrah' and he was clearly the most excited researcher I've known when it became clear that funding to support his fieldwork in this basin was forthcoming through State of Nevada technical oversight program for the proposed Yucca Mountain high-level waste repository. In his 70's when the field support became available (Roger was my choice for this important study of the paleoclimates in this hydrographic basin) he was confronted with a dilemma shortly thereafter: he wasn't convinced he could manage both the Tecopa study and editorial efforts on the DNAG volume at the same time. He asked me what I thought (pointing out his age) and I urged him to take on the editorship.

No memorial to Roger would be complete without elaborating on his other

claim to fame, Roger the Character. In 1963 I was briefed by two colleagues before I met Roger in person. Burke Maxey (also considered by many as a character) pointed out that if I was interested in Quaternary geology he was the master. Another colleague, however, while acknowledging Roger's impressive Lake Lahontan work, couldn't wait to tell me about Roger, the Character. Roger's standard field attire in warm weather up to about that time (Roger began to cover up shortly before I met him due to a bout with skin cancer) was essentially nothing but a cowboy hat, boots, and the minimum (controversy continues as to exactly what) in terms of modesty. This attire, when added to Roger's favorite mode of transportation for fieldwork in the desert bolsons horse created quite a sight. One can imagine the effect of a seemingly naked cowboy riding around the desert, brown as a berry, going nowhere, as he rode up and down the dry washes. Roger used a horse (he owned a series) because he would systematically examine every wash bank or topographic feature that offered the possibility of an exposure, however limited, plus the horse offered efficient access to key localities often widely spaced in the typically soft ground terrain. In this way he could review possible stratigraphic relationships over very large areas in a reconnaissance sense, and then begin more detailed studies of the key localities.

One account (provided initially by my new colleague, but verified later by participants) captures some of the behavior patterns that set Roger apart. On the day of moving to a new field camp location in the eastern Carson Desert (SE corner of Eightmile Flat), Peg Wheat, his field assistant, was left to set up camp, while Roger took off by horseback for reconnaissance in the new area (Roger rarely wasted any time available for field work). It seems that around midday, Roger was close to Salt Wells, a former Pony Express stop and at that time the location of an isolated bar-restaurant on US Highway 50, near the northwest corner of Eightmile Flat. Roger stopped in to order two cases of fine French wine (Roger, regardless of location or conditions, liked his creature comforts, one of which was a glass or two of top quality wine for dinner). He also expected good service (also regardless of location or conditions). Apparently, out of nowhere, he marched though the door and up to the bar in his warm weather field attire. Without preliminaries, he asked if he could order the two cases of wine (the husband and wife owners had never seen anything quite like this and at first suspected a prank). The order, however, if real, would make them a good profit. They agreed, but suggested he check back after they could find out from their distributor the price and time of delivery. This satisfied Roger as reasonable, and he walked out the door, got on the horse, and rode southeast out across Eightmile Flat, a direction to absolutely nowhere (except the new field camp). All of this, including the transportation mode and the direction, was observed with total amazement by the owners. He was their first almost naked cowboy, and his order (high-end French wine by the case) was beyond anything the locals could cook up. There was a happy ending, as Peg Wheat was assigned to check back, knew the couple, and assured them that Roger was real and good for cost of the wine. Roger got his supply of camp wine without missing a beat in his field work, the husband-wife owners had a bar story that was hard to make better by exaggeration, and some 50 years later it gives the reader some insight as to why many thought Roger was a character. From Roger's standpoint it was just good logistics. Or was it?

During the past 20 some years, I interacted with Roger frequently. We traveled together, several trips in rough foreign settings, and his demeanor never changed. He expected good service and quality wine and food, and again, the circumstances didn't matter (while at times just getting by was my objective). During these years Roger's appearance was such that he was sometimes mistaken for Willie Nelson (especially in foreign settings). He got a kick out of this, and once he confided that during his undergraduate years at Cornell he had become very active in theatre. Even considering it as a career! Until he told me this, I had assumed he was oblivious to what was 'normal' behavior. Was it possible that Roger the Character was, at least in part, staged'

Roger's primary focus in life was his professional work, which largely dealt with Quaternary research whenever possible. From the time I met him, until the last two months of his life at the age of 92, he was actively engaged and remarkably productive. He met and married Harriet, his wife, in 1941 who was, at that time, a geologist employed by an oil company. She had just been evacuated to the US after being stationed in the Philippines. Harriet was his loving helpmate and companion. She looked after Roger in all matters domestic, as well as helped him with his work (see the DNAG volume K-2 Forward). Roger spoke of Harriet and his family often (three sons and daughters-in-law, and associated grandchildren) and he ranked his family very high in his priorities. However, other than Harriet, who knew him very well, the rest of his family may not have known the complete Roger. If one did not work with him and share in his primary interests and goals. He was opaque; he rarely revealed his feelings and passions (unless he disagreed with you on something he considered important!)

He had strong feelings about some conservation issues, and what he perceived as bad government drove him wild. He felt this way about the proposed high-level nuclear waste repository at Yucca Mountain (a “dry” design in highly fractured welded tuffs in the vadose zone didn't seem that dry during pluvial climates to Roger, or to me, for that matter). Another major issue led Roger into forming, with several others, the Colorado Chapter of the Sierra Club when the Bureau of Reclamation planned to dam all the major canyons of the Colorado River (they added the Marble Canyon Dam, but not the next one as planned).

Anybody who has attempted research has their secret list of heroes in science; I'm fortunate to have known two as colleagues and friends. Roger was one of them.

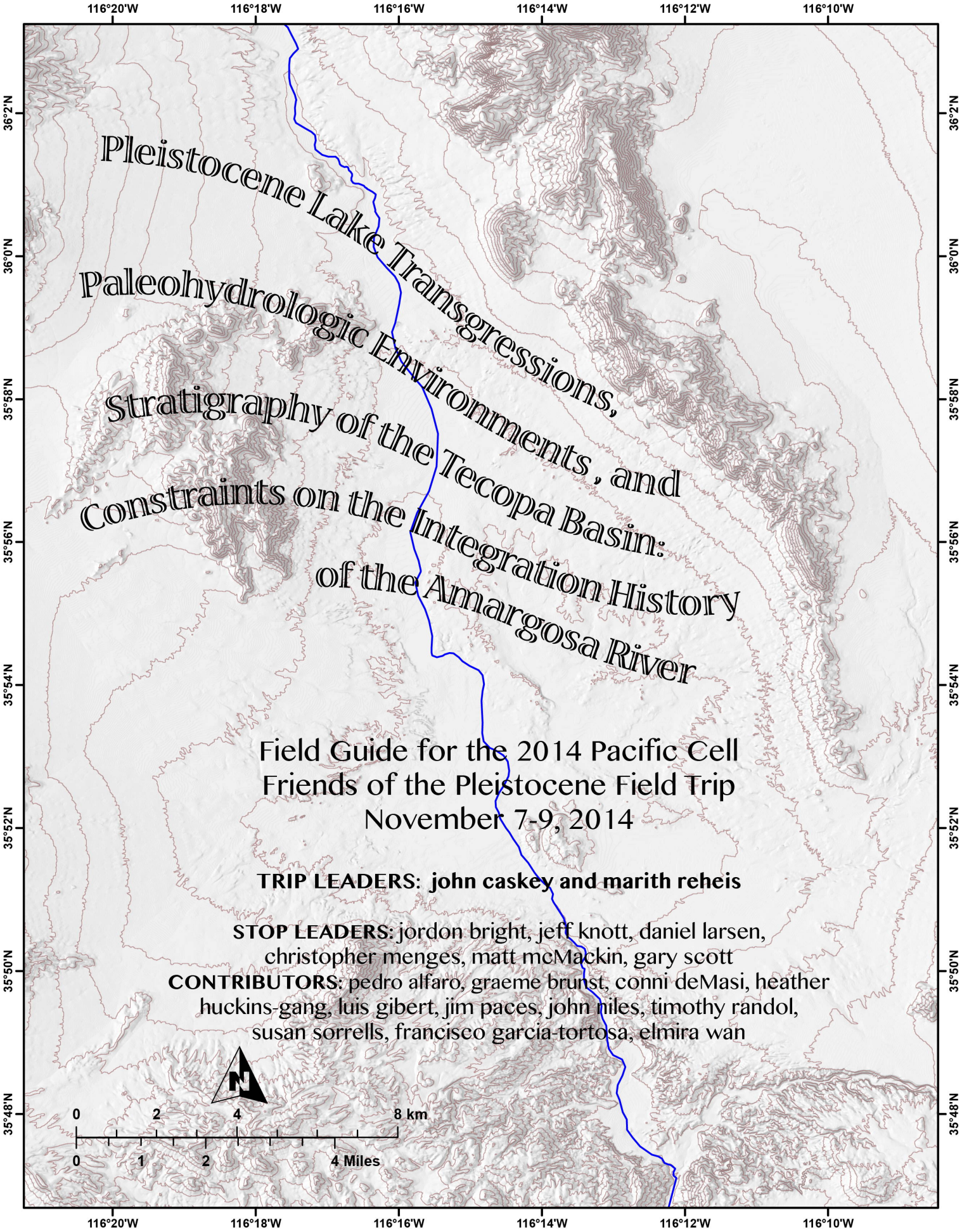
Richard M. Forester (1948-2014)

Photograph by Bob Thompson

Eulogy by Saxon Sharpe

Richard M. Forester passed away suddenly at his home in Lakewood, Colorado on March 27, 2014. He was born December 23, 1948 in Hancock, NY to Clayton and Ellen Forester. (His father was one of the previous owners of The Hancock Herald). His parents predeceased him. He is survived by his sister Mary Beth Michaud of Middletown, NY, a niece Elizabeth, a nephew Robert, a great niece Olivia, an aunt Helen Lester, and several cousins. The nonmarine science community lost an innovative and critical thinker. Rick pioneered the use of nonmarine ostracode species as indicators of present and past hydrologic conditions. He demonstrated that each species' position in the solute evolutionary path defined ecological and hydrochemical conditions that convey valuable information about the hydrology of the nonmarine water body. He also identified the importance of groundwater in the hydrologic budget of many lakes by demonstrating that many ostracode species prefer groundwater discharge environments. Rick extended these concepts to the interpretation of fossil ostracode assemblages and showed that the paleohydrology of water bodies (extant or extinct) could be reconstructed from them. Rick's expertise in studies of past and present hydrologic systems was pivotal to investigations of the long-term impacts of climate and groundwater on waste isolation at the proposed geologic repository for high-level radioactive waste at Yucca Mountain in southern Nevada. Rick led the U.S. Geological Survey's Yucca Mountain Climate Program from 1991 to 2000. His reconstructions of past climate in the Southwest established the foundation for predictions of future climate and its potential impact on repository hydrology for the Department of Energy. His genius for numbers and patterns allowed him to recognize a relation between the Devils Hole, Nevada, isotope record and Milankovitch cycles that could be projected into the future to estimate climate cycles. Rick had an astute and analytical mind, because of that he was able to effectively lead a multidisciplinary team working on the Yucca Mountain Climate Program, as his knowledge was very broad and did not stop at ostracodes and groundwater. Rick

worked in the USGS from 1977 until he retired in October, 2005. During his career, he discovered and described new ostracode species from different environments in western North America. He devised sampling protocols and documented ostracode species' modern geographic ranges, climates, and aquatic chemistry and, with Alison Smith, Don Palmer, and Brandon Curry, made this information available online at the North American Nonmarine Ostracode Database (NANODe). Rick interacted across scientific disciplines with numerous USGS scientists, including nonmarine diatom specialist Platt Bradbury and other paleolimnologists, to devise approaches to use multiple indicators to construct past lacustrine and wetland environments. In 2005, Rick was nominated and elected by his peers as a Fellow of the American Association for Advancement of Science for meritorious efforts to advance science in the fields of past climate and hydrology, in particular working with nonmarine ostracodes. He authored over 100 professional publications and reports during his career. He served on numerous Ph.D. committees, a tribute for someone not in academia. He mentored many graduate students on the identification and the ecological and hydrochemical significance of nonmarine ostracode assemblages and engendered a new community of nonmarine ostracode workers in North America and overseas. He was passionate about the graduate students he helped and his commitment to their science and well-being continued well after their graduation. His research interests were broad, and colleagues from many fields sought his collaboration because of his keen mind, ability to think outside of the box, and attention to detail. He was patient, generous with his time, and always ready to help, whether it be a student, a colleague, or a neighbor. He was fascinated by the science and mathematics of the stock market and created successful investment strategies for himself and a few lucky friends. He was good natured with a good sense of humor. He loved good beer, good red wine, and enjoyed anything containing chili peppers (the hotter the better), and dessert. Cats were always a favorite interest, and he particularly enjoyed the companionship of Tauron, a Persian-tabby cat of notable perspicacity. Tauron would sit next to the microscope and watch him work, taking great interest in the transfer of microfossils from picking tray to slide, on the tip of a brush. Indeed, after the passing of Tauron, he chose not to keep a pet, observing that he had met the best. Rick grew up in the small town of Hancock, New York, in the western Catskills. His family history was steeped in the culture of rural New York and the coal country of Pennsylvania around Scranton. His mother taught in a one-room schoolhouse, and his father owned the local newspaper. He was the first in his immediate family to go to college. He received a B.S. in Geology from Syracuse University in 1969. He began graduate school at the University of Illinois, Urbana, supported through a Teaching Assistantship. He completed an M.S. on stable isotope geology in 1972 under Tom Anderson, and a Ph.D. on probability and ostracode ecology in 1975 under Philip Sandberg. He was urged by John Carter, a research paleontologist at Illinois, to apply for a National Research Council postdoctoral position with the U.S. Geological Survey. Rick prepared a submission to Joseph Hazel, U.S. Geological Survey, proposing further work on probability and ostracodes. He was awarded the NRC postdoc, but found that he would be working on biostratigraphy of Miocene marine ostracodes at the USGS offices at the National Museum of Natural History in Washington, D.C. Upon completing the postdoc, he was hired as a Research Geologist by the U.S. Geological Survey in Denver, Colorado. His new research project was working with Tom Fouch and the Energy Team on nonmarine ostracodes from the Paleogene Green River Formation in Utah. He spent considerable time learning the difficult systematics and paleoecology of nonmarine ostracodes. Through this project, he became acquainted with Denis Delorme, Canada Center for Inland Waters, a specialist in nonmarine ostracode taxonomy, and with Blair Jones, a geochemist/hydrologist at USGS Reston, as Rick delved deeper into understanding the geochemical environment of nonmarine ostracodes. While working as a teaching assistant at University of Illinois CU Geology Department, Rick met and married Elisabeth Brouwers, currently an ostracode specialist and stratigrapher at the USGS. During their 15-year marriage they published several papers on estuarine ostracodes together.



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Friends of the Pleistocene Field Trip
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Pacific Cell FOP 2014: Tecopa Basin
Field Trip Road Log and Stop Descriptions (Days 1-3)

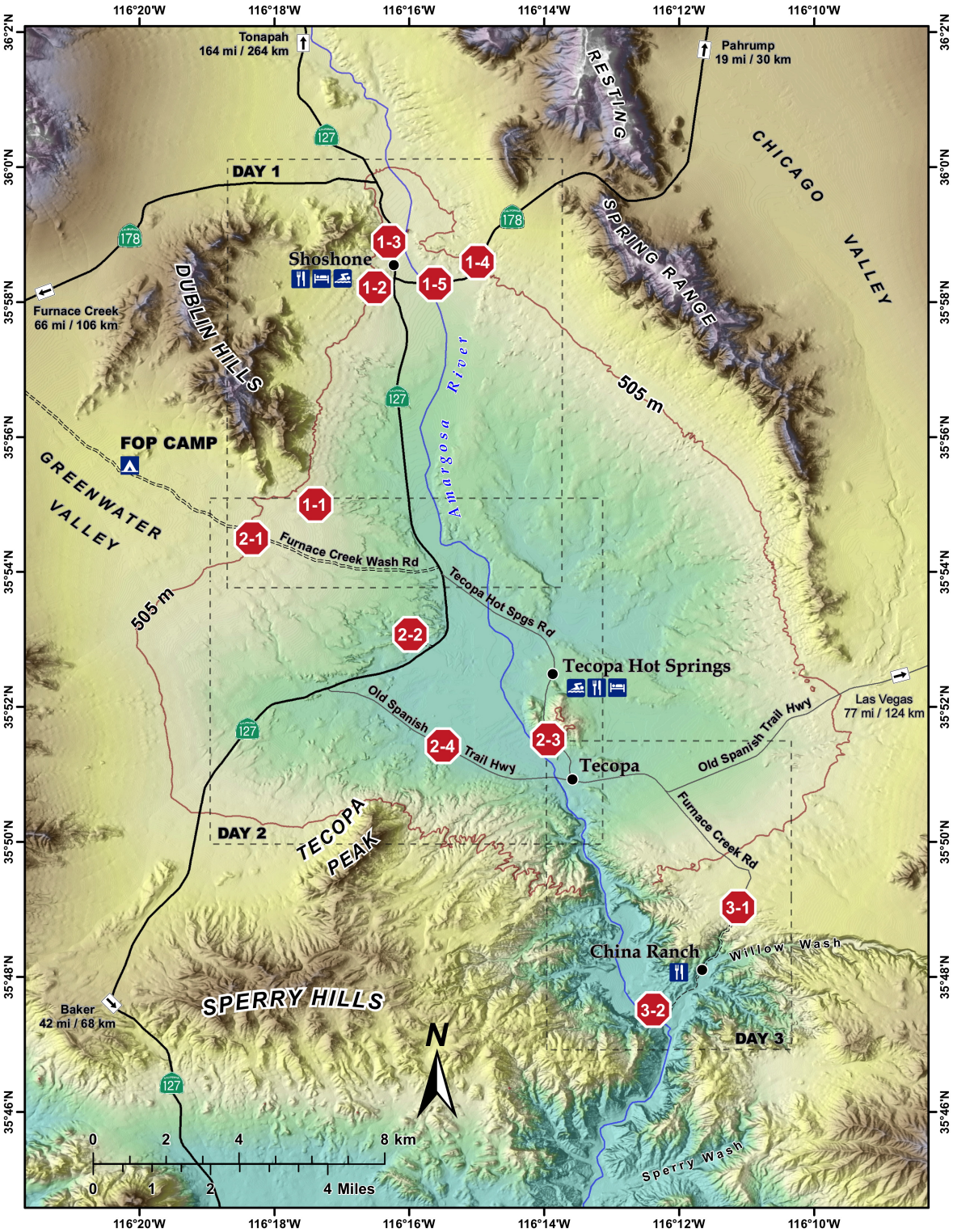


Figure 1. Tecopa basin FOP 2014 trip map showing stops for Days 1-3. An (optional) stop is also planned for Day 3 (Stop 3.3) to Eagle Mountain located 17 mi north of Shoshone along Hwy 127.

Day 1: Down Furnace Creek Wash (FCW) Road to east of Shoshone

The first day's primary objective is to lay out our recent observations related to the lake-level history and paleohydrologic environments of the Tecopa Basin, which are directly tied to the integration history of the Amargosa River watershed. We will be discussing these themes throughout the trip. At Stop 1.1, we'll summarize our main observations regarding lake-level history and present evidence that two prominent middle Pleistocene shoreline sequences (0.64 Ma to ~ 0.50 Ma) are significantly uplifted in the northern part of the basin relative to correlative shorelines in the southern part of the basin. The four stops near Shoshone at the north end of the basin provide excellent exposures of these two sequences where they transition with fluvial-deltaic deposits derived from the proto(?) Amargosa River. The Dublin Gulch dwellings (Stop 1.2) and the town of Shoshone share a fascinating Old Western history (Sorrells, A1.7). On the Stop 1.2 traverse we well be looking at spectacular quarry exposures of paleoseismically-induced(?) liquefaction features in Lava Creek ash beds (Scott) dated at 0.64 Ma (Lanphere et al., 2002). On this traverse as well as other Day 1 stops we will be hiking through some of the detailed stratigraphic sections of Larsen (2008) where he will discuss aspects of his research on local controls on basin facies.

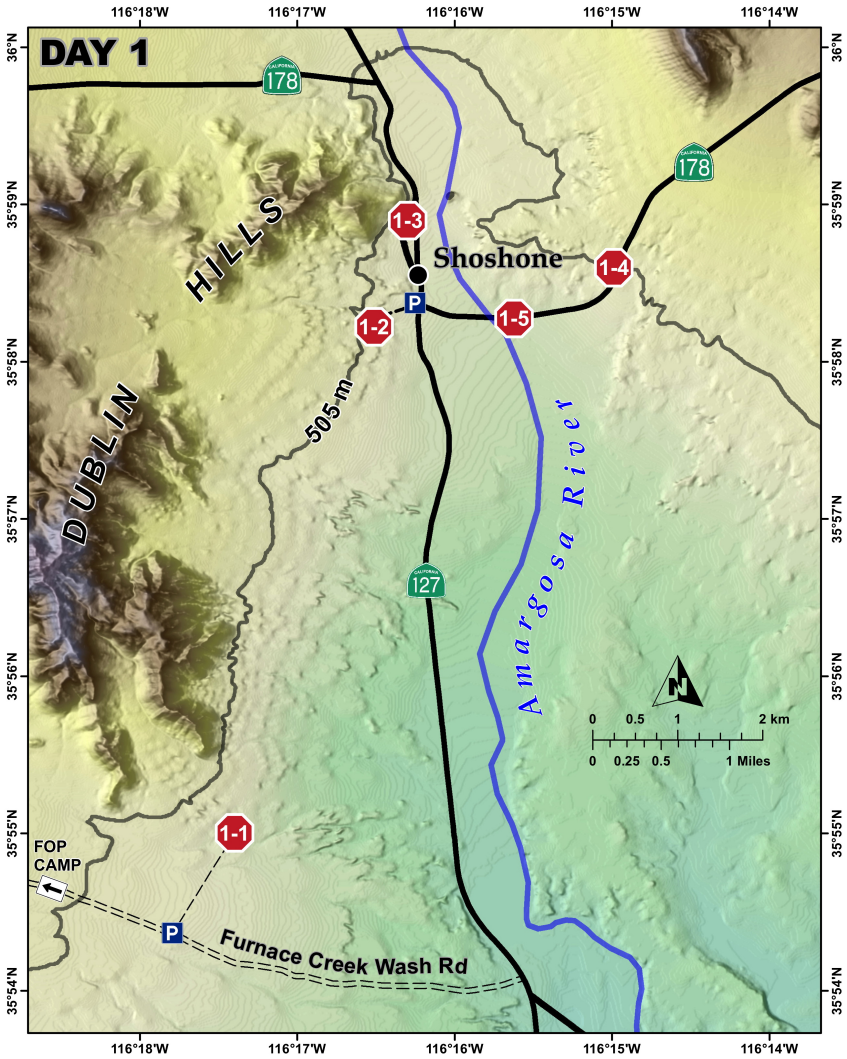


Figure 1.1. Field trip stops for Day 1; northern Tecopa basin. The dashed line from P to Stop 1.1 shows the approximate 2.5-km roundtrip traverse route to the Dolomite Hill area.

Mileage		ROAD LOG DAY 1
0.0		**Reset odometer on Furnace Creek Wash Rd at camp turnoff**
		At 9:00 the Dublin Hills (just N of camp); east-dipping upper Precambrian to Middle Cambrian strata (from lowest to highest): the Stirling Quartzite (at base); Wood Canyon Fm. (thick brownish unit); Zabriskie Quartzite (thin unit of pink quartzite), Carrara Fm. (reddish-brown or rusty), and Bonanza King Fm. (banded, grey carbonate). Overlying volcanic rocks west side of the Dublin Hills are Shoshone Volcanics (8.5-7.5 Ma).
1.0		At 10:00 on east side of basin, the same miogeoclinal strata form the Resting Springs (foreground) and Nopah (skyline) ranges. Nopah Peak is the highest of the two peaks separated at ridgeline saddle.
2.6		**Park area for Stop 1.1 presentations and traverse** - Please park single file along the right side of road, leaving plenty of room for vehicles to pass. After introductory presentations near the vehicles, we will be hiking to the gray dolomite knob ~1.3 km to the NNE (heading ~028˚) to look at Lava Creek-aged beach deposits at 488 meters above sea level (masl), and younger, higher shoreline deposits northwest of the knob. The hike is about 2.3-km-long round-trip. It is not a rugged hike. We plan to be away from the vehicles for about 2 hours.

Stop 1.1 Traverse to Dolomite Hill (There will be presentations near the vehicles then a ~2 hour round-trip traverse north to Dolomite Hill)**:**

- Introductions (**Caskey**)
- Previous research and current efforts related to Tecopa basin stratigraphy, lake -level history, and integration of the Amargosa River watershed (***Reheis, Bright, and Caskey; Appendix A1.1, A1.2***)
- Tectonic setting of the Tecopa Hump (***McMackin and Menge;, Appendix A1.3***)
- Summary of new observations and preliminary interpretations regarding the middle-to-late Pleistocene history of high-level lakes in the Tecopa basin (***Caskey, Reheis, and Bright; Appendix A1.4***)

Stop 1.1. Traverse to Dolomite Hill (Reheis): We are on the Greenwater fan draining southeast from the Greenwater Range and Ibex Hills. There are excellent views of most of the Tecopa basin from this area, so it is a good place for an overview and introduction. Views include, from north to south: the Dublin Hills, dissected deposits of the Tecopa basin between the Amargosa River and the Resting Spring Range, spring tufa mounds along that range front, the Hot Spring Hills and the Tecopa Hot Springs area, the Nopah Range, the Sperry Hills, the Amargosa River Canyon, and Tecopa Peak. In good light, one can see the highest-level strandlines (508 masl) observed in the southern part of the valley, which are of middle Pleistocene age. To see these, look along the lower bedrock flank of Tecopa Peak and just below the summit of the Hot Spring Hills and around the western side. On the northeast side of Tecopa Peak, 508-m shoreline

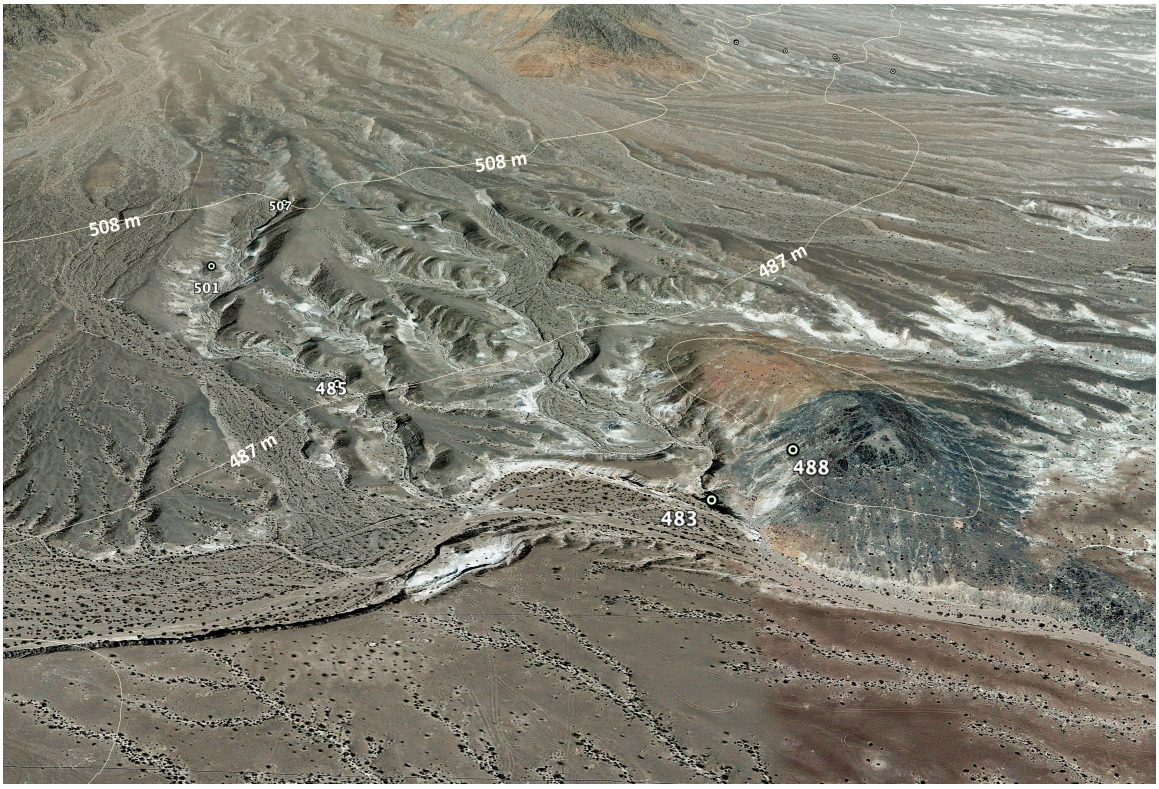


Figure 1.2 Aerial-oblique, Google Earth (GE) image of the vicinity of Dolomite Hill (Stop 1.1). View is to the NNW. 487 m and 508 m contours are shown for reference. For an approximate scale the distance between the 483 m mark at the base of the hill and the 501 m marker (up the deep, left incision) is about 500 m.

deposits are exposed in stratigraphic section above older shoreline deposits at 487 m that contain LC ash (0.64 Ma). Shoreline deposits of the *LC lake* are well exposed along the dissected fan piedmont on the north side of Tecopa Peak, and are visible from afar as the conspicuous white line that stands out from the surrounding darkly varnished fan units. We refer to these two Lake Tecopa phases simply as the *LC lake* (Lava Creek Lake) and the *High lake* (Caskey and Reheis, Appendix A1.4). We will walk due north to the southeastern corner of the Dublin Hills across fan deposits of Greenwater Wash, mainly mapped as the older two post-Lake Tecopa gravels by Morrison (unpublished mapping), and passing by a projecting knob of Miocene basalt en route. The basalt knob exhibits conspicuous strandline-like features that lie at the same elevation as the shoreline deposits we will see at Stop 1.1. Our target is an arroyo cut at the foot of the Dolomite Hill of middle Cambrian Bonanza King Formation (Figure 1.2). In the arroyo are exposures of carbonate-cemented beach gravel beneath Lava Creek ash and capped to the west by well-bedded gravels that may represent a fan-lake interface setting. On the southwest flank of the Dolomite Hill cemented dolomite-cobble beach gravels reach a thickness of 2-3 m thick. These gravels are plastered (i.e., cemented) onto Cambrian bedrock and capped by thin-bedded Lava Creek ash that can be traced to the west several meters down the paleoslope. The altitude of the highest beach deposits is 488 masl, identical to the highest *LC lake* deposits observed at several locations on the north and northeast sides of Tecopa Peak at the south end of the lake.

If there is time on the traverse, we will continue ~0.7 km northwest to see the southern Dublin Hills section of the higher, *High Lake* deposits. Along the way, northwest-trending arroyos expose oscillation-ripple-bedded Lava Creek ash followed by about 10 m of overlying gravel and interbedded sandy deposits with abundant root casts that Caskey and Reheis interpret as mainly groundwater-discharge (GWD) deposits. Farther up the wash and stratigraphically above the GWD, are about 5 m of green ostracode-bearing muds (including *Limnocythere sappaensis*) and beach sand and gravel that extend up to 507 masl. These post-LC lake deposits represent the highest, but not the youngest, of the high-level lakes we observe in the Tecopa basin.

		Return to vehicles. Continue down FCW Road. Proceed to Stop 1.2.
4.6		Turn left (north) onto CA 127. Proceed north for 5 mi. to **Stop 1.2.**
5.0		The basin deposits on the left contain a tephra (near the base of the upper resistant ledge) that was originally thought to be correlative to Tuff C mapped elsewhere in the basin (Sheppard and Gude, 1968; Hillhouse, 1987). A. Sarna-Wojcicki (unpubl. data) identified tephra mapped as Tuff C in the eastern side of the basin as the 2.1 Ma Huckleberry Ridge (HR) tephra, and so Tuff C was considered everywhere to be HR tephra. New $^{40}Ar/^{39}A$ dating by Gary Scott and others (Berkeley Geochron Center) now show that what was mapped as "Tuff C" here in the western part of the basin is actually 1.25 Ma. The new tephra age and implications for revisions to basin stratigraphy will be the focus of discussion on Day 2, Stop 2.2.
5.5		The well-rounded cobble conglomerate exposed along the road for about the next 0.8 mi (430-435 masl) resembles the coarser parts of middle Pleistocene Amargosa River gravels at the north end of the basin and also broadly channelized gravels that extend down the axis of the valley ~1 km to the east, that we refer to as *the axial-channel gravels*. However, in detail, the mostly fluvial gravels in this area locally appear to be reworked and somewhat restacked by waves along "an ancient" Lake Tecopa shoreline.☻
6.3		On the left: sandy clay deposits that directly underlie what are interpreted to be lacustrine sandy gravels at 435 masl yielded a variety of ostracode species including *Candona rawsoni*, *Cyprideis*, *Limnocythere paraornata, Cypridopsis vidua,* and *Potamocypris* (samples R053120A and TB14-20A). Collectively these varieties suggest a cold wetlands envirionment. The presence of *Cyprideis* indicates low ALK/Ca and not water brought in by the Amargosa River. A nearly identical variety of species were sampled from deposits at 430 masl that overlie the axial channel gravels 1-2 km north of Tecopa Hot Springs (sample site T14-52). Don't quote the Road Log on this!
7.0		Basin deposits exposed west of the road in this area include the 0.76 Ma

		Bishop ash (Tuff B of Sheppard and Gude (1968) and Hillhouse (1987)) and the 0.64 Ma Lava Creek (LC) ash (Tuff A). The rusty-orange weathering interval marks the LCB ash. A thick dark green section of highly altered mudstone and sandy mudstone lies between the two tephra.
8.0		At 1:00, the Butte-like hill in front of the Resting Springs Range is composed of tuffs of the Shoshone Pass volcanics (9.6±0.1 Ma, Calzia and Rämö, 2000). FYI, there are some very accessible roadcuts with remarkable exposures of the transitions through a complete cooling unit of a welded ash flow tuff (i.e., ignimbrite).
8.5		Approaching town of Shoshone, CA (12:00); skyline shows east-tilted, middle Cambrian carbonates the west and northwest are overlain by similarly tilted, dark brown basalt (12:00) the "DV basalt," which has been dated at 9.3±0.3 Ma (Calzia, unpublished whole-rock K-Ar date, 1994, in Calzia and Rämö, 2000). The DV basalt is thought to correlate with subhorizontal basalt flows of the same age found locally along the crest of the Resting Spring Range to the east. If so then the age of rotational block faulting and west-directed Miocene extension across Amargosa/Tecopa Valley is constrained to be younger than 9.3 Ma. Major extension across the greater southern Death Valley region initiated in the Kingston Peak area at 14-13 Ma. Extension generally migrated from east to west through time. The period of large magnitude extension across this part of the southwestern Great Basin was essentially over by about 7-5 Ma (Wright et al., 1984; Calzia and Rämö, 2000).
9.0		The Tonopah - Tidewater (T-T) Railroad grade is visible less than 100 m right (east) of the highway. The Tonopah & Tidewater railroad was a subsidiary of the Pacific Coast Borax Company. The standard gauge railroad was built during 1905-1907 between Ludlow (San Bernadino Co.) and Beatty, NV where it connected with the Bullfrog-Goldfield Railroad, which in turn, tied into several other Nevada railway lines in Esmeralda and Nye counties. It served the gold mines and other mines in the region through most of the 1930's and was eventually dismantled for scrap in 1942 during the war effort. (Source: http://mojavedesert.net/railroads/railroads-051.html)
9.2		The T-T Railroad cuts to the east expose large, south-dipping, sandy-gravel foresets interpreted as Amargosa River-derived deltaic deposits. The sediments have not been dated, but they were likely on move on the move during one of the two middle Pleistocene lake trangressions we are able to document between about 0. 64 Ma and 0.50 Ma.
9.5		Just as you pass by the turnoff on the right to CA Hwy 178 W (toward Pahrump,NV) pull into the large graded area that comes up immediately to

		your left (west). **This is the parking area our Stop 1.2. traverse into Dublin Gulch**. We should get assembled before departing for an ~2.0 km round-trip traverse to the pumicite quarry to look at spectacular liquefaction features exposed in the ash beds among other things. There will be one or two stops along the traverse for short presentations and discussion. Our **Lunch Stop** will be **when and where we park for Stop 1.3,** which is only a five-minute drive (less than a mile) from **Stop 1.2**.

<u>Stop 1.2 Dublin Gulch Traverse (Scott) (Larsen)</u>

The destination of this traverse is to the pumicite quarry exposures investigated by Gibert et al. (2011) (**Appendix A1.7**). Larsen (2008) has studied the stratigraphy of the Tecopa beds throughout much of the basin. Our traverse to the pumicite quarry is in the area described in his stratigraphic section LT-10 (**Appendix A1.5**).

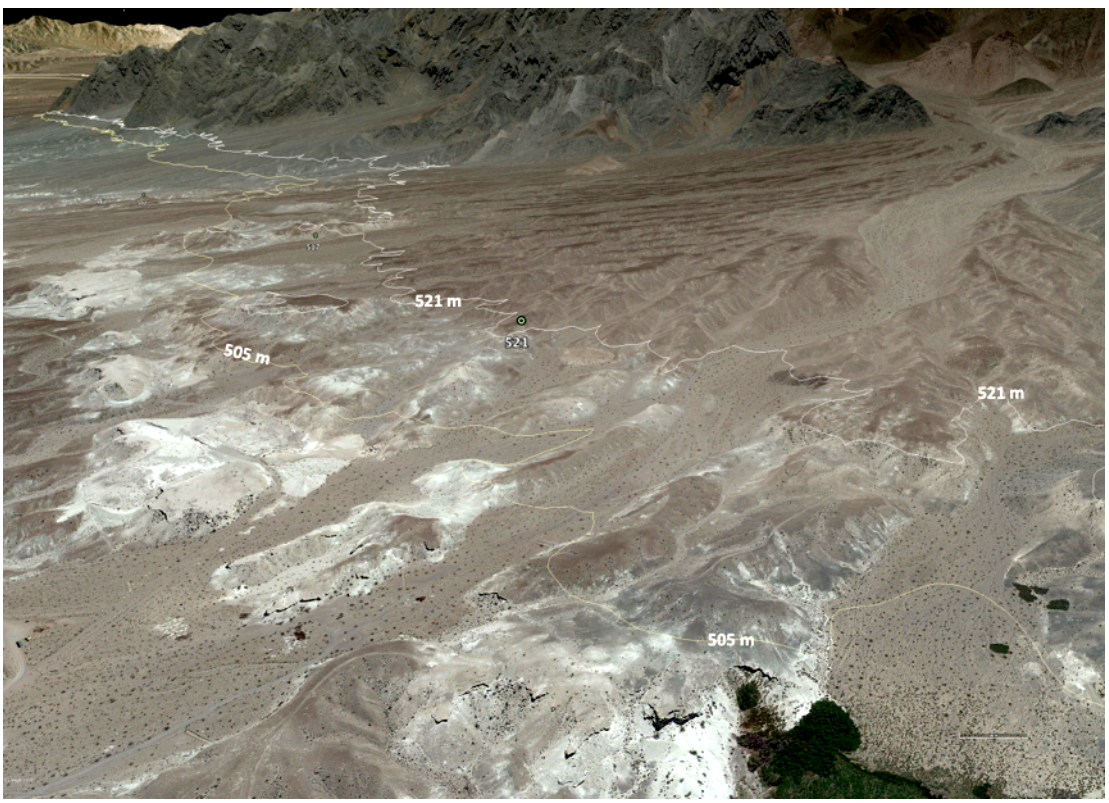


Figure 1.3. Aerial-oblique GE image of the Dublin Gulch area (view ~ S). Shoreline deposits of both the *LC Lake* and the *High Lake* are exposed in this area, although much of the LC-Lake deposits are quarried away. There are some interesting exposures in the area of small round waypoint marker (labeled 521 m) where the fan terminates along 521-m-shoreline(s) of the High Lake.

Optional stop on return from pumicite quarry

For those interested, from the pumicite quarry we can continue a few hundred meters west and 'up section' relative to the alluvial fan stratigraphy (Figure 1.3). In this general area we've been able to track rounded beach gravel up to about the base of the scarp at about 521 masl, which we take as representing the highest elevation of the *High Lake* here in Dublin Gulch. But is this a wave cut scarp? Note that on the image on (Figure 1.3) there is another smaller scarp expressed higher up that marks a small but obvious step in the upper fan surface.

Another interesting point to make here in Dublin Gulch is when these local fan gravels become reworked into well washed and rounded beach gravel they look nearly identical to some of the fluvial-deltaic gravels that were transported into the basin through Amargosa Valley.

Important to our story of the paleohydrologic environments in the basin are the ostracodes that we collected and that Jordon processed and identified. In at least one place, muddy sand within these beach deposits contained abundant *L. sappaensis*, indicating a lake supported by Amargosa River water, and lower down contained *Candona* that live in groundwater-discharge deposits. Its also important for us to keep track of where we find shoreline of different highest shorelines we find in Dublin Gulch are at 521 masl; 16 m higher than what we called *High lake* deposits we find only up to 505 masl northwest of the dolomite hill at Stop 1.

		Return to vehicles. Continue north into Shoshone on Hwy 127.
9.6		Just as you pass the Chevron Station, Charles Brown Market, and the Shoshone Inn, veer left off of Hwy 127 and onto Old State Hwy.
9.9		Continune past Death Valley High School (on left and), home of the Scorpians.
10.0		The Shoshone Trailer Park is on the right.
10.1		There is a large parking area to the right. This is **Stop 1.3 and it's Lunch time.**

Stop 1.3 'DV Basalt' strandlines (Caskey and Reheis) (2 hrs. including lunch)

The in place see looking east About 200 m northwest of the parking area are remarkable exposures of cemented lenses(?) of Lava Creek ash glued to and around large colluvial boulders of basalt. They are associated with scattered small well-rounded pebbles, some flattened discoid shapes. Caskey and Reheis interpret this feature as a deposit along the shore of the Lava Creek-aged lake, and probably at about the lake high-stand based it's elevation of 508 masl. It's 20 m higher than the LC cemented shoreline deposits we saw at Dolomite Hill this morning! About 10 or so meters north of where you lose the whitish tephra, and about 5 m higher in elevation, there horizontal ledge of really thick, and really dense tufa at an elevation of ~514 masl. If its a shoreline tufa then logically could represent a strandline of the younger *High Lake*. As discussed at Dublin Gulch we think we have the *High Lake* at up to 521 m just around the corner from here. We will wait until Stop 1.4 to talk about the details of the stratigraphy across the northern part of the basin because we have excellent views up there, and where the

sediment isn't eroded out by the river. There is an excellent road-cut exposure of very pristine and very coarse Bishop ash. You will also see alot of well-sorted and rounded, pebble gravel in this area and across the entire width of the valley.

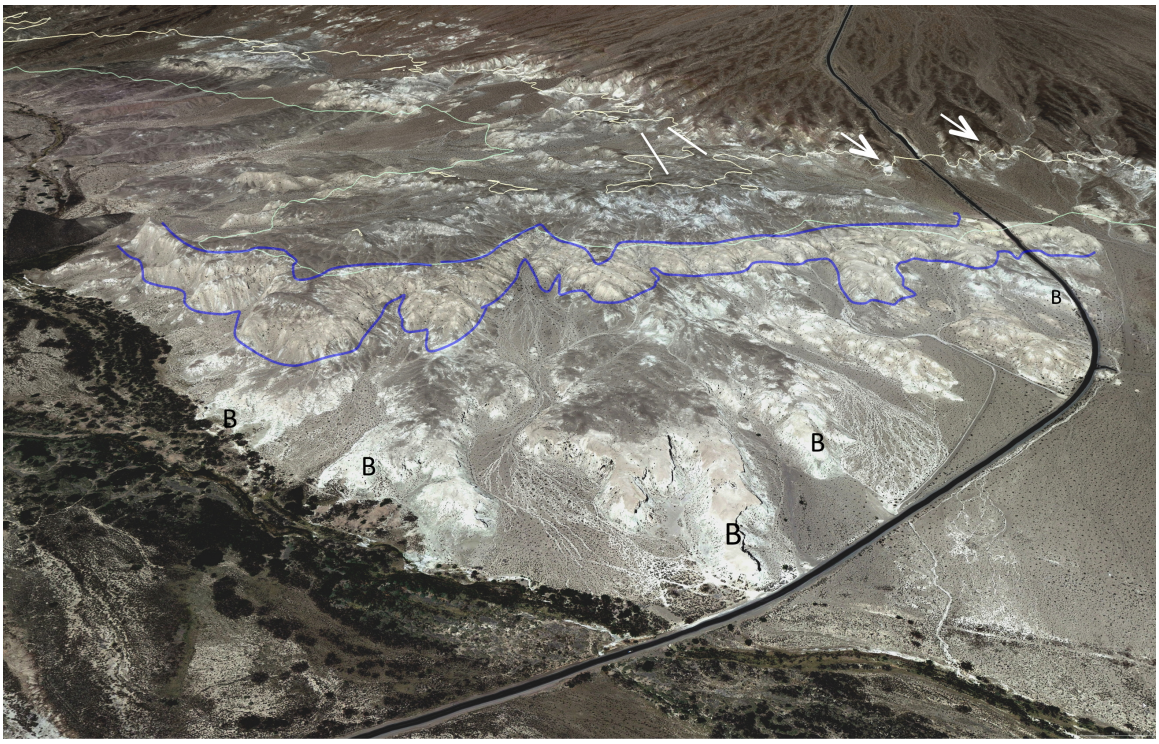
Relative uplift(?) of the LC and *High Lake* shorelines between Dolomite Hill and Shoshone (Caskey and Reheis):
The elevation of the highest lake gravels in the Dublin Gulch area is 521 masl, identical to the highest elevations of *High Lake* deposits 3.5 km to the south on the east flank of the southern Dublin Hills. About 2 km east of Shoshone at Stop 1.4, *High Lake* beach gravels are found at up to 524 masl, 16 m higher than the highest *High Lake* deposits we observed near the Dolomite Hill (Stop 1.1) and near Tecopa Peak at the south end of the basin. At Stop 1.4 we will discuss the various lines of evidence that support a basin-wide correlation of high-level, post-LC shoreline deposits and then our observations that indicate ~21 m of relative(?) north-side-up displacement of *LC Lake* deposits in the northern part of the basin since ~0.64 Ma.

10.1	**0**	**Return to vehicles. Head back to Shoshone enroute to Stop 1.4**
	0.6	Turn left onto CA Hwy 178 E (toward Pahrump)
11.3	**1.2**	river crossing Castle in the Clay location of Stop[%
	1.7	pass by turnoff to the landfill
	1.8	**road cut thru bishop**
12.1	**2.0**	**turnoff to Stop 1.4**

Stop 1.4: High beach gravel site(s) and traverse through High Lake delta (Caskey, Reheis, Larsen, and Bright) 1.5 hrs.
This is a great overlook and a part of the basin where there has been alot of work done. After a review of some of this work, we will argue the case for age correlation of the highest two shorelines we see throughout the basin, and comparisons of lake level curves of Morrison (1999) and Larsen (2008), which includes a review and some scrutinizing of some of Roger's conclusions. Then we shall take a traverse through what hope what appear to be nice deltaic gravels and sands.

(Reheis) To the southeast of the highway are extensive exposures of Tecopa basin sediments as old as ~1 Ma dissected by drainages emanating from the Resting Springs Range. These outcrops have been studied by many researchers; previously, they were particularly well exposed in the now-closed Shoshone landfill. Hillhouse (1987) identified the Brunhes-Matuyama chron boundary in deposits between the Bishop ash bed and an underlying bed of upper Glass Mountain tephra. Morrison (1991, 1999) described sections here and identified lacustrine deposits associated with the Lava Creek-aged lake. Perched above these deposits and locally

interbedded with them are lenses and mounds of spring carbonates. Larsen (2008) measured detailed sedimentary sequences south of the highway as section LT-8 and 8B. More recently, Caskey, Reheis, and Bright have measured, sampled, and analyzed numerous sediment samples from this sequence. Most valves were too fragmentary to identify. In this corner of the basin *L. sappaensis*, representing a lake fed by the Amargosa River, has only been found in the interval immediately below and above the Lava Creek ash. Sediments above the Bishop ash and those higher than about 500 masl above the Lava Creek ash contain a few ostracodes representing springs and groundwater discharge.



`. Upper and lower fluvial-deltaic sequences (dark deposits) east of Shoshone and north of Hwy 178. The orangish weathering unit separating the fluvio-deltaic sediments, is in part ground water discharge deposits, and not well washed fluvial gravels. White arrows and lines point to the highest-level beach deposits we have observed in the basin.

Larsen: Some sedimentary features that we may down from and south of the road) see include a thin interval of Lava Creek B filling a fluvial channel and interfingering of distal alluvial fan deposits, alluvial and eolian alluvial flat deposits, Amargosa channel and floodplain deposits, and thin marshy lacustrine muds. Larsen (2008) sections LT-8 and -8B extend from beds correlative to the Glass Mountain ashes south of the landfill northward to a gully south of CAL 178 highway containing an ash tentatively correlated by Morrison (1999) to the ~0.2 Ma Wadsworth(?) ash. The strata in sections LT-8 and 8B comprise silty to fine sandy distal alluvial fan to alluvial flat facies interfingering with sandy to pebbly Amargosa floodplain and channel facies, with several thin greenish-gray lacustrine margin mud intervals. Lacustrine intervals in these sections generally have root traces throughout, suggesting marshy lake margin conditions. The facies distributions in section LT-8 and -8B follow predictable Walther's Law

relationships and lack evidence of coarse-grain fan progradation into the basin, such as that observed in section LT-10. The post-Bishop ash lacustrine interval is correlative to lacustrine incursions at LT-1, LT-10, and LT-2; however, the thick Lava Creek B lacustrine interval seen at section LT-10 is not present at LT-8 and is only hinted at in section LT-1. These observations support the inference of basin tilting or possibly movement along a fault on the western central part of the basin, focusing lacustrine deposition in that part of the basin. The series of observations associated with correlation of the sections at LT-1, 8, and 10 suggest caution needs to be applied in using elevation of shorelines to correlate highstand lacustrine intervals in the Tecopa Basin.

North of the highway is a climbing sequence of interbedded fluvial and deltaic deposits of the Amargosa River. We will walk through this sequence to examine the complex facies changes and argue the interpretation of fluvial vs. deltaic deposits. Sediment samples collected in sandy mud deposits at or near the very top of the section in this area contained *Candona*, *Cypridopsis*, and a complete *Limnocythere staplini*, which complicates the story that the sediments themselves tell us about the environment. It's worth discussing.

Optional stop 5:

Larsen: At optional stop 5, the lower part of section LT-1 is exposed (Larsen, 2000). Some features that we may see include Amargosa River channel sands, subaerial deltaic sands, floodplain deposits with weak paleosols, several ash beds, enigmatic tufa blocks above the Bishop ash, and greenish-gray lacustrine muds. This section largely records sedimentation from the ancestral Amargosa River with a prominent lacustrine incursion above the Bishop ash and below the Lava Creek B ash.

Considering the observations at each stop in the north-central Tecopa basin, the preserved Pleistocene stratigraphy shows influence of Amargosa River discharge and sedimentation throughout. The lacustrine intervals suggest that the maximum lake level and extent increased upward through the section; however, the distribution of indicators of lake expansion varied due to tectonic influences.

Day 2: Down Furnace Creek Wash (FCW) Road to Tecopa

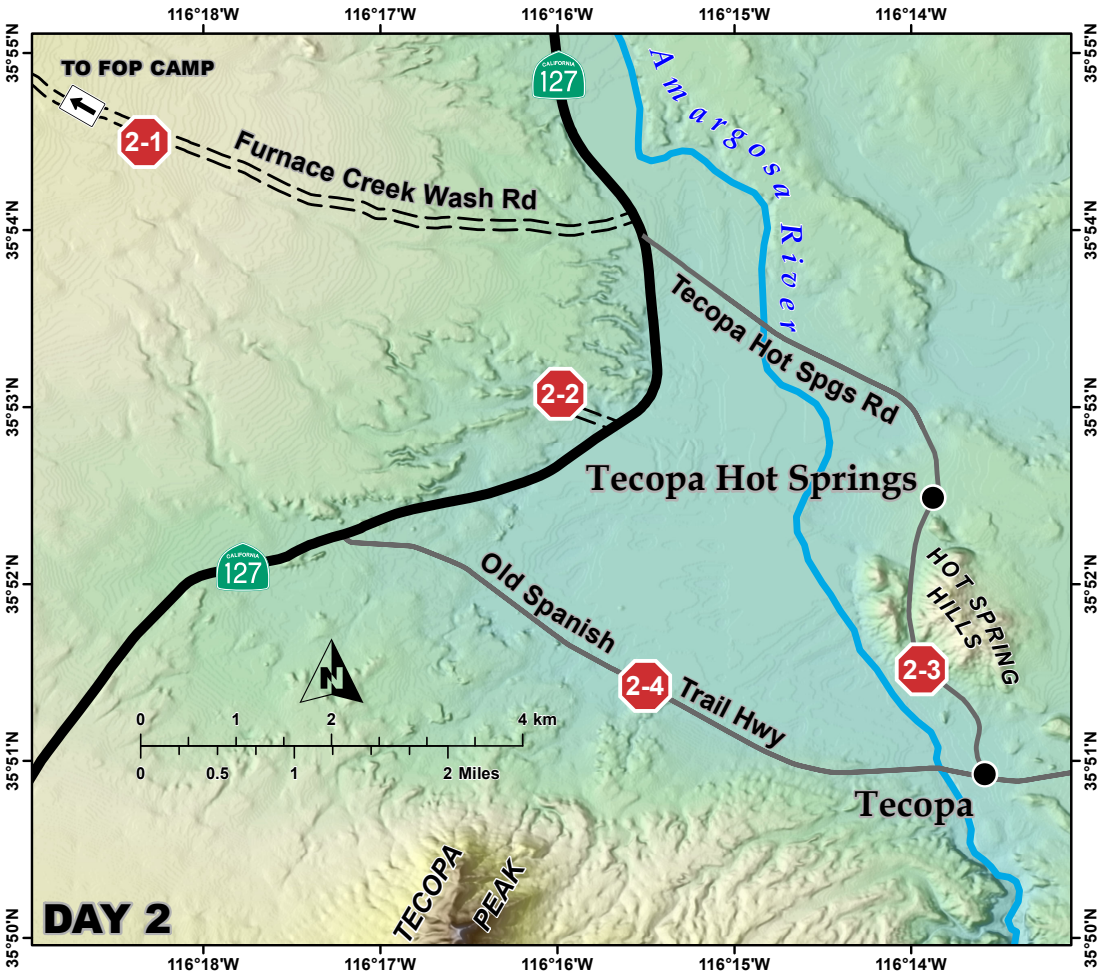


Figure 2.1. Trip map for Day 2 Stops. BTW the beautiful field stop maps for this trip maps were made by John Niles who many of you know. Thanks John! They are really nice.

Mileage		ROAD LOG DAY 2
0		reset odometer at pullout from camp
2.0		**Stop 2.1.**

Stop 2.1. What we will see: 100 m south of the road is a southwest-facing arroyo cut that exposes 0.5-1.5 m of well washed and well bedded pebble gravel that is inferred to represent a beach gravel. This unit extends about 150 m continuously along the arroyo cut and overlies brown fine-grained deposits interpreted as groundwater discharge deposits interbedded with thin fan gravel. On the western end of this outcrop is a basal rounded gravel with backset (NW-facing) beds overlain by a westward-thickening, NW-dipping bed of sand (sampled for ostracodes but none were found). Locally, gravels are cemented by carbonate that resembles lake tufa. Caskey and Reheis tentatively interprets this site as evidence for a lake younger than

the post-Lava Creek highstand that lies stratigraphically about 15 m above the ash bed; this younger lake extends only up to ~505 masl and thus would be inset within the older sediments. We suggest that during some previous highstand, Lake Tecopa may have discharged over the southern threshold and eroded it from about 510 to 505 masl, thus accounting for this younger inset highstand. We have not yet been able to date this beach gravel; possibly it is of early OIS 6 age.

		Return to Vehicles and proceed down fan to Hwy 127 where you will turn right (south).
4.6		**Turn south** onto Hwy 127
5.9		**Turn right** into the broad drainage that has come to known in our circles as "Camp Wash"
		Stop 2.2. Please try to park toward the south side of wash to keep the outcrop open and visible for the group.

Stop 2. Camp Wash. **Early Pleistocene basin history (Scott, Larsen, Reheis, Bright)**
2-3 hours

What we will see: This large drainage we call "Camp wash" for its common usage has excellent exposures from near the highway to as much as 5 km to the northwest. To the east, deposits are exposed below and above what has been mapped as "Tuff C" by all previous workers (Sheppard and Gude, 1968; Hillhouse, 1987) and correlated to outcrops of "Tuff C" that lie 6-8 km to the ENE that have been identified by glass chemistry as the Huckleberry Ridge ash (~2 Ma; Hillhouse, 1987; Sarna-Wojcicki et al., 1987, and later publications; Morrison, 1991, 1999; Larsen, 2008). Stratigraphic sections have been measured by all of us and previous workers through this sequence and up through the Lava Creek ash along this and adjacent drainages. In addition, paleomagnetic data have been collected and analysed by previous workers (Hillhouse, 1987; Larson et al., 1991). However, the chronology and the interpretations of all this previous work must be reconsidered in light of new data obtained on "Tuff C" at this location in Camp wash and nearby. This tephra contains euhedral biotite, which was not previously recognized, and therefore the tephra cannot have been derived from the Yellowstone eruptive centers. New dating by Al Deino and Gary Scott (Berkeley Geochronology Center) indicates that this tephra is ~1.25 Ma. The tephra layer mapped as "Tuff C" along the east side of the Amargosa River also contains biotite phenocrysts and is likely the same 1.25-Ma tuff. Reheis has carefully traced the "Tuff C" identified as correlative to the Huckleberry Ridge ash from its relatively unaltered outcrops 8 km to the ENE as far as possible to the west, and found that this tephra layer dips gently west and disappears below the surface about 2 km NE of Tecopa Hot Springs.

Reheis and Bright sampled for ostracodes here and also in the same sequence along the east side of the Amargosa River. At this location we found *Limnocythere* fragments just below the tuff; along the river, at a lower altitude, we found *L. staplini* just below and above the tuff. *L. staplini* can live in lakes but require low ALK/Ca water, so the shallow lake associated with the 1.25-Ma tuff was not fed by the Amargosa River. No other ostracodes were found in the other

parts of either measured section, with the exception of elliptical *Candona* valves in one sample below the Bishop ash that probably indicate a groundwater discharge setting.

Larsen: Camp Wash lies between Larsen (2008) sections LT-3 and LT-7. The section below Tuff C is primarily light greenish gray to pinkish gray lacustrine mud with varying amounts of diagenetic pores and molds after gaylussite. Tuff C is white to green and gray and shows extensive diagenetic alteration as well as diagenetic pores after gaylussite. Facies above Tuff C are generally light gray to pinkish gray and include wave-rippled tuffaceous and epiclastic sands interbedded with cherty mudstone, all of which are interpreted to record shallow lake to playa conditions. Of significance, several ash beds are present in section LT-3 that show ball and pillow and other soft-sedimentary deformation. The enclosing muds show no evidence for such deformation, except to provide a medium to founder into. The deformation of the ash beds is not eroded at the top or unconformable, suggesting the deformation is associated with sediment loading rather than contemporaneous seismic activity (e.g., Gibert et al., 2011).

Silicate diagenesis is evident in all samples from sections LT-3 and -7 (Larsen, 2008). Remnants of zeolite alteration are present, but most of the alteration is present as replacement of detrital clays and early-formed diagenetic phases by extremely fine-grained (< 10 micron) potassium feldspar, illite, and searlesite, with cristobalite being present in cherty beds. The diagenetic history is interpreted to have involved an initial period of syndepositional alteration of detrital and volcanic ash debris by zeolites, trioctahedral smectite, and perhaps magadiite, followed by later fluxing of hypersaline brine through the sediments to form the potassium feldspar, illite, and searlesite. Alteration of volcanic ash in saline, alkaline-lake waters is well documented, both in modern environments (Lake Magadi: Eugster, 1970; 1980; 1986) and ancient deposits (Tecopa: Sheppard and Gude, 1968; Pliocene Lake Olduvai: Hay, 1976; Hay and Kyser, 2001). However, the influence of fluxing brines through buried lake sediments may require specific conditions, as does the formation of searlesite, a borosilicate mineral (Larsen, 2008).

Lake level and extent prior to Tuff C are interpreted to have been low and localized to the south-central basin. Lake level decreased following Tuff C for an extended period, but renewed expansion of the lake prior to upper Glass Mountain ash falls at 1.0 Ma is evident based on extent of lacustrine mudstone at sections LT-6 and -6B (Larsen, 2008). The new age assignment for Tuff C at 1.25 Ma solves a previous problem of low sedimentation rates (or even deflation) in the basin prior to emplacement of the upper Glass Mountain ashes (Morrison, 1991, 1999; Larsen, 2008). Tuff C was initially correlated to the Huckleberry Ridge ash at 2.1 Ma, leaving only 3 to 5 m of sediment deposited in the basin over approximately 1 million years. The new age assignment for Tuff C suggests sedimentation rates were approximately constant between 1.25 and 0.76 Ma. Furthermore, identification of the Huckleberry Ridge ash in the southeastern part of the basin (Reheis and Caskey, pers. comm.) suggests that the sedimentation record between the Huckleberry Ridge ash and Tuff C needs further exploration.

Little evidence for significant Amargosa River discharge and sedimentation is evident prior to the deposition of the upper Glass Mountain ashes. However, a lacustrine bar deposit at section LT-4B between the two prominent upper Glass Mountain ash beds contains rounded, compositionally diverse clasts; similar rounded diverse clast compositions are observed in sheet delta sands at section LT-1. The increase in lake level and appearance of the rounded, diverse clast compositions suggests that increase in lake level at approximately 1.0 Ma was related to

increased discharge from a more integrated Amargosa River drainage system. This interpretation contrasts with that of Menges (2008), who places drainage integration of the upper Amargosa River with the Tecopa basin at 150 – 100 ka based on geomorphic observations.

Scott has measured paleomagnetic properties upward from the 1.25-Ma tuff in Camp wash to the Bishop ash bed and has identified the Jaramillo event. In addition, numerous examples of soft-sediment deformation can be viewed along this and the adjacent drainage.

Stop includes lunch. Divide up for short walks up drainages to view paleomagnetic evidence for Jaramillo subchron (north fork) and seismites that involve the Bishop ash bed (south fork).

0.0		Return to vehicles and reset odometers when you get to Hwy 127. Turn left north.
1.2		Make hard right turn to head southeast toward Tecopa Hot Springs. within about the first mile you get great views of the Southern Nopah Range (10:30), Kingston Peak at 11:00 and at 12:30 is the Alexander Hills.
2.2		Crossing the Tonopah Tidewater railroad grade
2.8		On the left is the trailhead for the rustic hot springs. They are really nice springs but widely known about anymore.
		Proceed through the Tecopa HotSprings area and beyond to Stop 2.3. The bedrock you are going through is the Precambrian Stirling Quarztite

5.0		**Parking for Stop 2.3** might have use a stretch of road. Please be careful along the road.

Stop 2.3 Grimshaw Lake (Caskey, Reheis, Larsen, McMackin) 1.5 hr?

(Caskey) Marith had listed in the draft outline for the trip a number of items for discussion at this site. Bring them all up and open it up for discussion. The image below might get some thoughts rolling.

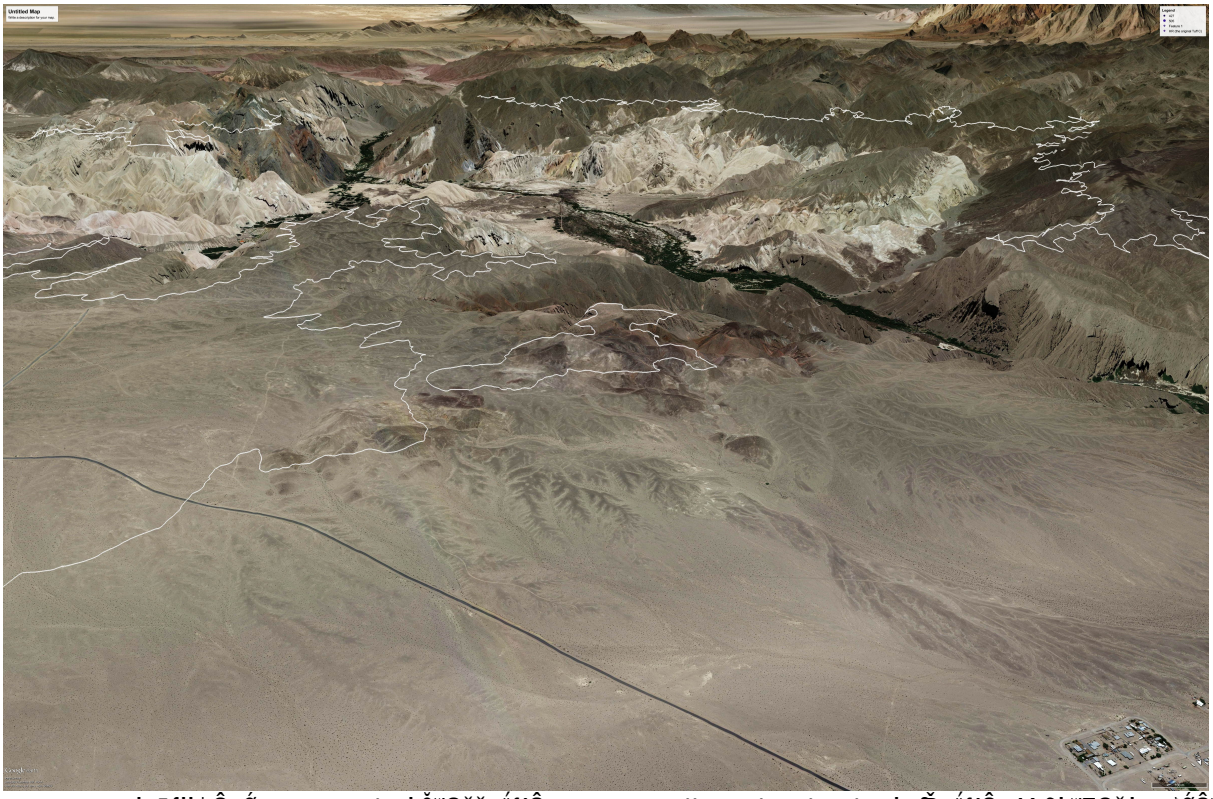


Figure 2.2. hōfljdzŜ Øiew south l·ŎŃƧăă ŭKŜ Sperry Hills paleodivide l·yŘ ŭKŜ !Y l·ŃƎƧăl· włØŜŃ /l·yèƧy. !y ŜÈdzKdzYŜŘ bedrock pediment óYłŘŘŧŜú łă ŜÈLJƧăŜŘ ły ŭKŜ ŧƧøŜŃ l·ŃŜl·ă ƧŦ l· YƧŃŜ ŎƧyŭłyŧdzƧdză dzLJLJŜŃ ŜŃƧăłƧyl·f ădzŃŦl·ŎŜ ƧŦ {LJŜŃŃè łłŧf ŘłØłŘŜ ŭKl·ŭ ŜÈłăŭŜŘ LJŃłƧŃ ŭƧ ŘƧøyŎdzŭŭłyƎ ƧŦ ŭKŜ Ŏl·yèƧy¢ ¢KŜ lowest LJƧłyŭ on that surface łă now рлу Y ó ŎƧyŭƧdzŃ ŦƧŃ ŃŜŦŜŃŜyŎŜú low enough to represent the spill-point level of the *High Lake* phase. The brief duration of the *Terminal lake* at 505 m suggests this lake breached from an only slightly lowered spill point of the High Lake phase.

Larsen: At this location we can see several features of interest regarding Pleistocene shoreline features in the Tecopa basin and modern spring activity. Looking northeast, the top of the Tecopa Hills has a pointed apex with beveled margins at all sides. The beveled margin is interpreted as a wave-cut terrace at 515 m. Rounded to angular boulders of Precambrian Stirling Quartzite lie on the wave-cut surface. Tracing along the Tecopa Hills skyline to the east, a prominent gravel beach deposit, cemented by $CaCO_3$, is present at 498 m near the fault contact between the Precambrian Stirling Quartzite to the north and Wood Canyon Formation to the south. Rounded pebbles of both Stirling Quartzite and Wood Canyon Formation are present in the gravels. A few other potential beach features are locally observed on the Tecopa Hills, but are not as prominent as those discussed above. Looking to the southwest, high shoreline gravels are present at elevations ranging from 483 m along the north face of Tecopa Peak to 505 m along the lower east slopes of Tecopa Peak; even higher elevation (approximately 515 m) $CaCO_3$-cemented gravels and talus exist near the contact of Pleistocene Tecopa gravel and the Wood Canyon Formation along the east slope of Tecopa Peak.

The surface that we are standing on is covered by a veneer of Amargosa River terrace gravels. The gravels are rounded to subangular and represent diverse rock types in the

Amargosa River drainage. The age of the gravels is speculative, but likely correlative to one of the Late Pleistocene terraces of Morrison (1999).

In many locations around the Tecopa Hills and elsewhere in the Tecopa basin (Fig. 3X), groundwater discharge creates springs at the surface. Larsen et al. (2001) identified three end-member spring compositions based on major solute chemistry, and stable oxygen, hydrogen, and boron isotope compositions: 1) pH-neutral, $Na+Ca+Mg+SO_4+HCO_3$ water with moderate to high total-dissolved solids, 2) mildly alkaline, $Na+Cl+SO_4+HCO_3$ water with high total dissolved solids, and 3) alkaline, $Na+HCO_3+CO_3$ water with high total dissolved solids. The temperatures of most of the springs range from 17 to 32ºC; however, waters in the Tecopa Hot Spring area range from 9 to 47ºC. Type 1 waters are typified by those discharging at China Ranch, Resting Spring, and Shoshone, and appear to originate from recharge to the regional carbonate aquifer system in the Spring Mountains, Nevada. Type 2 waters are warmer and isotopically and chemically distinct from the carbonate aquifer waters. These waters discharge in the Tecopa Hot Spring area and appear to originate from fractured Stirling Quartzite, especially at the intersection of faults (Wamalwa et al., 2011), and, again, recharged in the Spring Mountains. Type 3 waters are alkaline and discharge in the central part of the Tecopa basin. Their low stable hydrogen and oxygen isotope compositions and high boron content suggest that they originate from the Amargosa River valley aquifer system and interacted extensively with the saline, alkaline lake deposits.

Detailed mapping in the vicinity of the Tecopa Hills and subsequent geochemical and isotopic analysis of springs in the Tecopa Hills has provided further constraint on the source and controls on spring discharge in the Tecopa Hot Spring area (Larsen and Sperry, 2011). The Tecopa Hills expose a faulted and fractured, largely homoclinal section (mean bedding: 360˚, dip 45˚E) of Precambrian through Lower Cambrian strata comprising quartzite, shale, and lesser quantities of dolostone. Faulting styles include an older low-angle detachment fault that is displaced by younger N-S trending normal faults and later SW-NE trending vertical faults (mean vector: 70˚). Prominent joint orientations include bedding-plane perpendicular sets in quartzite, as well as pervasive SW-NE trending vertical sets (mean vector: 81˚). Reconnaissance investigation of similar-age strata in the Resting Springs and Nopah ranges reveals similar joint styles and slightly more NE trends (mean vector 56˚ for each range). Hot and warm spring discharge in the Tecopa Hills area occurs mainly on the NE side of the hills and is aligned mainly along SW-NE trending faults, although minor seeps likely emerge from the SW-NE trending joint sets along much of the NE margin of the hills. Thermal spring water geochemistry suggests type-2, slightly alkaline, $Na+Cl+SO_4+HCO_3$ water containing moderate to high dissolved solids mixed with type-1, pH-neutral to slightly alkaline, $Na+Ca+Mg+SO_4+HCO_3$ water containing moderate dissolved solids. Noble gas composition, Na-K geothermometry, and helium isotopic composition ($R/R_a = 0.08$) of type-2 waters are consistent with deep regional ground-water flow path from the Spring Mountains along with a component of crustal Noble gas flux. The type-1 waters are interpreted to originate from the Spring Mountains along a shallower regional ground-water flow path within the regional carbonate aquifer system.

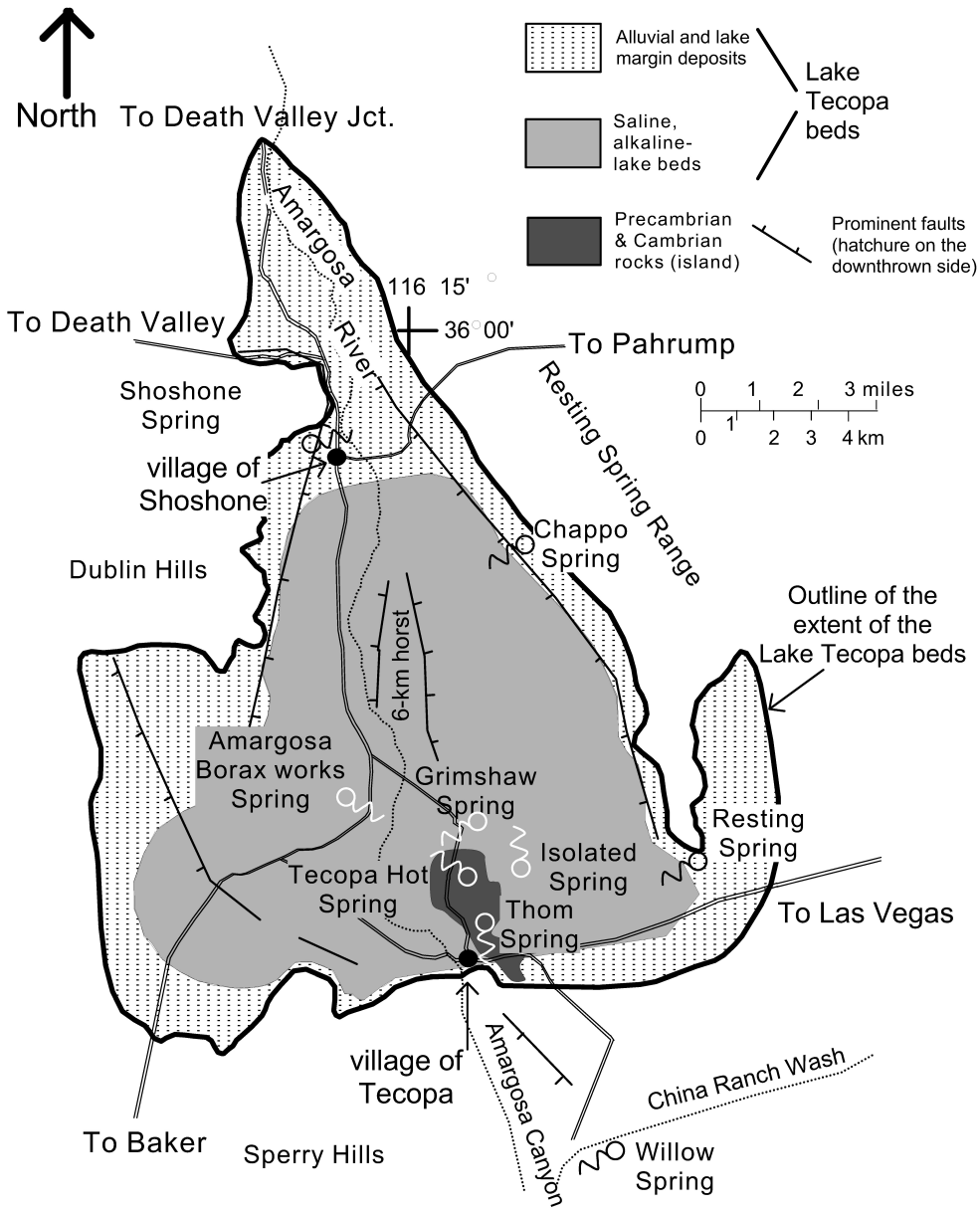


Figure 2.3: Generalized map of the Tecopa basin and environs showing the distribution of springs (modified from Larsen et al., 2001). Prominent faults are from Morrison (1999).

Reheis: The outcrop exposed on the southwest face of Grimshaw hill (informal name) contains about 13 m of fine-grained deposits unconformably overlain by about 2 m of Amargosa River gravel and sand. The fine-grained deposits consists mostly of brown sandy mud and pale brown sandy silt with a few thin sand beds and possibly diatomaceous beds. Eight sediment samples from this section contained only a few ostracodes representing groundwater taxa. Near the base is a tephra layer whose glass chemistry is most similar to upper Glass Mountain tephra beds. This deposit is interpreted to consist mainly of playa and groundwater discharge deposits whose age may be about 1 Ma. ,

Stop 4: **Sedimentology and stratigraphy of early Pleistocene deposits above and below Tecopa tuff (formerly tuff C). Implications for tectonic deformation.**(2 miles west of Tecopa on Old Spanish Trail highway. 2 hrs. Larsen and Kristian Olsen.

At this stop we will see strata below Tuff C in Section LT-9A. These strata illustrate the stratigraphic relationships between alluvial fan and fan delta sedimentation and the early lake episodes. A map of the area with the Lava Creek B, Bishop and Tuff C ashes in the Tecopa beds and structural data is shown in Figure 2.4. Sections LT-9A through 9D are shown in Figure 5X. Section LT-9A comprises a series of alternations of fan delta, braided stream and shallow lacustrine deposits that reflect delta lobe switching and variations in lake level. Also note that the dip of the Tecopa beds is greater than elsewhere in the basin, which can clearly be seen for Tuff C east of sections LT-9A and 9B in Figure 2.4. Looking toward the south and southeast from the top section LT-9A, one can see that the sand and mud with interspersed ash deposits in section LT-9A grade laterally into the coarse-grained alluvial fan deposits of the Tecopa fan. The Tecopa fan comprises several individual alluvial fan sequences coalesced into an area of largely pebbly sand and conglomeratic strata. Section LT-D is composed almost entirely of fan deposits (Figure 2.5). Age control in the fan deposits is limited to rare exposures of the Lava Creek B, Bishop and Tuff C ashes.

Several lines of evidence suggest that tectonic processes influenced sedimentation and deformation of the Tecopa Lake beds. Northward progradation of the Tecopa fan systems occurred following Tuff C deposition during times of relatively dry climate conditions, inferred from low lake levels, suggesting tectonic uplift of the Tecopa Hump drove coarse-grained sedimentation northward. Basinward tilting and resulting higher dip of Tuff C (Fig. 2.4) and underlying strata relative to that of the Bishop and Lava Creek B ashes suggests that uplift did not continue throughout the depositional history. However, post-Lava Creek B basinward tilting and increased dip are observed in the uppermost part of section LT-9C, suggesting that at least two periods of uplift were contemporaneous with the preserved depositional record of the Tecopa Lake beds. Additionally, localized accommodation produced by the uplift resulted in steep depositional gradients unique to Lake Tecopa's southern margin in the basin. Despite enhanced accommodation space, lack of evidence for migration of saline, alkaline brine into these areas, as evident by limited zeolite and potassium feldspar alteration, is consistent with broad uplift as opposed to fault-bound accommodation.

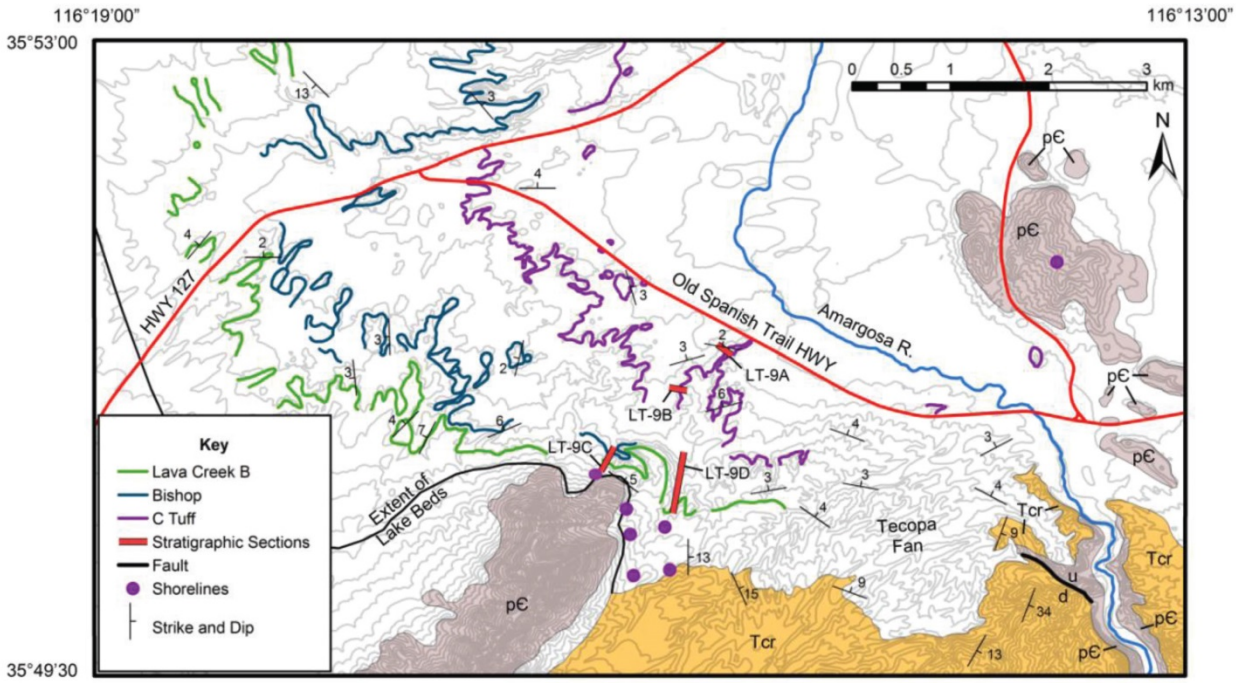


Figure 2.4. Map of south central part of the Tecopa basin showing the generalized extent of Pleistocene Tecopa Lake bed exposure with distribution of the Lava Creek B, Bishop, and Tuff C ashes, Miocene China Ranch beds, and Precambrian and Cambrian rocks (pC) (modified from Olson and Larsen, in review). Other non-colored areas are Holocene deposits.

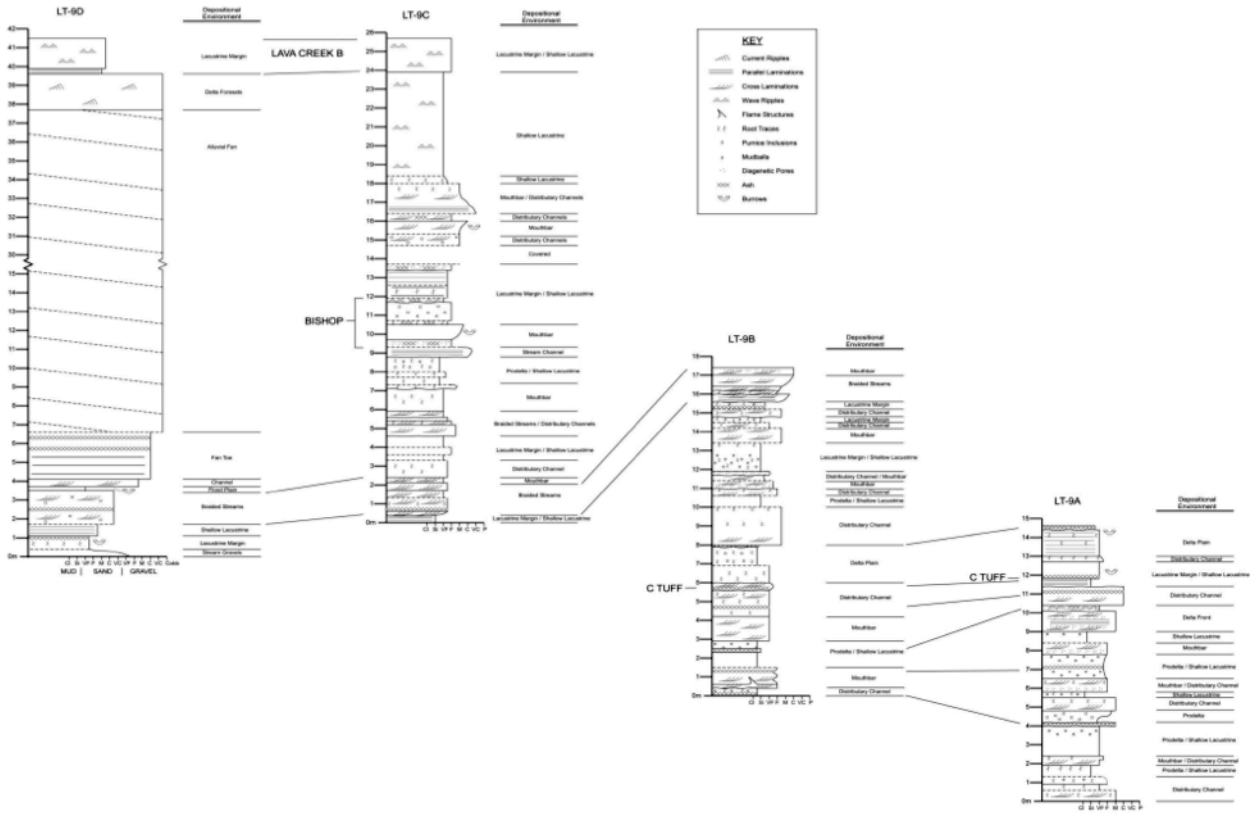


Figure 2.5. Stratigraphic sections 9A, 9B, 9C, and 9D. See Figure 2.4 for section locations.

<u>Day 3</u>: <u>Chjina Ranch to Tecopa</u>

No Road Log today. GONE FISHING! once you pullup stakes and see the all the vehicles rolling then plug these coordinates intoi your GPS and let your navigator get you here. Don't hurry we'll wait for you.....yea right.

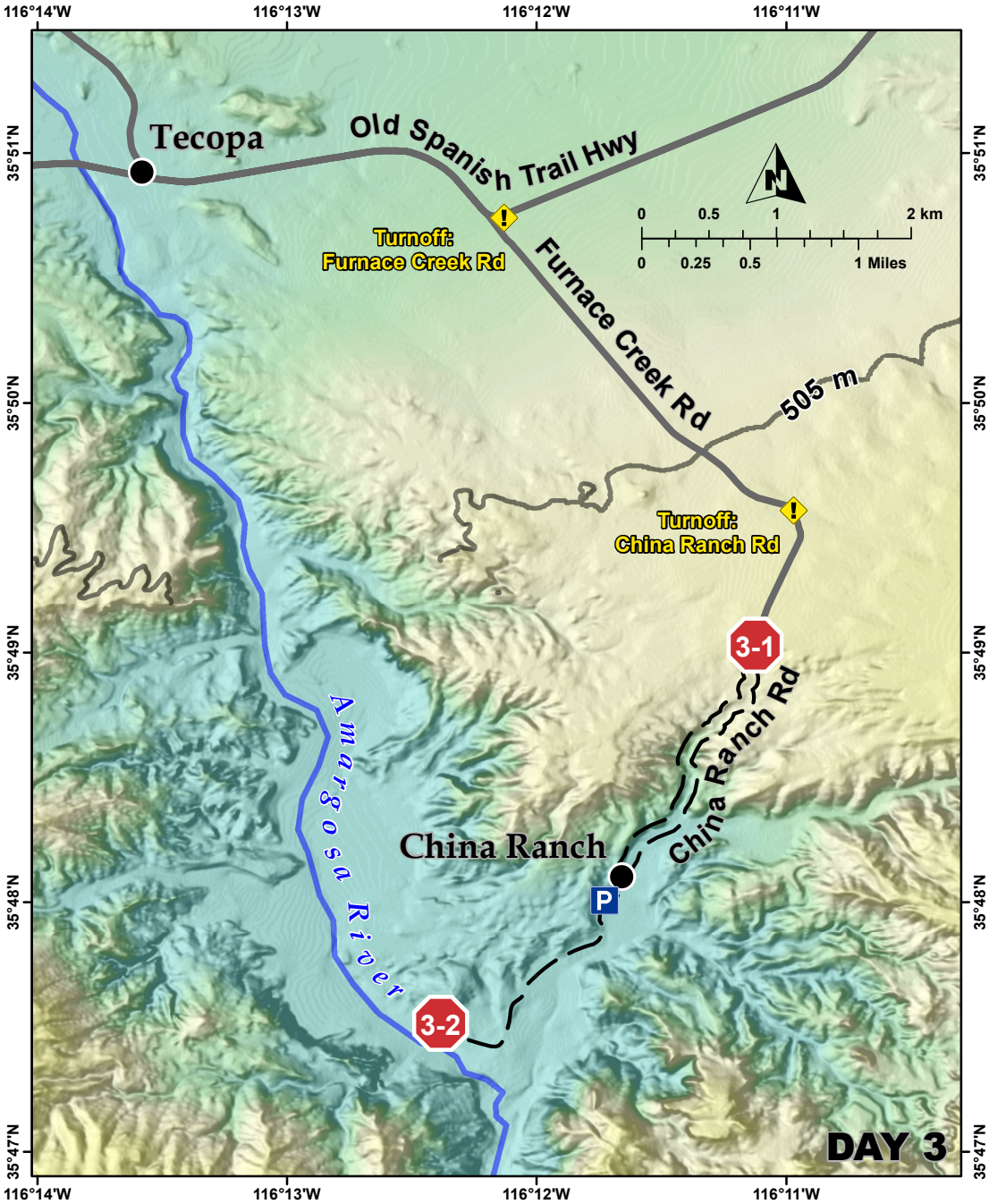


Figure 3.1.A. Field trip map showing Day 3 Stops 3.1 and 3.2. For the third Day 3 stop (Stop 3.3 at Eagle Mtn) please refer to Figure 3.1.B.

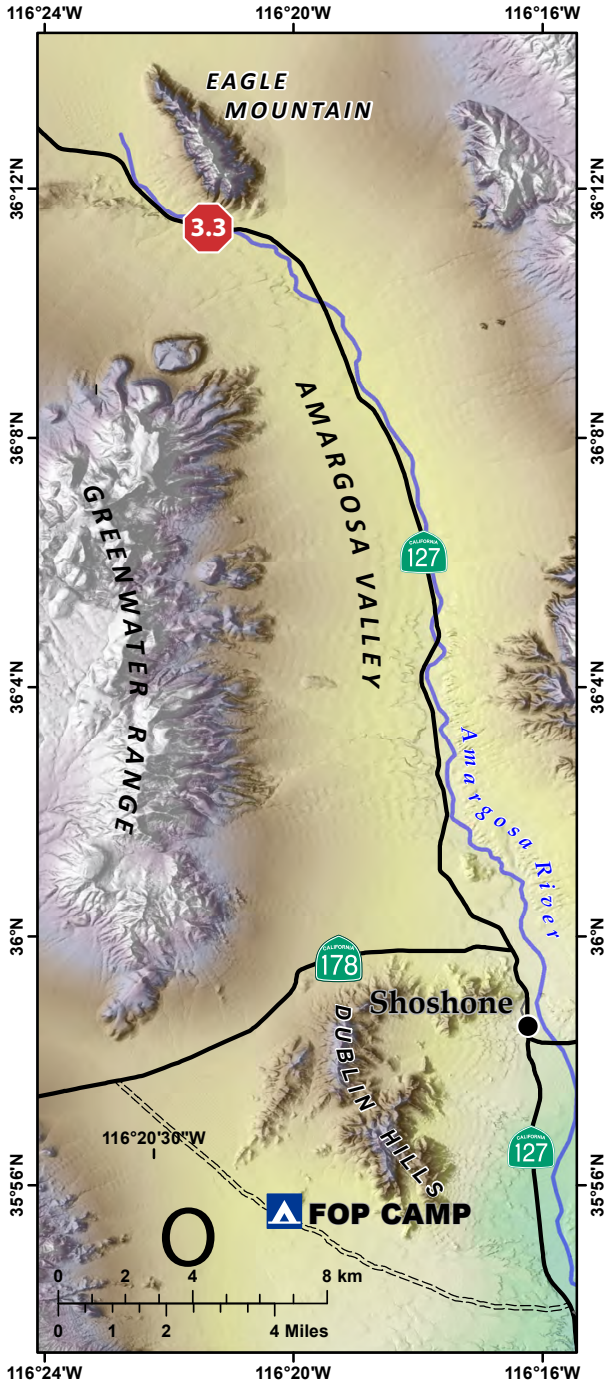


Figure 3.1.B. Field trip map showing Day 3 OPTIONAL STOP 3.3. For the third Day 3 stop it just a short drive north out of Shoshone.

Turn off to Furnace Creek Wash Rd 35.845689°-116.202189°
Turnoff China Ranch Rd 35.826333°-116.182894°
Day 3 Stop 1 35.817151°-116.185411°

Getting to FOP! — Directions to camp, south side of Dublin Hills

From the South: From I-15 at the freeway "town" of Baker, CA:

• from the Mad Greek restaurant, head north on CA-127 for 51 mi; look for left (west) turn onto Furnace Creek Wash Road. This is a dirt road. There is a corner street sign where the road intersects the CA-127) The left turn is roughly 100 m or so beyond the well-marked, hard-right turn on paved fork toward Tecopa Hot Springs.

• turn left (west) onto Furnace Creek Wash Rd; go 4.8 miles. Just as you are about to reach a narrow ridge of volcanic rocks that lowers to the road from the right, make a hard right turn onto unmarked dirt road

• Welcome Home! From here just drive down the desert-paved fan a ways and look for a place to set your camp! We will be at this same camp for all 3 days of the trip.

From the North: From US-95 at Beatty, NV:

• travel S on US-95 for 30 mi to the turnoff S onto NV-373

• take NV-373 S for 16.3 mi to the NV-CA state line. Here the road name changes to CA-127. Continue S on CA-127 for 33 mi to the intersection with CA-178 on the right (W); from here you have two options for getting to camp

• Option 2 (below) is only about one mile longer, but it is considerably less dusty than Option 1. Option 2 also takes you through the small town of Shoshone (just a half mile farther down up the road) in case you need any basic supplies like beer or more beer.

Option 1 (the dustier, slightly shorter option):

• turn right (W) onto CA-178; go 6.8 mi to Furnace Ck Wash Rd.

• turn left (S) onto Furnace Creek Wash Rd. and continue for 3.9 mi

• As you pass the low ridge of volcanic rocks on your left, veer left to turn off the main

dirt road and down into the camping area that will be our home for the duration of the FOP.

Option 2 (the less dusty, slightly longer option):

• from the intersection of CA-127 and CA-178 W continue S on CA-127 ~1 mi to Shoshone.

• from Shoshone, continue S on CA-127 for another 5 mi (~6.7 mi total miles S of CA-178 W intersection) to Furnace Ck Wash Rd, which is a dirt road that comes up on the right (W) just before you reach the well-marked left fork toward Tecopa Hot Springs

• turn right (west) onto Furnace Creek Wash Rd; proceed 4.8 mi. Just as you are about to reach the narrow ridge of volcanic rocks from the right make a hard right turn

• Welcome Home! From here just drive down the desert-paved fan a ways and look for a place to set your camp! We will be at this same camp for all 3 days of the trip.

The lat-long coordinates for the turnoff to camp from Furnace Creek Wash Rd. are 35.9013 N, 116.259588° W.

The well-paved and varnished fan surfaces in this area should make for an unlimited number of nice, semi- flat, campsites.

Please do everything possible to minimize disturbance to the fan surfaces around our camp, and upon departing camp on Sunday please restore your campsites as much as possible to their natural undisturbed state.

Appendices — Research Summaries and Reprints to Accompany Field Trip Stops Described in Road Log, Days 1 through 3.

Day 1 Appendices (A1.1 - A1.6)

Appendix A1.1

General Introduction to the Tecopa Basin Field Trip (Reheis, Caskey, and Bright)

The Tecopa basin-fill sediments have been studied for decades to investigate basin stratigraphy, sedimentology, diagenesis, chronology, and paleoclimate record. Sheppard and Gude (1968) first examined the chemical alteration of the sediments and recognized three main tephra layers—named Tuffs A, B, and C—which formed the basis for correlation of stratigraphy throughout the basin. Jack Hillhouse (1987), with the help of Andrei Sarna-Wojcicki et al. (1987; also Izett et al., 1970) correlated these tephra to the Lava Creek, Bishop, and Huckleberry Ridge tuffs, respectively, using glass chemistry and paleomagnetic data. Roger Morrison (1991, 1999) made the first attempt to construct a comprehensive lake-level history for the Tecopa basin, and proposed that the last major highstand reached ~543 m early in OIS 6. Chris Menges (2008) used geomorphic analysis to interpret an integration history for the Amargosa River through the Tecopa basin, and concluded that the river was blocked from entering the basin for a long period during at least the middle Pleistocene. Finally, Dan Larsen (2008) recently refined the lacustrine history using detailed sedimentology and patterns in the style and degree of diagenetic alteration, and continues to work in the basin today with his students.

For the last few years, John Caskey, Marith Reheis, and Jordon Bright have been studying the lake history in the basin by looking for preserved shoreline deposits and (or) strandlines around the basin as first order indicators of lake levels and by analyzing sediments for ostracodes as indicators of depositional and hydrochemical environment. Additional work is being done by Gary Scott, Al Deino, and others on dating the tuffs, paleomagnetic chronostratigraphy, and soft-sediment (e.g., seismogenically induced) deformation expressed within tephra layers throughout the basin (Gibert et al., 2011). Much of this recent work, still in progress, will add considerable detail to our understanding of Tecopa basin and Amargosa River history. We will show some of this recent work on this field trip as well as new work by Larsen and his students.

Beach sands and gravels that are in stratigraphic context with dated tephra layers document lake highstands, allow better reconstruction of lake level, and permit evaluation of basin deformation (Table 1). During the next three days, we will show you several examples of beach gravel and sand associated with lakes of Lava Creek age (640 ka (Day 1, Stops 1.1, 1.2, and 1.3) and describe others. We will also show evidence of younger beach gravels, typically 12-15 m stratigraphically above those of Lava Creek age (Day 1, Stops 1.2, 1.3, and 1.4) that document a younger lake that may have an age of ~600-500 ka based on estimated deposition rates and preliminary U-series analysis by Jim Paces (USGS). Gravels and strandlines representing both of these lakes are present from the north end of the basin to the south end and indicate as much as 20 m of tilting or displacement up to the north (Table A1.1). Whether this is due to basin-wide tilting or to one or more faults is as yet unclear. In addition, we have identified a distinctly younger possible beach gravel (Day 2, Stop 2.1,) that is inset within older deposits and may

represent the last lake highstand prior to incision of Amargosa Canyon. This has implications for the timing of downcutting and the arrival of the Amargosa River in Death Valley and the history of Lake Manly (Forester et al., 2005; Caskey et al., 2013). We also recognize beach deposits associated with the Huckleberry Ridge tephra in the southeastern part of the basin, the only area where it occurs (the tephra mapped as equivalent by Hillhouse (1987) along the Amargosa River is much younger—new data to be presented on this trip by Gary Scott). We have not found unequivocal nearshore deposits associated with any other stratigraphic intervals thus far. In sum, we can document with certainty only 5 lakes that were large enough (had enough fetch) to construct recognizable gravel shorelines.

Our other major approach has been to analyze and identify ostracode species and assemblages in sediment samples collected from measured sections in a range of altitudes. The distribution of ostracode species is primarily controlled by the relationship between the ratio of carbonate alkalinity (ALK) to calcium (Ca) and the total dissolved solids content (TDS) (Forester, 1983, 1986; Forester et al., 2005), and also by water temperature and habitat permanence (Curry, 1999; Quade et al., 2003). The Amargosa River has a high level of alkalinity relative to calcium (high ALK/Ca), in contrast to most of the springs and groundwater present in the Tecopa basin (Larsen et al., 2001). Thus, in combination with sedimentology, ostracode assemblages should be able to discriminate between lakes dominated by Amargosa River water and lakes, wetlands, or groundwater-discharge deposits sustained mainly by local groundwater.

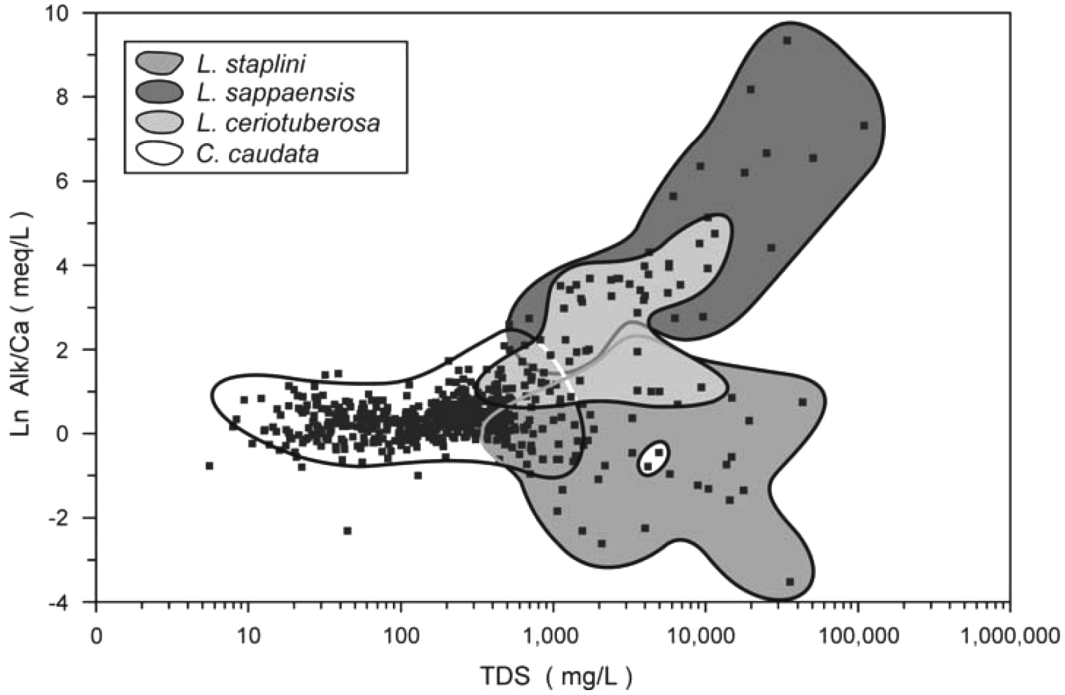


Figure A1.1.1 Total dissolved solids (TDS) versus the alkalinity/calcium ratio (ALK/Ca) for ostracode species. Ostracode samples (black squares) in the plot are common to continental environments such as springs, wetlands, and lakes mostly throughout the U.S., and to a lesser extent, Mexico plotted as. Modified after Forester et al. (2005). This paper can be found in full in Appendix A3.

Identification of ostracodes in the Tecopa basin is hampered by the effects of sediment diagenesis and by outcrop weathering; many samples yielded no ostracodes, or only non-diagnostic valve fragments. However, many samples from Lava Creek- and post-Lava Creek-aged lake deposits contained common to abundant *Limnocythere sappaensis*, an ostracode that requires a high ALK/Ca ratio and can tolerate high pH levels, indicating Amargosa waters. In contrast, lake deposits associated with the Huckleberry Ridge tephra contained *Limnocythere staplini*, which requires a low ALK/Ca ratio and thus was likely supported largely by spring discharge from a carbonate aquifer. Several other ostracode species that provide information on environmental conditions (Forester, 1983; Forester et al., 2005) are present in some samples from these and other typdes of deposits. For example, some samples from inferred deltaic settings near Shoshone contained ostracodes that live in spring-fed wetlands, and samples from sediments deposited between ca. 2 Ma and 0.7 Ma either were barren or contained ostracodes that live in shallow lakes, spring-fed wetlands, or groundwater. To date, these results suggest that the Amargosa River either was not present or contributed only minimal flow to the Tecopa basin during the early Pleistocene, and has implications for the integration history of the river as proposed by Menges (2008).

Appendix A1.2

Ostracodes (Jordon Bright) [the following abstract is slightly modified after Pigati et al., (2014)

Ostracodes (or ostracods) are a diverse group of generally millimeter-sized crustaceans that inhabit nearly all aquatic environments. Rather than possessing a singular carapace that protects the head (e.g., lobsters and shrimp) or dorsum (e.g., crabs), the ostracode carapace is bivalved and their anatomy has been reduced such that the carapace encapsulates the entire body. The ostracode carapace is composed primarily of calcite, but also contains a small amount of organic matter and trace elements. Upon death, their soft anatomy quickly decomposes, leaving behind disarticulated valves that are readily preserved in both lacustrine sediments and GWD deposits. Ostracode valves can be identified using morphological characteristics that are unique both at the genus level (e.g., adductor muscle scars) and species level (e.g., size, shape, surface topography).

The distribution of ostracode species in hydrologic settings, while complex, is not random. Many ostracode species are generalists and inhabit a wide range of aquatic environments. Others are more specialized (or limited) in their ecological requirements and display intimate associations with discrete habitats. The primary variables that control ostracode species distributions are the ratio of calcium (Ca) to carbonate alkalinity (ALK), the total dissolved solids (TDS) content and temperature of the host water, and habitat permanence (Forester, 1987; Quade et al., 2003). In most cases, ostracodes that inhabit water with high Ca/ALK find low Ca/ALK chemistries lethal, and vice versa (e.g., Forester, 1983). Thus, fossil ostracode assemblages can be used to infer the chemical compositions of past water bodies (Forester et al., 2005; Pigati et al., 2014).

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Appendix A1.3

Tecopa Hump and Integration of Amargosa River through Amargosa Canyon (McMackin and Menges)

At the south end of the basin the Sperry Hills mark a topographically high structural culmination that we refer to as the Tecopa Hump, an antiform that approximately continues 15 km in both directions from this site along a WNW-ESE trending axis with an asymmetric cross-profile much higher and steeper on the southwest (southern Death Valley) side.

In contrast to a classical anticline, the Tecopa Hump has a core of complexly faulted bedrock overlain by Late Miocene Basin assemblage, including the China Ranch beds (light colored sediments in middle and foreground) and extensive fanglomerates shed from adjacent ranges or uplifted bedrock in the Hump. Late Tertiary and older Quaternary fanglomerates dip as much as 10˚ to 15˚ north or south on the flanks of the Hump. We consider the Tecopa Hump to have not only played a major role in the original development of the boundary and paleodivide of the Tecopa basin, but also exposed and structurally inverted the core of an even older late Miocene sedimentary basin. The Hump also appears likely to have influenced the location and timing of the initial breach through the paleodivide that first integrated the Amargosa River between Tecopa basin and southern Death Valley. Both flanks of the Hump are extensively dissected by persistent and locally extensive incision (obvious in this area) which both predates and postdates initial integration of the Amargosa River in late-middle Pleistocene time. Persistent dissection into Miocene to Proterozoic basement rocks in this section of the Hump highland is especially well recorded by an inset or stair-stepped suite of at least three major and up to five or six detectable levels of strath or strath-fill terraces on the main-stem and tributary drainages (including Willow Creek) of the Amargosa River. We propose that the topographic form and geomorphic characteristics of the Hump, when combined with incision history (ie., spatial patterns, timing, and divergence-convergence height variations of strath terraces) suggests a continued low-level of uplift on this structural boundary of southern Death Valley throughout at least mid-late Quaternary time.

Appendix A1.4

Table A1.4.1. Summary of evidence for Tecopa basin lake levels and associated paleohydrologic environments

"Highest elevations" were obtained from differentially-corrected GPS data collected in the field using a Trimble GeoXH receiver (est. vertical precisions of ~±10 cm). Highest elevations of the oldest two middle-Pleistocene lakes (the *LC lake* and the *High lake*) is given for both the N and S ends of the basin, which reflects our preliminary interpretation that the pair of shoreline deposits northern part of the basin is uplifted 16-21 m relative to correlative shoreline deposits to the south. One presumption to this claim is that both of two different-aged shorelines in the northern part of the basin correlate with the pair of different aged terraces in the southern part. Beyond that we recognize that there can be significant uncertainties in how we use differences in shoreline elevations to estimate things like relative uplift because elevations of *highest observed* shoreline deposits can't be assumed to correspond to given lake's highstand elevation. *Highest observed* shoreline elevations are at best considered as minimum estimates lake high-stand elevations. We do not have sufficient observations of lakes older than the LC lake recognize any kind of deformation "signal" except that older basin deposits those of at least Huckleberry Ridge (HR) age (2.1 Ma) and perhaps older, are locally quite deformed (folded and faulted) in the SE part of the basin. Other notes: 1) The lowest elevation values provided in the table are mostly poorly constrained; 2) We have not yet identified unequivocal shoreline deposits that are of approximate Bishop ash-age. It is included in the table as a place marker because we haven't given up the ship.

Age	Highest elev. (masl) N end	Highest elev. (masl) S end	Lowest elev. (masl)	Range (masl)	Evidence, observations, and caveats; paleohydrologic environment where inferred from ostracodes
2.1 Ma (HR ash)	?	437	411	26	• highest identified HR beach sand is associated with beach sand and gravel. • range between highest and lowest was surveyed east of Tecopa Hot Springs Hills where basin deposits are highly deformed at least locally (faulted and folded), so range likely affected by tectonics; sediment compaction also possible (no data on this though) • ostracodes mostly *not compatible with Amargosa River (AR) waters.*
1.25 Ma (“Tecopa tuff”)	?	423	414	~10	• green mud sample just below tuff yielded ostracodes, so a possible shallow (i.e not a high-level)lake sampled from Ostracodes present cannot live in AR waters.
>0.76 Ma (pre-Bishop ash)	??	??	??	??	• AR gravels are observed below Bishop locally (north end of basin), so pre-Bishop lake of some size seems likely(?) • No ostracodes below Bishop ash in Camp Wash (Stop 2.2)—this is about the lowest elevations observed for Bishop ash in the basin (435 masl)—or in higher-level Bishop east of Shoshone • no unequivocal stratigraphic or micropaleo evidence for a high-level ~Bishop-aged lake (refer to Larsen, optional *Stop 1.5*; and Larsen (2008)
0.64 Ma* (***LC Lake***)	508	488	442-445	>45	• widely recognized throughout basin; easily identified; almost always with LC ash • highest-level LC lake shorelines are sometimes buried along basin margins • *LC-lake* high-stand elevation is well estimated in various parts of basin (e.g., multiple, well-preserved (buried) constructional beach ridges along N, and NE flank Tecopa Pk ; • lowest known exposure of LC layer is along CA 127 (Hillhouse, 1987) **Highest identified *LC lake* levels (north part of basin)** • *north of Shoshone*—highest likely deltaic gravels expressed as meter-high pebble-gravel foresets located 400 m, bearing 318˚ from Junction of Hwy127 and 178W (**508 masl**); • *NW side of Shoshone trailer park* (**Stop 1.3**)— strandline of cemented LC-ash around large

					bouldery talus blocks of *DV basalt* (**508 masl**) *amazingly preserved!* • *Dublin Gulch LC quarry, SW of Shoshone*—distinct symmetric oscillation ripples and ripple laminations in LC-ash (**500-505 masl**); • *E of Shoshone*—lower of two distinct fluvial deltaic sequences exposed north of CA Hwy 178E contains LC ash filled channels. **Highest identified *LC lake* levels (south part of basin)** • *N side of TP*—Seven exposures of high-level shoreline deposits **483-487 masl**, elevation between high-level shoreline sites deposit sites **487 masl**; uplift. • Dolomite hill (see Stop 1.1) **(488 masl)**; • 1.1 km due north of dolomite hill **(485 masl)**; **Paleohydrologic environments based on code assemblages** • Abundant (?) High ALK/Ca -tolerant ostracodes (*Limnocythere sappaensis, or Lsapp*) in samples collected near TP and the landfill east of Shoshone (south of Hwy 178, which suggests lake supplied by significant AR-fed lake (high ALK/Ca).
~600-500 ka (***High Lake***)	525	508	<443(?) (similar to LC-Lake(?))	>65(?)	**Highest identified High Lake levels (south part of basin)** • *NE side of TP*—several-m-thick section of beach sands and very well-rounded and polished pebbles (508 masl); excellent wash exposures in this area show *High Lake* deposits are ~17 m stratigraphically upsection from *LC lake* deposits. • ***There is a 21-m-difference in maximum shoreline elevations of LC and High lakes at TP.*** • *Erosional strandlines (narrow cut benches) on north bedrock flanks of TP and Hot Spring Hills*— both strandlines have well rounded and polished pebbles found up to **508 masl** • 500 m bearing 300˚ from the south side of Dolomite Hill—~2.5-m-thick section of beach sand and lesser amounts of v. fine pebble beach gravel; top of section at **507 m;** stratigraphic relations exposed in washes between Dolomite Hill and High lake site show that the High Lake deposits are up section of the LC Lake as well as **20 m higher** in elevation. • *1.2 km due north of Dolomite Hill (Stop 1.1) southernmost flank of arm of Dublin Hills*—thick section of well sorted and stratified pebble beach gravel **508 masl**. **Highest identified High Lake levels (north part end of basin)** • East flank of Dublin Hills 2.5 km due north of Dolomite Hill— up to 15 meters of mostly fine-pebble beach gravel; does include ~1-m-thick interval of interstratified alluvium; uppermost lacustrine deposits at (**521 masl)**; LC lake deposits seem to project to a level just beneath the exposed section of beach gravels. In this area the High Lake gravels are overlain by fan deposits with Stage 3+ carbonate development (east flank of Dublin Hills). • Dublin Gulch (see Stop 1.2 traverse in roadlog); alluvial fans appear to transition into beach deposits along a High Lake stand at ~**521 masl** • the stratigraphically higher of the two fluvial-deltaic sequences exposed at the north end of the basin is almost certainly associated with basin infilling during the *High Lake* phase (Stop 1.4); thin intervals of very well-rounded and polished pebble gravels show up high in the section exposed north of Hwy 178 (Stop 1.4), several excellent exposures of beach gravel in this same area range from523-525 masl. • the elevation difference between the highest levels of the LC Lake and High Lake is ~21 m at TP an only 13-17 m at the north end of the basin. The apparent jump (or drop) in elevations

					of both lakes appears to occur across a north-south distance of only about 1 km along the SE Dublin Hills. There are nice NE-striking Bedrock faults that project right through this area. no obvious scarps , but....? **Numerical age data** • U-series tufa age (from TP area) is 604 ±110 ka (2σ);; ostracodes (*Lsapp*) suggest AR-fed lake (high alk/Ca); Our middle Pleistocene *High Lake* deposits at north end of basin identified as OIS-6 (and last maximum) lake deposits by Morrison (1999).
184-130 ka(?) (***Terminal Lake***) possibly early OIS-6	???	505	443(?)	>60	Defined by erosional benches and washed, reworked, and size-sorted fan gravels throughout much of the basin at ~505 masl, and a thin sequence of beach gravel backsets on the Greenwater fan (Stop 2.1); interpreted as a brief "terminal" Lake Tecopa; referred as the Terminal Lake stand; though somewhat circumstantial, the apparently brief high stand may be due to breach event through the Sperry Hills divide in early OIS-6, consistent with ostracode evidence of AR-flow into Death Valley by 184 ka (Forester et al., 2005); upper age limit of AR integration constrained as prior to ~130 ka (U-series tufa age of +46 m Lake Manly shoreline that reworks AR-derived gravels in southern Death Valley); lowest elevation not known, but assumed to be similar to or lower than the upper elevation of a thick pristine section of fine-grained basin sediments that accumulated on the east flank of the Hot Spring Hills.
Terminal Stillstand(?)	NA	430-435	-	-	Reworked sands and gravels commonly observed at 430-435 masl; commonly marked by noteable well rounded and polished clasts; 430-435-m-elevation may correspond to the projected top of now incised bedrock (i.e., Stirling Quartzite) in the upper Amargosa Canyon wh; the bedrock "high" is well exposed in the canyon walls (both sides of canyon) from 1-2 km downstream of the canyon head at Tecopa (Figure 2.3.2a). The resistant bedrock "high" is interpreted to directly relate to the 430-435-m high "stillstand" lake. If breaching of the Sperry Hills divide was initiated at the time of the Terminal Lake Tecopa highstand (505 masl), then initially rapid incision of the canyon through the relatively nonresistant upper section of alluvial deposits (Figure 2.3.2b), would logically have stalled when river downcut to the level of the bedrock.
late Holocene	NA	413	399	14	Youthful strandlines expressed as <20-cm-high wave-cut scarps at and near Grimshaw Hill, and "bathtub rings" on the low flanks of hills along the southern margin the basin; cause for lake unknown, but presumed to be from a landslide dam in the narrower sections of upper AR canyon.

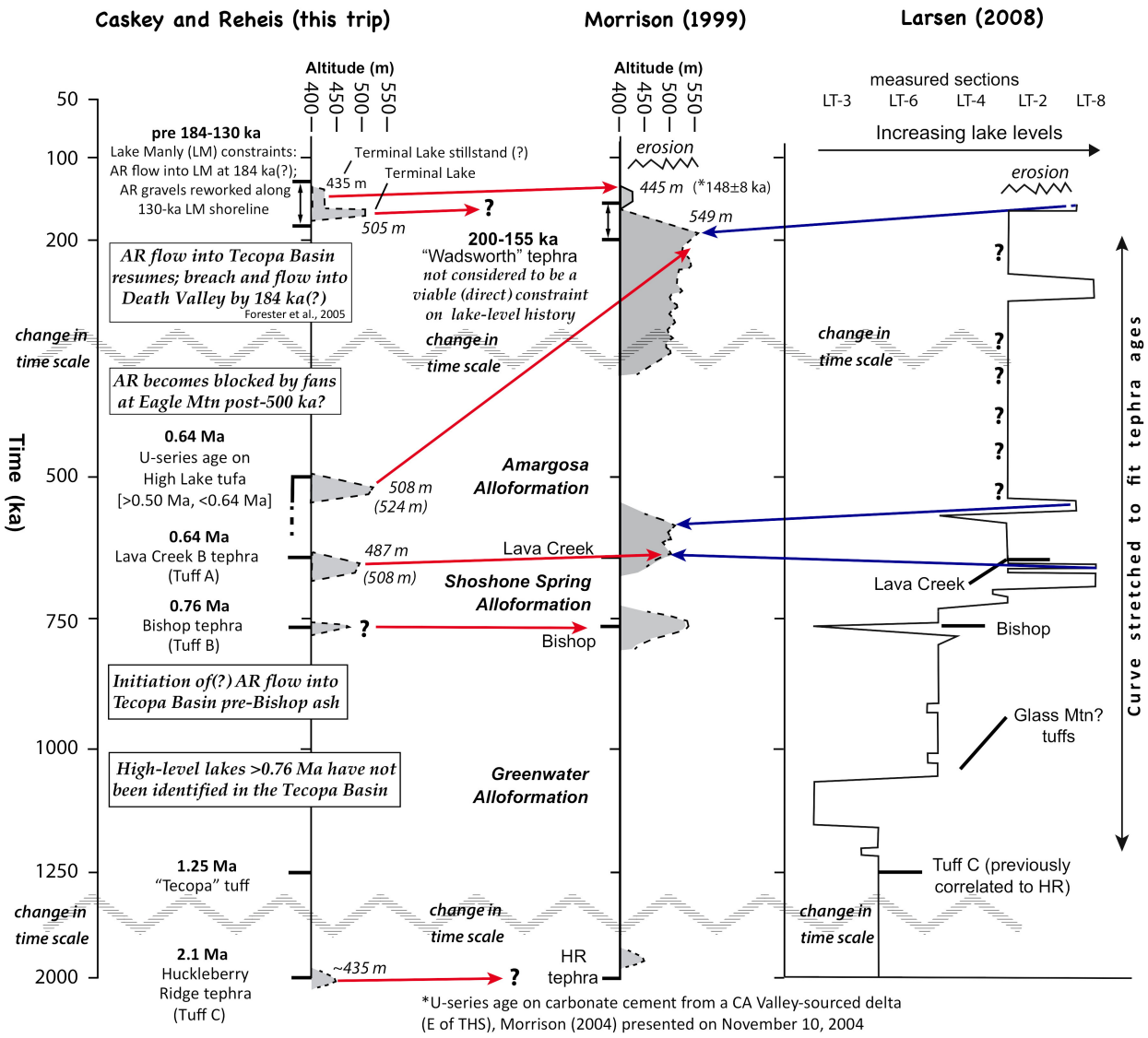


Figure A1.4.1. Comparison of lake-level interpretations by Caskey and Reheis (this trip) to those of Morrison (1999; red arrows), and Larsen (2008; blue arrows). Curves from previous publications were redrawn, and the Larsen (2008) curve was stretched to fit tephra ages (chronometric time scale not used in original curve). Interpretations of geomorphic events in the drainage basin shown in boxes on left; highstand elevations of Caskey and Reheis shown in italics, with southern basin elevations listed first and northern basin elevations in parentheses. Note two changes in time scale between 200 and 500 ka and 1250 and 2000 ka. Tephra layers shown by heavy black lines.

A1.5. Stratigraphy of the Tecopa Beds, north-central basin (section LT-10; see) (Larsen)

At the several stops in the north-central part of the Tecopa basin we will examine a largely fluvial and alluvial record of Pleistocene sedimentation interspersed with volcanic ash and thin intervals of lacustrine muds and deltaic sands and various shoreline features. These deposits record not only waxing and waning lake levels, but also subtle tectonic controls on sedimentary facies distributions. The main environments represented include alluvial fan deposits derived from the adjacent Dublin Hills and Resting Springs Range, ancestral Amargosa River channel, floodplain, and deltaic (subaerial and subaqueous) deposits, marshy to open water lacustrine muds, and calcareous tufa from spring deposition. Stratigraphic correlation of several sections (LT-1, -1B, -2, -8, -8B, -10) (Figs.A1.5.1 and A1.5.2a) measured in the Tecopa beds in the north-central part of the Tecopa basin are shown in Figure A1.5.2a-c). At stop 2 we will walk through parts of section LT-10. Key features at this section include a thick interval Lava Creek B ash deposited in a subaqueous deltaic setting with underlying Amargosa River gravels and thin lacustrine mudstone and overlying mid- to distal alluvial fan deposits. The close association of alluvial fan and Amargosa channel deposits at section LT-10 is interpreted to be a result of periodic westward tilting or faulting, causing Amargosa channels or lacustrine incursions to be focused along the western margin of the north-central part of the Tecopa basin.

A1.5. Stratigraphy of the Tecopa Beds, north-central basin—Larsen

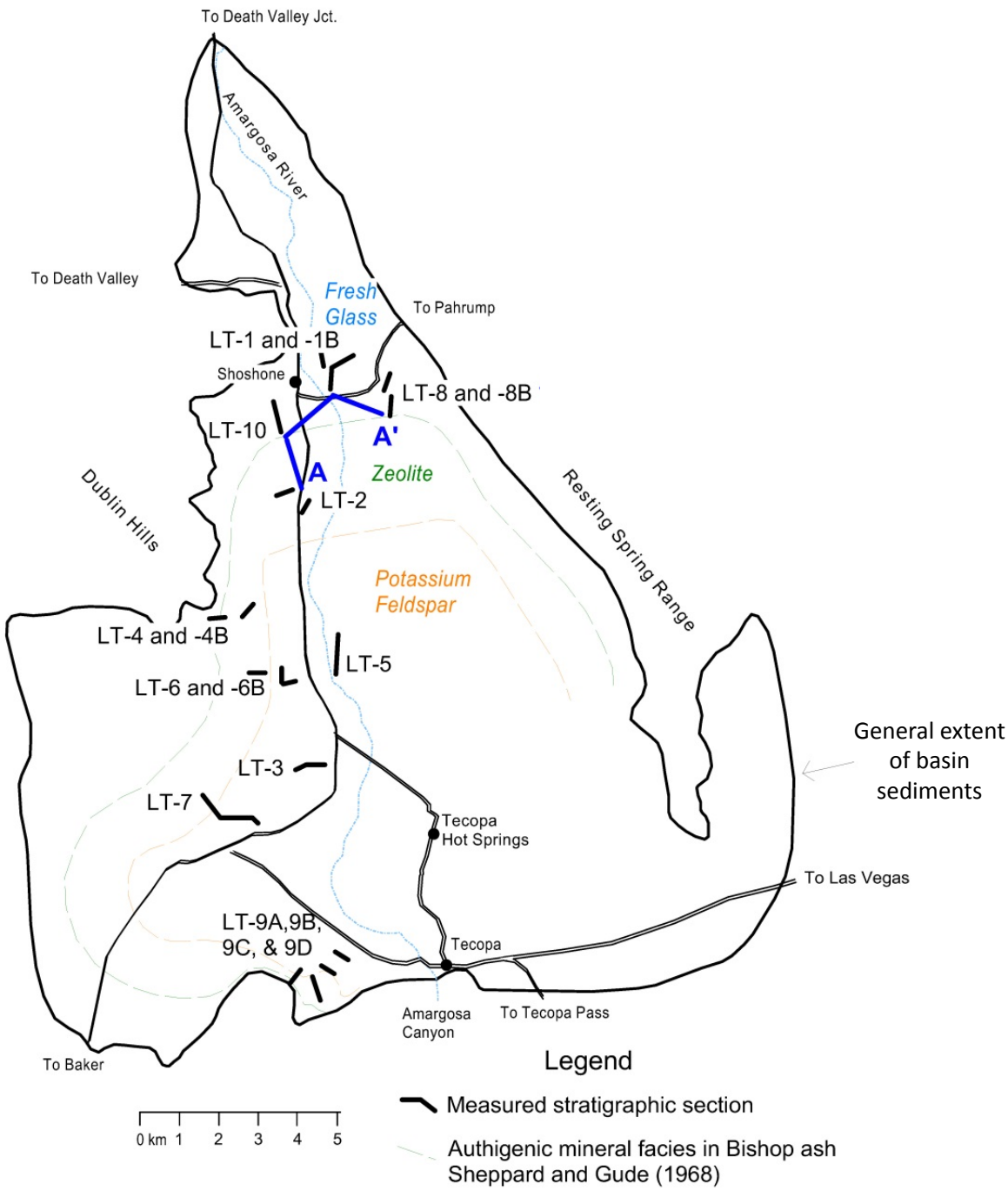
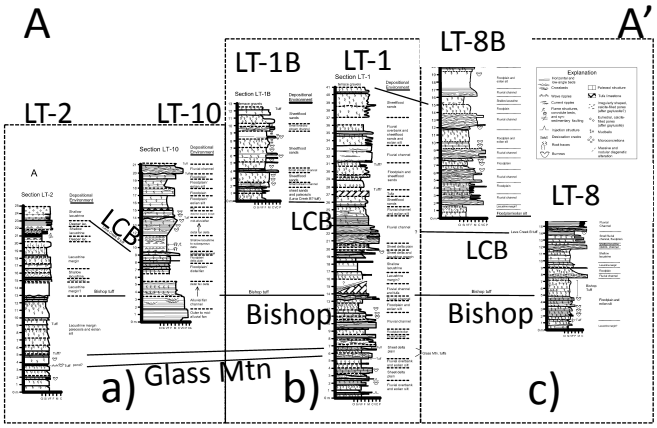
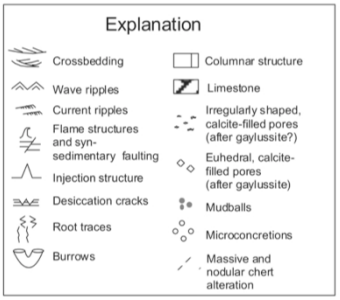


Figure A1.5.1. Map of the generalized outline of exposure of the Tecopa Lake beds (modified from Larsen, 2008) showing section locations discussed in text.

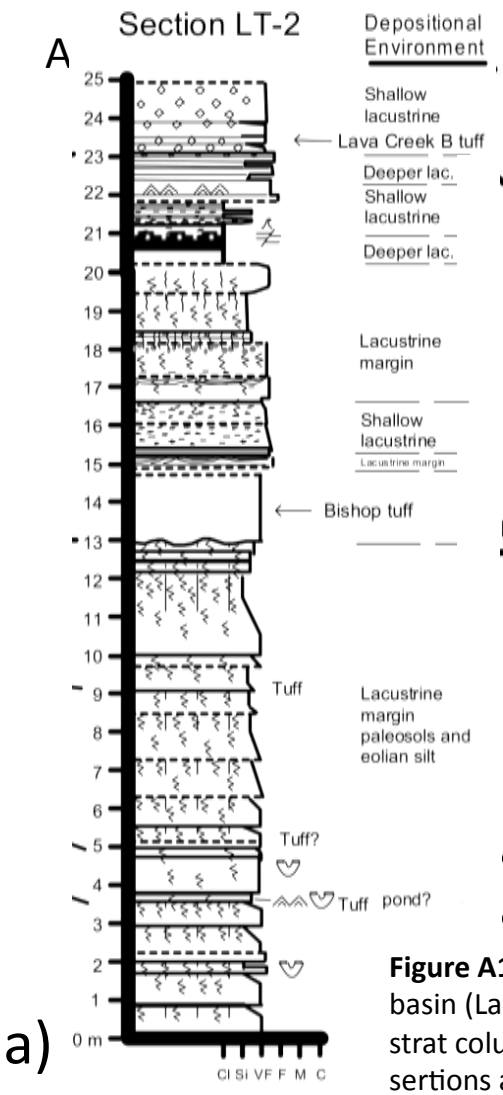
A1.5. Stratigraphy of the Tecopa Beds, north-central basin—Larsen





Index for strat columns shown on this page and next two pages

(East side of Hwy 127)



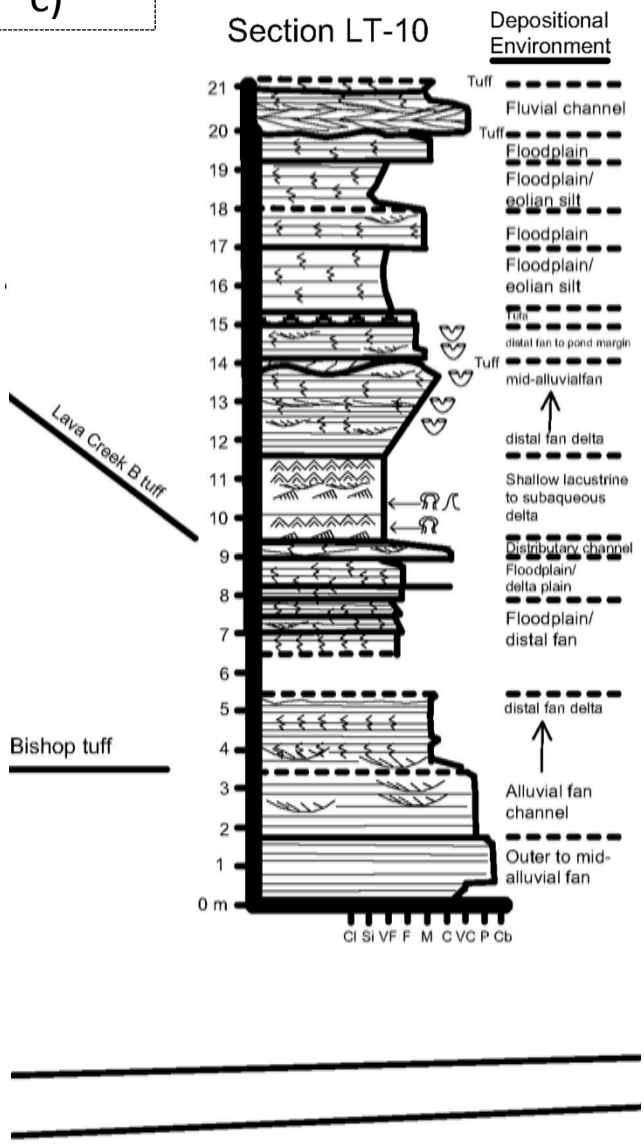


Figure A1.5.2a. Strat columns for the north-central Tecopa basin (Larsen, 2008). See Figure A1.5.1 for location of these strat columns and refer to inserts above correlations to othe sertions and lithologic explanation.

A1.5. Stratigraphy of the Tecopa Beds, north-central basin—Larsen

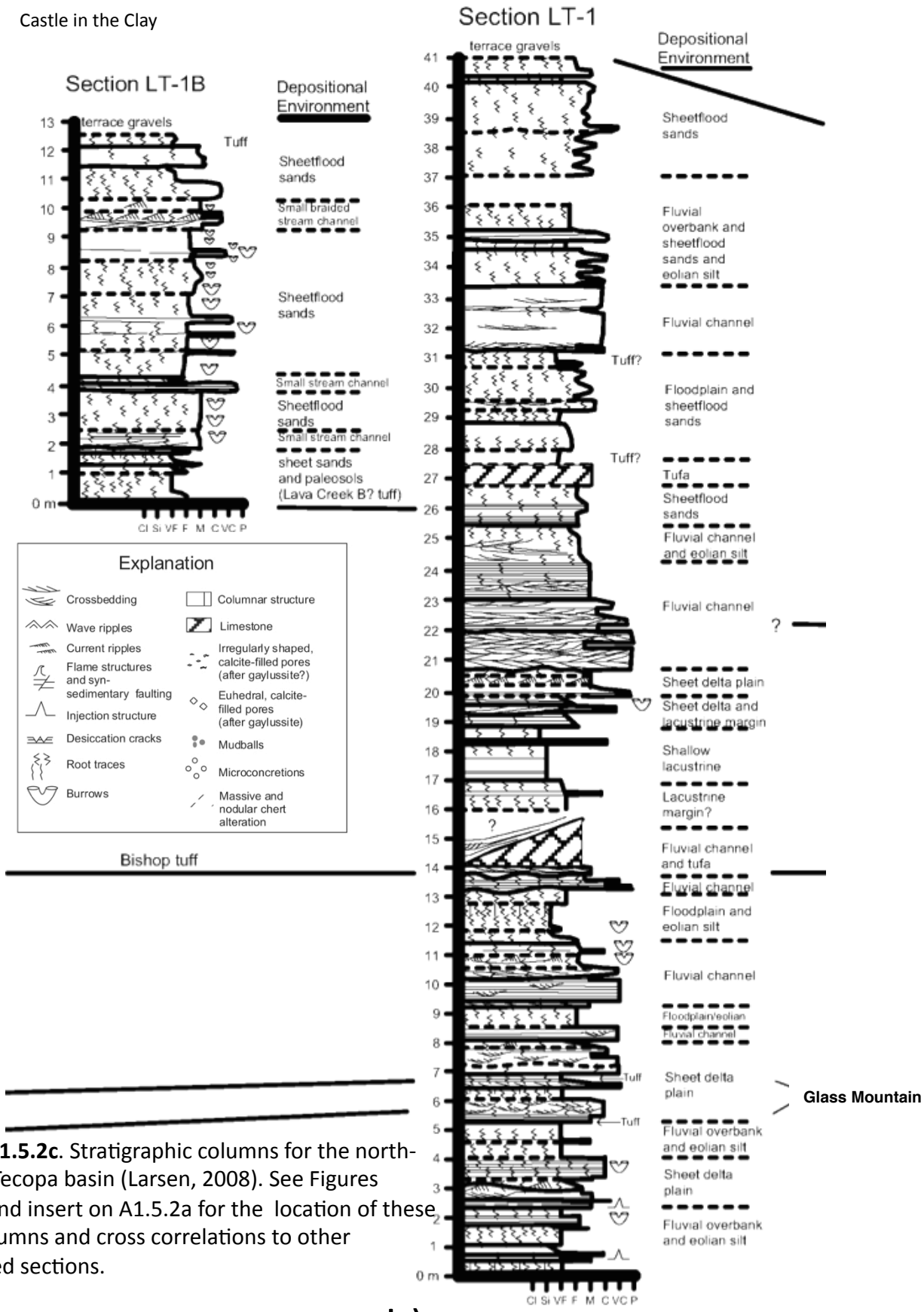


Figure A1.5.2c. Stratigraphic columns for the north-central Tecopa basin (Larsen, 2008). See Figures A1.5.1 and insert on A1.5.2a for the location of these strat columns and cross correlations to other measured sections.

A'

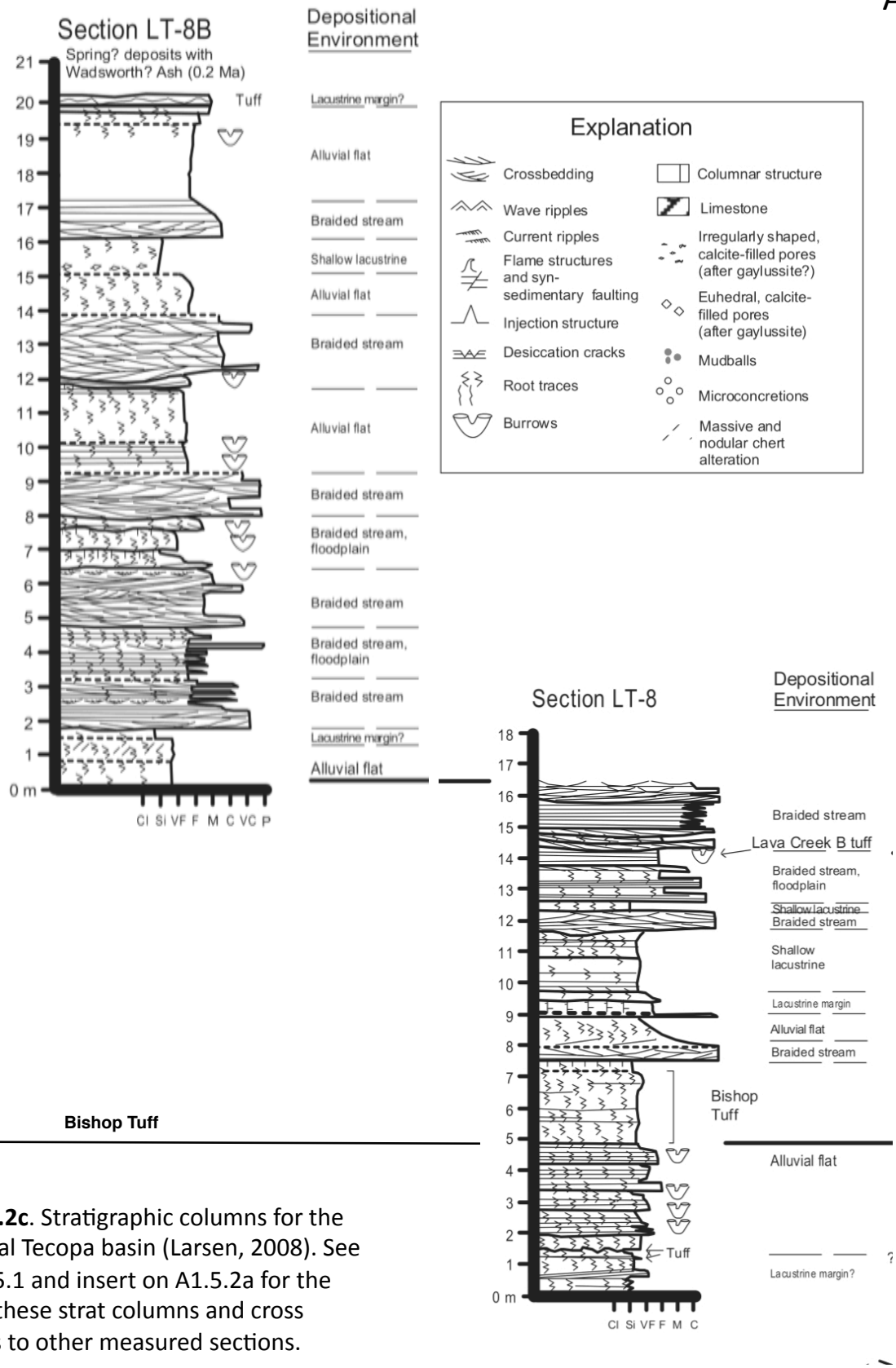
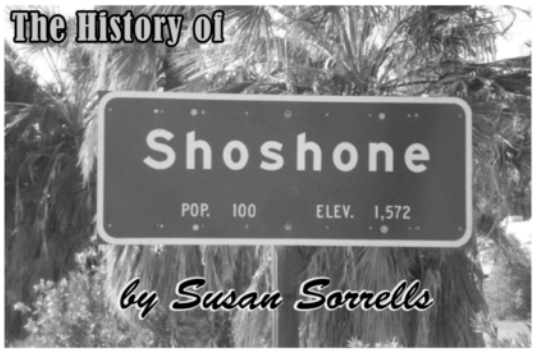


Figure A1.5.2c. Stratigraphic columns for the north-central Tecopa basin (Larsen, 2008). See Figures A1.5.1 and insert on A1.5.2a for the location of these strat columns and cross correlations to other measured sections.

c)



The history of Shoshone in many ways embodies the culture of mining in the first half of the 1900's. Even today there are historical structures in Shoshone that hint at the dynamics of a bustling mining community. However in order to understand the history of the village and the men and women who lived there it is important to recount the history of the area.

The first explorers entered the Amargosa regions in the 1820's when horse traders from New Mexico were blazing the Spanish Trail through Resting Springs and Tecopa. These traders probably visited Yaga, now known as Tecopa Hot Springs, which was the largest Southern Pauite village in Death Valley and the Amargosa. Shoshone however was off the beaten path. The impetus for the traffic on the Spanish Trail was horses not gold. All this changed in 1849 when gold was discovered in California and the gold rush began in earnest. In that same year two Mormon missionaries, James Brown and Addison Pratt found the first gold discovered in Death Valley itself at Salt Springs, a stop on the Spanish Trail between Resting Springs and what is now known as Baker. This discovery became an important trigger for the many prospectors roaming throughout the western United States to turn their attention to the Amargosa and Death Valley in pursuit of their dreams. It was during this era that the Death Valley prospector was born. The Gunsite silver and the Alvord gold in the Panamints and the Brefogle gold in the Amargosa also fired the imagination of these prospectors.

Twenty-five years later, in 1875 the mining activity in the area increased when William and Robert Brown found rich silver and lead deposit in the Resting Springs mining district just south of the stop on the Old Spanish trail called Resting Springs. They set up a town called Brownsville on Willow Creek creating a new rush to the area. Jonas Osborne further developed this property and although he struggled for thirty years to make a profit on these claims with little success his hard work paid off in 1906 when, with the arrival of the Tonopah and Tidewater Railroad, he sold his property for $175,000. In subsequent years his deposit produced over $4,000,000 making it one of the most prolific mines in the region. Although Brownsville, or Tecopa, as it later was called, was a thriving community, including a stage line from San Bernardino, Shoshone was known simply as Metbury Springs named after its sole inhabitant, who had come to the area to seek work as a miner at Brownsville.

In 1904, when Borax Smith incorporated the Tonopah & Tidewater Railroad towns started to spring up along the route and the population began to shift toward these station stops. In the late1800's borax was rising in prominence as a profitable mineral and in order to capture the market William Coleman built Harmony Borax works in Death Valley, then, in 1882, decided to build the Amargosa Borax works on the edge of the Tecopa wetlands to ship borax to Dagget by mule team in the summer when it was deadly hot in Death Valley. At the turn of the century as the borax market began to expand, Borax Smith, Coleman's successor, began to build a railroad to carry borax to markets from the rich borate regions in the area, and in 1906 the Tonopah and Tidewater Railroad was being constructed along the banks of the Amargosa. By June of 1907 the borax ore was trammed from the Lila C mine to the

railroad siding at Zabriski about 7 miles south of Shoshone. By August the Tonopah & Tidewater had reached Death Valley Junction and was shipping ore from the Lila C. In 1908 there were 85 registered voters in the Ballarat district, 68 in the Emigrant district, and 33 in the Tecopa district. There was no Shoshone district. In 1910 Dad Fairbanks, a freighter and storekeeper who had moved from Ash Meadows to Greenwater, a boomtown that had ended up being nothing more than a stock scam, decided to move to Shoshone to start a boarding house for the railroad. While he was in Shoshone Fairbanks became caught up in the colorful life of the prospector and placed claims on many properties in the area such a lead and silver prospect called the Redwing mine just outside Shoshone and a gold mine in Confidence Wash.
About the same time a Paiute elder named Tom Weed moved his family to Shoshone from Manse, Nevada. His wife Nellie was a well-known basket maker and traded her exquisite baskets for supplies. While in Shoshone Tom worked in many of the mining companies in the community such as the Pacific Coast Borax and Filtrol oil, He was known as a skilled worker and received a premium salary as documented in the "Day Books" kept by Dad Fairbanks in his store. Tom was also a dedicated community member and took part in community activities such as building the swimming pool in Shoshone. The residents of the Shoshone-Tecopa community enjoyed this swimming pool for many years. He and his family were very good friends with the Fairbanks family and this friendship between the two families has continued throughout the years.

From 1907 to 1922 the Shoshone Tecopa area was developing into a hotbed for prospectors and promoters. The town of Old Tecopa (formerly Brownsville) was moved from Willow Creek to its present location and by 1910 the Tecopa Railroad was in operation creating further expansion of the Noonday lead and silver district. A new mineral, talc, also appeared on the scene as a profitable commodity in the Southern Amargosa when Dave Morrison of China Ranch sold his claim to L. Lindsay who developed the Western Talc Mine just southeast of old Tecopa. Even Bob Montgomery of the Shoshone Montgomery and Skidoo fame sought to augment his fortune by staking silica claims several miles south of Shoshone with seven other partners. In 1916 Dad Fairbanks also experienced a change in fortunes. He discovered a clay deposit that in 1920 he sold to Associated Oil. The company later became known as Filtrol Oil and developed into a thriving business in Shoshone employing many miners and operating into the 1930's. A spur from the Tonopah and Tidewater Railroad to the clay deposit was built in the 1920's and was not removed until 1942.

From the 1920's through the 1930's Shoshone was fast becoming a thriving hub of commerce and social life for the surrounding area. About that time Dad Fairbanks asked Charlie Brown, his son-in-law, to join him as a partner in his business. In 1910 Charles Brown had married his sweetheart, Stella Fairbanks, Dad Fairbanks daughter, in Goldfield. In 1920, when he received the offer from Dad Fairbanks he was living in Tonopah with his family. Charlie Brown then moved to Shoshone and together these two men with their families turned a stop on the Tonopah and Tidewater Railroad in a thriving hub of commerce and social life for the surrounding area. Prospectors and others who were arriving daily could buy their supplies before setting out to stake their claims in Death Valley an area highly mineralized where many a prospector ventured to find his fortune but sometimes ended up losing his life. One hopeful prospector attracted another and before long many of the famous prospectors and mining promoters such as Shorty Harris,

George Montgomery, Siberian Red, Death Valley Scotty, the Ashford brothers, and Sir Harry Oakes called Shoshone their home. Since these men by nature were gamblers it didn't take long for them to establish the Mesquite Club where high stakes poker games took place, sometimes lasting for days. Dad Fairbanks was a first rate poker player and was known for keeping a game straight. These high stakes poker games were more important than banks and it is rumored that many of the successful prospectors in the area got their "venture capital" from these legendary poker games held at the Mesquite Club. The fame of the Mesquite Club spread throughout the region enhancing the reputation of Shoshone as the "in place" to stay before embarking into the surrounding desert.

By 1927 Dad Fairbanks at the age of 70 decided to move to Baker, California to take advantage of the business opportunity created by a new highway that was passing from Barstow through Baker to Las Vegas and beyond. By this time Charlie Brown had accumulated a sizable nest egg and was able to purchase with cash Dad Fairbanks' interest in the business in Shoshone. Being a man with foresight Charlie saw the importance of the automobile, which had replaced the horse and buggy, and was able to negotiate a deal with Standard Oil to become one of the first Standard Oil dealers in the state of California.

But Charles Brown also had other interests in addition to his business in Shoshone. Having lived on his own since his early teens he consistently was committed not only to his own business and family but also to his community. In 1923 he was asked by the members of his county to run for county supervisor and was elected in 1924. He was the first supervisor from the Death Valley Amargosa region and in order to attend the monthly meetings in Independence he traveled over 1000 miles: he would take the Tonopah and Tidewater Railroad to Crucero then catch another train to Los Angeles, then another to Mojave and finally arrived in Independence. He remained a supervisor until 1938 when he was elected to the state senate, an office he held until 1962. Shoshone was continuing to grow as a center of activity and by 1932 there were 65 voters in the Great Register, more than Tecopa and almost as many as in Death Valley. As the population of Tecopa and Shoshone, grew so did the social life.

From the 1920's through the 1930's Dublin Gulch, on the outskirts of Shoshone, also was developing as a destination and was becoming well know in the area. In 1922 the first precinct in Shoshone was established and included 19 voters and Harold Ashford, John Sheridan John Crowley and Joe Volmer all appeared on the registry. All of these men listed mining as their profession and probably put their skills to work by building their cave dwellings in the Gulch. It was at this time that many historical references to Dublin Gulch and its residents begin to appear. Although Joe Volmer first appeared on the Great Register in Skidoo in 1908 it is speculated that Joe Vomer met Charlie Brown in Tonopah where they became good friends and when Brown moved to Shoshone Volmer followed him. As in the region as a whole, there was great diversity in the Gulch. There were three Republicans loyal to the Queen on the right side of the Gulch while on the left side lived two Socialists and a Democrat. Since Joe Volmer also brewed and sold alcoholic refreshments, the Dublin Gulch became a social center for miners, promoters and prospectors throughout the region. Volmer also was on the school board, so many community and county officials also visited him at his home in the Gulch.

It is also speculated that shortly after 1920 when Volmer arrived in Shoshone he struck a deal with Fairbanks and Brown to build the first cave dwelling for himself then to

Dublin Gulch — Photo by Bruce Hawkins

build an adjacent one to rent out. John Cowley and his friends John Sheridan and Smoky Dickson soon moved in. By 1922 Sheridan, a well-known prospector also had struck it rich one and a half miles north of Shoshone. Sheridan was one of the oldest employees of the Pacific Coast Borax Company and as such recognized formations, which usually indicated Colmanite. He found such a formation while prospecting near his cave dwelling and when he dug down into the mud and at 8 feet he hit a large Colmanite lode. Pacific Coast Borax Co quickly bought the property for $50,00 and the mine became known as the Gerstley. The three friends, Sheridan, Crowley and Dickson, decided to celebrate their good fortune by buying a brand new pickup truck. Since they did not want to move from their homes they blasted out a garage next door and concocted a garage door to protect their prize. This garage next to their home can still be seen today.

By 1926 the Shoshone precinct had risen from 19 voters to 55, and Charlie Dobe Nels appeared on the rolls. His home however was more elaborate than the others since instead of a tin stovepipe, his home was constructed of masonry bricks. He also was more innovative in his method of prospecting: instead of a burro he use a Harley Davidson with a sidecar to transport him and his gear to remote locations in search of the next big strike.

Across the Gulch from these successful but eccentric prospectors lived three brothers from England who, at the turn of the century, had become very successful miners and mining promoters in the rich goldfields of Randsburg. Harold Ashford, the youngest brother came to Death Valley in 1906 to develop a gold mine in the Black mountains on the east side of Death Valley called the Golden Treasure. Harold Ashford, however did not show up in the Tecopa and

Shoshone area until 1914 when he appeared on the Great Register of Inyo County as living in Zabriskie. In 1920 Henry Ashford was included in the census in Pahrump, Nevada. Although Henry and Louis were both in the area during these years they probably were not on the Great Register of voters because there is no evidence that they had become citizens. From 1906 until 1932 the Ashford brothers engaged in a cycle of selling their properties then ending up in expensive litigation related to the sale. As Carl Glasscock, a desert historian astutely observed, all the Ashfords got out of the Golden Treasure was and I quote "just enough gold to keep them in groceries and lawsuits for more than a third of a century."

In 1932, however, their luck changed and the LA times reported from Randsburg that a mining engineer, Ralph Reed, had arranged a sale of the Ashford property and mill at Death Valley for a down payment of $25,000 and had agreed to put another $25,000 into equipment and to start development at once. Just four years later, according to the LA Times, on 2/16/36 Louis Rudolf Ashford passed away in Los Angeles. Henry and Harold continued to be a part of the Shoshone 9/29/14 social scene as reported in the Inyo Register by the Death Valley Rambler, in his column called " Death Valley Doin's. The real author of this column was Dr Schrum who had a practice in Death Valley Jct. Although Harold lived in Los Angeles he often visited his brother Henry in Shoshone. Henry continued to live in his cave dwelling in Dublin Gulch until he moved to Big Pine just before his death on March 27, 1947. He was 83. Henry was the last of the original residents of the Gulch and with his death an era ended.

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Superposed deformed beds produced by single earthquakes (Tecopa Basin, California): Insights into paleoseismology

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ABSTRACT

Well-exposed soft-sediment deformation (SSD) structures occur in Tecopa paleolake beds largely composed of ash from the Bishop (760 kyr) and the Lava Creek (640 kyr) ultra-plinian eruptions. In both cases the SSD structures affected volcaniclastic deposits concentrated by overland flow from the surrounding topography into the lower slope and lake environments. Re-sedimentation of ash produced alternations of volcaniclastic layers with different grain sizes and ash content that allowed for reverse density gradients, which favoured liquefaction and deformation. In the Tecopa Basin, extensive outcrops show these deformed strata continuing laterally for hundreds of metres, along with three-dimensional exposures. This study illustrates the 3-D complexity of soft-sediment deformation including some novel morphologic features. Of particular interest were examples of superposed deformed beds that laterally change into one single deformed horizon. Heterogeneities in ash concentration, grain-size, and water content produced contrasting permeabilities that were barriers to liquefaction and soft-sediment deformation. The probable causes for liquefaction were Middle Pleistocene paleo-earthquakes. There are several active faults in the region with enough seismic potential to produce moderate to large earthquakes. The fact that a single deformed deposit laterally merges into multiple superposed deformed beds indicates that multilayer liquefaction could be produced by a single shaking event. In these cases, SSD layers are insufficient and unreliable indicators of paleo-seismic recurrence intervals.

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1. Introduction

The particular geological history of the Tecopa basin, located in the Amargosa Desert in SE California (Fig. 1), has favoured the presence of well-exposed load-type soft-sediment deformation structures. The Tecopa lake beds show >70 m of outcropping deposits that record the geological history of the region between >2 Ma and ~0.2 Ma (Hillhouse, 1987; Morrison, 1999). This enclosed basin acted as a trap for Plio-Pleistocene ash eruptions in the Western United States. Many of these ash-rich deposits show soft-sediment deformation (SSD) structures, and can be used as key marker beds for correlation around the Tecopa Basin.

Soft-sediment deformation structures in lacustrine deposits have been widely described in the geological record (Sims, 1973; Hempton and Dewey, 1983; Alfaro et al., 1997; Rodríguez-Pascua et al., 2000; Moretti and Sabato, 2007; among others). Deformation structures generated during liquefaction by shaking from earthquakes are referred to as seismites (Seilacher, 1969). Lacustrine sedimentary environments have been noted as excellent recorders of earthquake events, owing to the presence of water-saturated sediment with high susceptibility to liquefaction (Sims, 1973, 1975).

Here we report a study on soft-sediment deformation structures that occur in the ash-rich Pleistocene deposits associated with the Bishop and the Lava Creek eruptions. These two tuffs outcrop extensively in the Tecopa Basin. The kilometric continuity and three-dimensional exposures of these deformed beds permit detailed observations of novel morphologic features developed during liquefaction. The most interesting observation is the presence of complex deformation where superposed deformed beds laterally evolve into one single deformed horizon. We discuss the seismic origin of these SSD features and if these superposed deformed beds are the result of single or successive events.

2. Geological setting

The Tecopa Basin is located in the Basin and Range physiographic province, 50 km SE of Death Valley, in what is now the Amargosa River valley. The basin is approximately 500 km^2, limited by the Dublin and Ibex Hills in the west, the Resting Springs and Nopah Ranges to the east and the Sperry Hills to the south. These surrounding mountains

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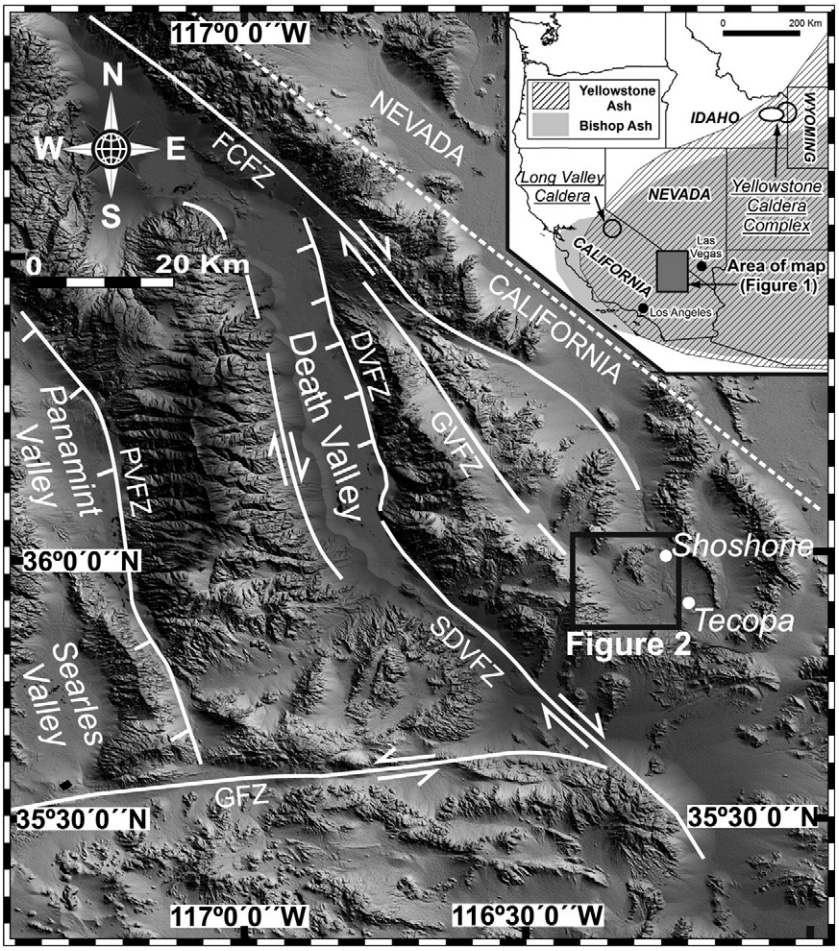

Fig. 1. U.S. location of Death Valley region showing the Long Valley and Yellowstone calderas. The digital elevation model shows the major geographic features and active faults in the Death Valley region, with the location of the study area. DVFZ: Death Valley Fault Zone, FCFZ: Furnace Creek Fault Zone, GFZ: Garlock Fault Zone, GVFZ: Grand View Fault Zone, PVFZ: Panamint Valley Fault Zone, SDVFZ: Southern Death Valley Fault Zone.

(basement in Fig. 2) are composed of Precambrian gneisses and metasediments, Early Paleozoic marine rocks and mid-Cenozoic volcanics. During the Late Pliocene and much of the Pleistocene, the Tecopa Basin was the site of an ephemeral shallow lake system. Toward the end of the Middle Pleistocene (~0.2 Ma) the Amargosa River breached the southern margin of the Basin and integrated with Death Valley (Morrison, 1999), thus initiating arroyo cutting that exposed >70 m of Pleistocene sedimentary deposits. Beds are sub-horizontal with a general dip of ~1° basinward. This geometry has been interpreted as the result of differential compaction on a slight depositional dip (Sheppard and Gudde, 1968; Hillhouse, 1987).

Numerous rhyolitic vitric ash layers are intercalated within the Tecopa beds. Two of them are chemically correlated with ultra-plinian eruptions from the western US: the Bishop (760 kyr) and Lava Creek (640 kyr) Tuffs (Izett et al., 1970; Izett, 1981). Detailed paleomagnetic studies have been carried out immediately below the Bishop Tuff, showing the Matuyama to Brunhes polarity change (Hillhouse and Cox, 1976; Hillhouse, 1987; Valet et al., 1988; Larson and Patterson, 1993). Both the Bishop and Lava Creek tuffs are key lithologic marker beds that can be identified at many outcrops along the basin.

Infilling of the Tecopa Basin was dominated by mud and silt in the central part and sand to clay around the margins, along with isolated carbonate beds and paleosols (Hillhouse, 1987). Paleoflow directions and fluvial facies distribution show an axial fluvial system coming from the north (ancient Amargosa river). Transverse channels drain into the basin through the Chicago (east) and Greenwater (west) Valleys. Around the margins of the basin occasional reports of ostracods, diatoms, vertebrate fauna, and footprints have been made, along with the more common paleosols, and wave and current ripples (Hillhouse, 1987). The central facies shows mudstones with calcite-filled casts of gaylussite (Larson, 2008), a salt found in alkaline and saline lakes from California, usually associated with halite and trona. Sheppard and Gudde (1968) described the occurrence of thin beds of finely crystalline dolomite, which could be related to periods of higher lake level in an alternating wet/dry environment. Morrison (1999) reported halite in some mudstones below the Bishop Tuff and Starkey and Blackmon (1979) reported sepiolite as the dominant authigenic clay-type mineral in the basin. Sepiolite is found mainly near ash beds, probably being precipitated when silica became abundant, through dissolution of volcanic ash in the Mg-bearing lake waters. Sepiolite occurs in modern desert lakes with high salinity (Hardie, 1968; Parry and Reeves, 1968).

Diagenetic processes affected the ash beds as well as most fine-grained sediments over large parts of the central Basin. Sheppard and Gudde (1968) describe three concentric zones with increasing alteration toward the center of the Basin: the fresh-glass, zeolite, and K-feldspar diagenetic facies. This diagenesis reflects chemically zoned pore waters, which were fresher near the margins and increasingly alkaline and saline basinward. According to Sheppard and Gudde (1968), this zonation was probably inherited from a depositional environment where the groundwater of paleo-Lake Tecopa was moderately-to-highly saline with a pH >9, except around the margins. All SSD structures developed before early diagenetic cement and lithification of the sediments. Deformation structures are ubiquitous in

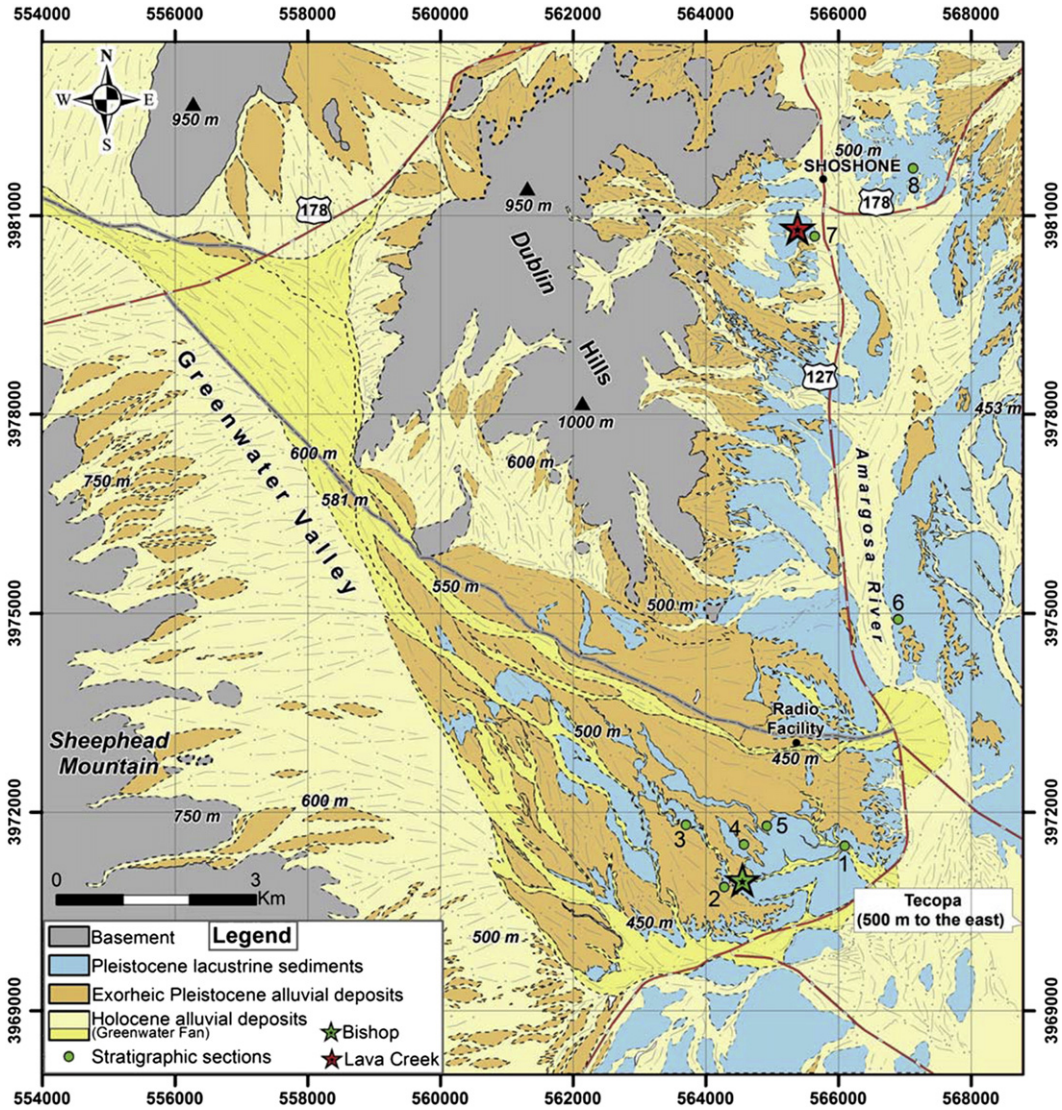

Fig. 2. General geological and geomorphological map of the Shoshone–Greenwater Fan Sector (western of Tecopa Basin) (based in Hillhouse, 1987). Inside is the location of the study seismites ouptcrops. Green and red stars indicate the location of Bishop and Lava Creek studied outcrops.

the central K-feldspar facies, common in ash-rich beds of the zeolite facies, and only locally seen in some ash beds of the fresh-glass facies.

We interpret the paleogeography of the Tecopa Basin as an ephemeral saline lake surrounded by local freshwater ponds near springs and fluvial inlets. Paleo-Lake Tecopa probably experienced large fluctuations in water level, chemical composition, and salinity reflecting the climate-driven hydrologic budget (see other examples in Rosen, 1994). During dry periods, paleosols would develop in marginal areas while sediments accumulated in a central playa by groundwater-fed discharge or occasional floods. During wet periods, a shallow lake would be temporarily established (Fig. 3). For example, the present interglacial climate does not allow the occurrence of a persistent lake in Tecopa owing to evaporation greatly exceeding precipitation (Hillhouse, 1987).

3. Stratigraphy of the deformed beds

Eight stratigraphic sections have been studied, five on the west side (Greenwater Valley), two on the north (near Shoshone) and one more central (Amargosa River) (Figs. 2, 4 and 5). Sections near Shoshone and in Greenwater Valley include the Bishop Tuff and some of them also the Lava Creek Tuff, which are distinctive beds with SSD structures. The central deposits represented by section 7 (Amargosa river) are correlated with section 1 in Greenwater Valley using a basal biotite-rich tuff (Tuff C of Hillhouse, 1987) (Fig. 4). The composite section from both the Amargosa and Greenwater areas indicates the existence of several other tuffs and deformed beds.

The sections in the Greenwater area are dominated by fine-grained deposits, largely mudstones and siltstones, unconformably covered by post-Lake Tecopa alluvial gravels. These sections are included within the zeolite diagenetic facies (Sheppard and Gudde, 1968), showing strong SiO_2 cementation, nodules and phillipsite neoformation. The correlation between sections uses the Bishop Tuff as a stratigraphic marker (Fig. 4). Among the different SSD structures that have been identified, the most spectacular examples occur in the volcaniclastic deposits of the Bishop and Lava Creek Tuffs, which were selected for this study. Average sediment accumulation rates of ~10 cm/kyr can be calculated for section 3 using the 11.6 m separating the Bishop Tuff (760 ky sensu Sarna-Wojcicki et al., 2000) and Lava Creek Tuff (640 kyr sensu Lanphere et al., 2002) (Fig. 4).

The two sections near Shoshone (8 and 9) are in the fresh-glass diagenetic facies (Figs. 2 and 5). In section 8, deposits below the pumicite

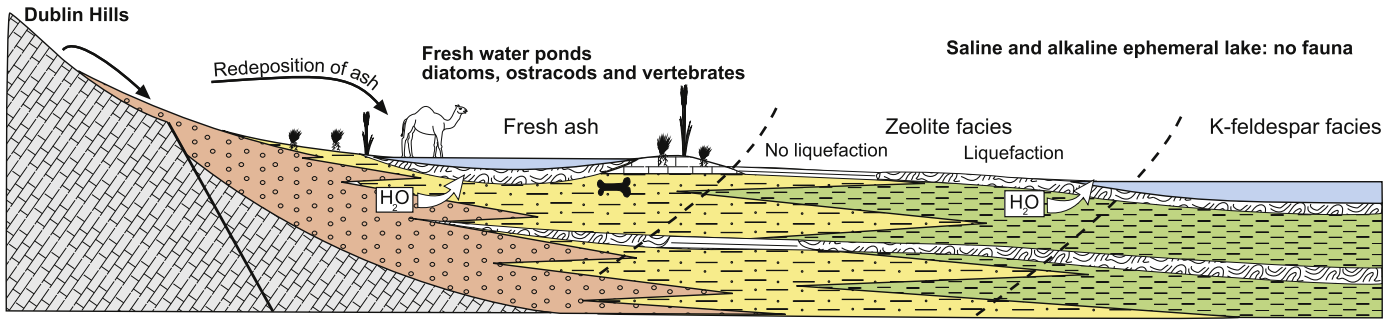

Fig. 3. Sedimentary model for Tecopa Basin, showing local ponds near the mountains and a central ephemeral lake. The figure shows the location of the two main ash beds affected by SSD and the distribution of diagenetic facies affecting these volcanic deposits described by Sheppard and Gudde (1968).

quarry consist mainly of alluvial sandstones covered by palustrine carbonates that include gastropods and paleosol development and volcaniclastic deposits related to the Bishop Tuff. The Lava Creek Tuff in this area is a 2.5 m thick deposit, apparently concentrated by re-deposition and overland flow from the surrounding uplands. Wave ripples are common, indicating shallow water in a pond environment. SSD processes variably affect these multiple beds of ash. In this quarry, Lava Creek Tuff is covered by coarser-grained alluvial deposits.

In section 9, the Bishop Tuff was affected by deformation and SiO_2 replacement during diagenesis. A fine-grained detrital succession, representing distal fluvial deposits, lay below this ash. Immediately on top of the Bishop Tuff are palustrine carbonates. An average sediment accumulation rate of ~7 cm/kyr has been calculated for deposits lying between the Bishop and Lava Creek Tuff. Here the Lava Creek Tuff occurs within fine-grained fluvial deposits (Fig. 4).

4. Study of soft-sediment deformation outcrops of the Tecopa Basin

Deformed beds are common across large parts of the Tecopa Basin. We chose to study the soft-sediment deformation structures in two prominent ash beds, the Bishop and Lava Creek tuffs. These two deformed key beds, characterized by inter-layered sandy and silty sediments, have a regional kilometric extension. Two high-quality outcrops were selected: along an arroyo in the Greenwater Valley; and in an abandoned quarry near Shoshone (Figs. 2, 4 and 5).

4.1. Soft-sediment deformation structures in Bishop Tuff (Greenwater Valley)

Well-exposed deformed beds <2 m thick are composed of two lithological units. The overlying unit consists of silt and fine sand, rich in volcanic shards; the underlying unit is made of fine-grained ash and mud. This deformed horizon is continuous and extends laterally for hundreds of metres (Fig. 6), allowing the observation of structures in both cross-section and plan view (Fig. 7).

The deformed beds can morphologically be described as pillows and sagging load structures (sensu Anketell et al., 1970; Alfaro et al., 1997). In vertical section, these load-type structures show curved and concentric laminae, which are slightly curved or planar at the base and highly dipping or vertical along the edges (Fig. 8). The thickness of these load-type structures varies between 0.5 and 1.5 m. Narrow fluidization channels, <1 m wide, separate individual load-type structures. In plan view they show an elliptical shape with the major axis ranging from 0.7 m to 3 m. The direction of the major axis has been measured in several outcrops separated by hundreds of metres and shows no preferred orientation (Fig. 7; see Alfaro et al., 2010).

A revealing morphological feature of this deformed bed is the presence of local lateral changes, from simple large load-type structures, into more complex structures involving two deformed beds with smaller load structures separated by less deformed (or undeformed) thin layers (Fig. 9). The transition to more complex deformation occurs within a few metres, and appears related to local facies changes. The simple reverse density gradient systems change to multilayered systems characterized by an alternation of sands with thin layers of fine-grained ash and mud. Some of these thin layers separate two deformed horizons or act as detachment layers generating unconformable surfaces (Figs. 8 and 9).

In the Greenwater the size of SSD structures in the Bishop Tuff decreases north-eastward (towards the margin) (Fig. 4). The deformation appears again farther north near Shoshone. These changes appear to correlate with areas less susceptible to liquefaction, probably without groundwater saturation or less appropriate density gradients, although an early diagenetic cementation could also be responsible.

4.2. Soft-sediment deformation structures in Lava Creek Tuff (Shoshone Quarry)

Excellent outcrops of deformed Lava Creek Tuff are located in the pumicite quarry near Shoshone. For 100 m along the quarry walls, several load-type and fluid-escape structures are exposed. Morphologically they are mainly sagging load structures, drop structures and pillows (sensu Owen, 2003) varying from centimetres to several tens of centimetres in size (Figs. 10, 11, 12 and 13).

The soft-sediment deformation that involves the Lava Creek Tuff is characterized by a main horizon ~1 m thick composed of sandy sediments with several intercalations of thin laminae of silt. Locally, this horizon shows complex internal deformation involving different deformed sandy layers separated by undeformed silty laminae (Fig. 13). Therefore, the main deformed horizon has a variable number of superposed beds involved in deformation. The most relevant feature is that these superposed deformed layers merge laterally into a single deformed layer as the smaller load structures pass into a single bigger load structure (Figs. 11 and 13). In the larger load structures, the silty laminae are cut by large fluid-escape structures, which extend from the bottom to the top of the deformed unit. The maximum thickness of the fluid-escape structures is ~1 m, through the total thickness of the Lava Creek Tuff (Figs. 11 and 13).

In one of the studied outcrops (Fig. 11) an apparently undeformed internal layer from the main horizon laterally evolves into an intensely deformed unit. Probably, local sedimentary conditions (degree of water saturation, depositional packing, etc.) were responsible for these local variations in style and magnitude of deformation.

5. Discussion of soft-sediment deformation

5.1. Trigger mechanisms

Various natural processes can trigger soft-sediment deformation in lakes, such as: overloading (Allen, 1982; Owen, 1987), wave-induced cyclical and/or impulsive stresses (Molina et al., 1998; Alfaro et al.,

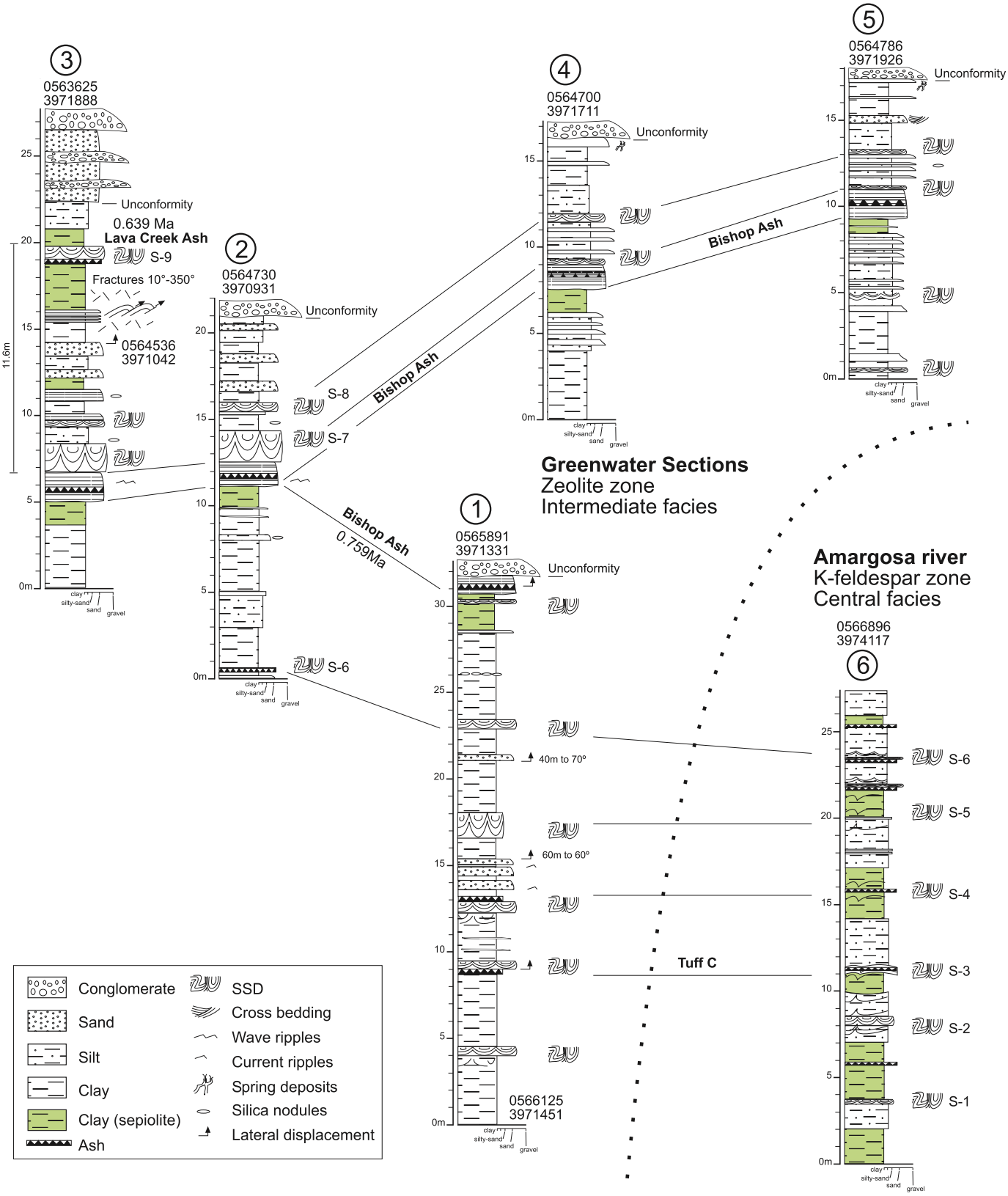

Fig. 4. Greenwater stratigraphic sections showing the location of Bishop and Lava Creek ash beds. The figure shows 9 superposed SSD beds which can be correlated locally. S7 and S9 occur immediately above the Bishop and Lava Creek tuffs in deposits originated after ash in the surrounding area was redeposited in the basin. This figure shows also a correlation with deposits outcropping in the Basin in the central facies near the Amargosa River. Both sections occur in different diagenetic zones (Sheppard and Gudde, 1968). Each section shows the UTM coordinates.

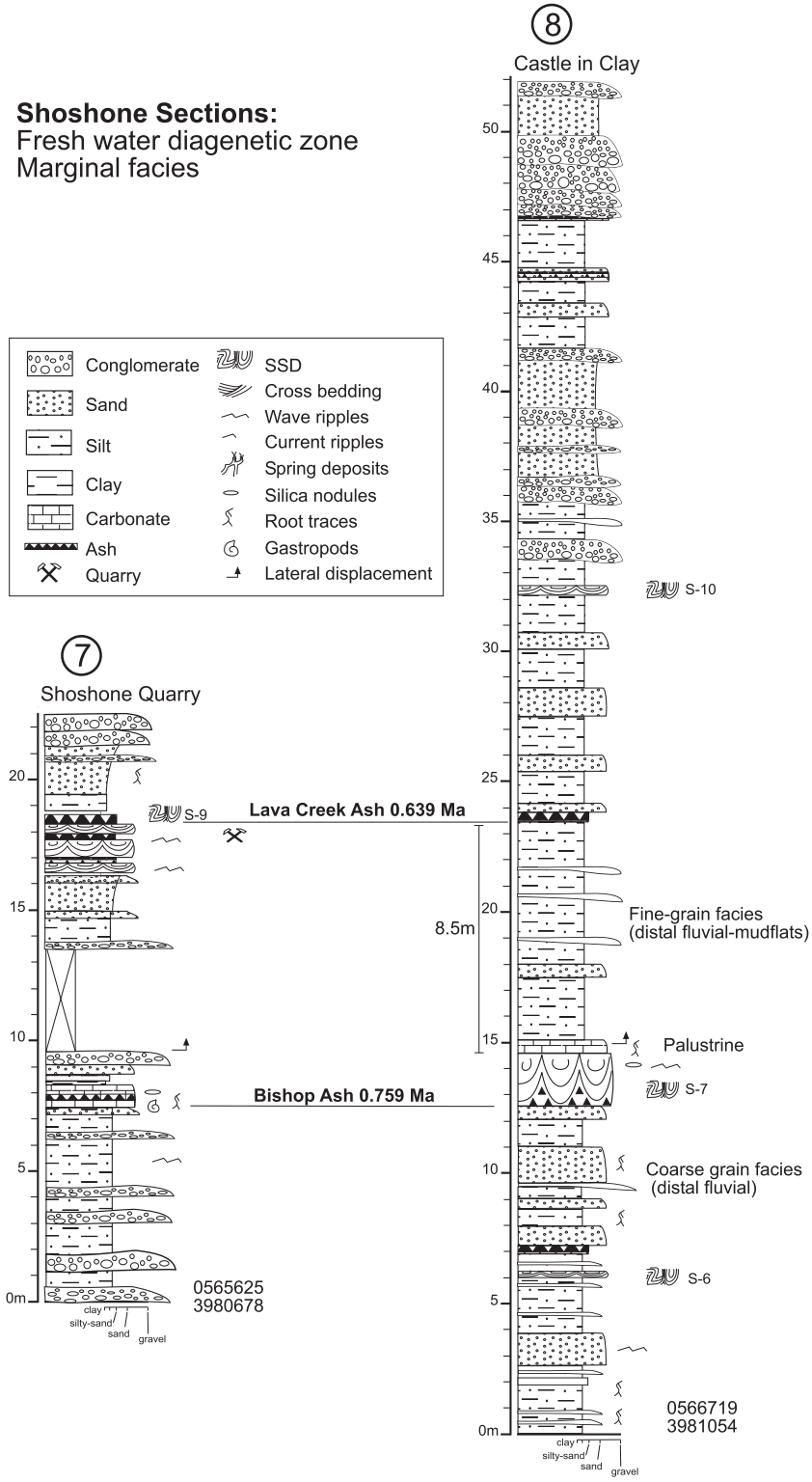

Fig. 5. Stratigraphy around Shoshone town, near the pumicite quarry and Castel and Clay area. Both sections includes the Bishop and Lava Creek, in marginal facies associated to fresh water environments (palustrine and fluvial) occurring during the Pleistocene in the North Margin of the basin. The figure shows the location of beds containing SSD and ash layers and UTM position.

2002), sudden changes in groundwater level (Owen, 1987), and earthquakes (Seilacher, 1969; Obermeier, 1996).

After a facies analysis we can discount overloading since overlying units do not appear related to rapid sedimentation, being typically composed of thin laminae from several millimetres to tens of centimetres in thickness. We also discount the effect of water waves, given the low wave-energy conditions of the Tecopa paleolake with its small areal extent. Finally, these soft-sediment deformation structures are morphologically very different from those produced by groundwater movements (Obermeier, 1996).

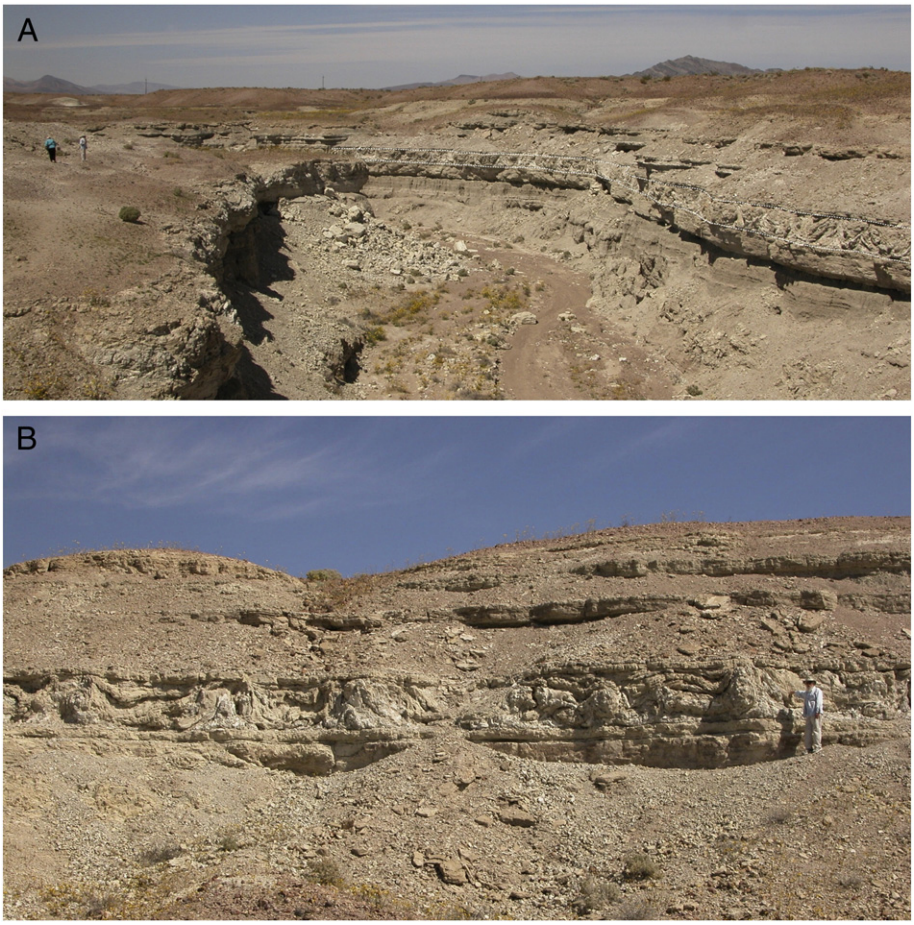

Fig. 6. A. Panoramic view of the Bishop Ash deformed beds, marked with a white dashed line. B. Detail of the Bishop ash deformed bed with large load-type structures separated by narrow fluidization channels.

We interpret the soft-sediment deformation structures described in the Bishop and Lava Creek Tuffs (Tecopa Basin) as seismites (sensu Seilacher, 1969). A seismic origin is also favoured because: (1) deformation structures continue laterally for hundreds of metres, occurring in the central and marginal facies and being overlain and underlain by planar units not involved in soft-sediment-deformation, and (2) the study area is in a region subjected to moderate to large earthquake shaking events.

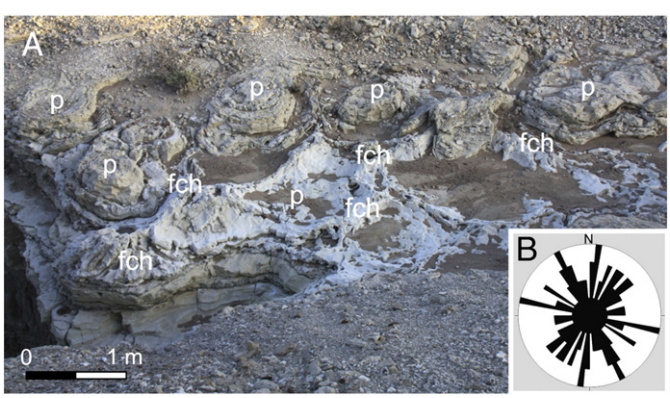

Fig. 7. A. Plan view of the deformed Bishop Ash. Elliptical pillows (p) are separated by narrow fluidization channels (fch). Observe the reverse density gradient system at this outcrop characterized by a brown overlying unit represented by fine sand and an underlying unit by light grey fine grain ash and mud. B. Rose diagram showing the direction of major axis showing no preferred orientation. N = 76, largest petal: 4.00 values, 5% of all values.

The Tecopa region was tectonically active during the Plio-Pleistocene. Nearby faults are responsible for historical earthquakes in the vicinity (Jennings, 1994). In the Tecopa Basin, several local faults have potentially generated earthquakes during Pleistocene. Additionally, other large faults in Death Valley and Panamint Valley, as well as the Garlock Fault zone (Jennings 1994), could be important. Most of these faults are large and have the seismic potential to generate earthquakes with magnitudes >7.0. These high-magnitude earthquakes could produce liquefaction at distances from the epicentre of between 100 and 200 km, as can be deduced from empirical relationship between magnitude and maximum epicentre distance for liquefaction sites proposed by Ambraseys (1988).

5.2. Significance of superposed deformed beds

A common morphological feature of both the Bishop and Lava Creek Tuff-rich deposits is that the main deformed beds are divided, at least locally, into several superposed deformed layers separated by thin laminae of fine-grained sediment. These minor deformed units, with smaller load-type structures, can be seen to merge laterally into a single deformed bed, apparently where more favourable conditions existed for the formation of large load-type and fluid-escape structures.

The following field observations indicate that a single shaking event or closely spaced shaking events caused liquefaction that produced superposed deformed structures.

- Large fluid-escape structures usually cut through the entire thickness of the deformed bed. Fluidized sediment flowed from the

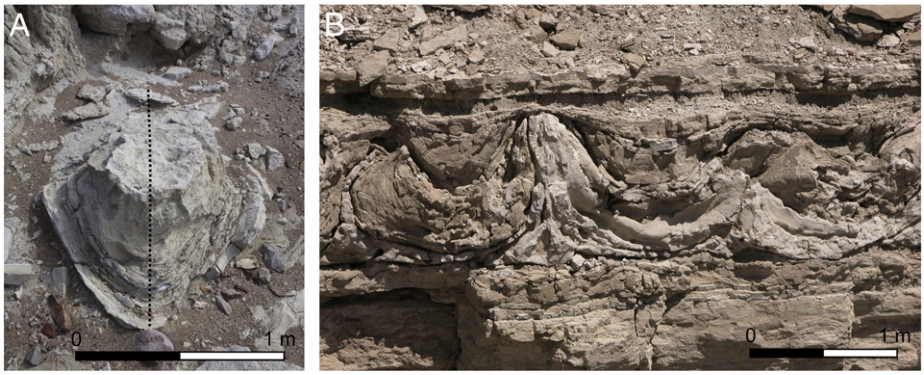

Fig. 8. A. Three-dimensional exposure of a load structure showing its internal structure. The black dotted line indicated the major axis of the ellipse in plan view. B. Detail of the deformed Bishop bed, showing to the left of the picture unconformable surfaces between laminas of superposed load structures.

lower layers, past the overlying beds, until reaching and flowing onto the surface existing at that time. These large fluid-escape structures were formed after the uppermost involved layer was initially deposited. Deformation is also constrained to have terminated just after the accumulation of the horizontal layer of sediment transported by fluidization, and the next bed being deposited on top of the deformation package.

- Thin superposed layers showing deformation are separated by undeformed laminae of silty sediment in some areas. However, laterally these same laminae were cut by fluid-escape structures. Silty laminae that separate ash-rich layers acted as local barriers to permeability during liquefaction (sensu Moretti et al., 1999).

These specific internal structures in the main deformed beds, composed of several superposed thinner deformed units, are interpreted as the product of earthquake shaking. The internal laminae of fine-grained sediments acted as permeability barriers to produce detached load-type structures. According to experimental results, load-type structures begin as detachments from an upper layer, collapse and move down into the underlying layer with a lesser bulk density (Kuenen, 1958). The style of deformation described in our study area is similar, showing pillows and drop structures. Our observations support a model where several superposed load-type structures were detached simultaneously during a major earthquake, with the lateral formation of larger, unified load and fluid-escape

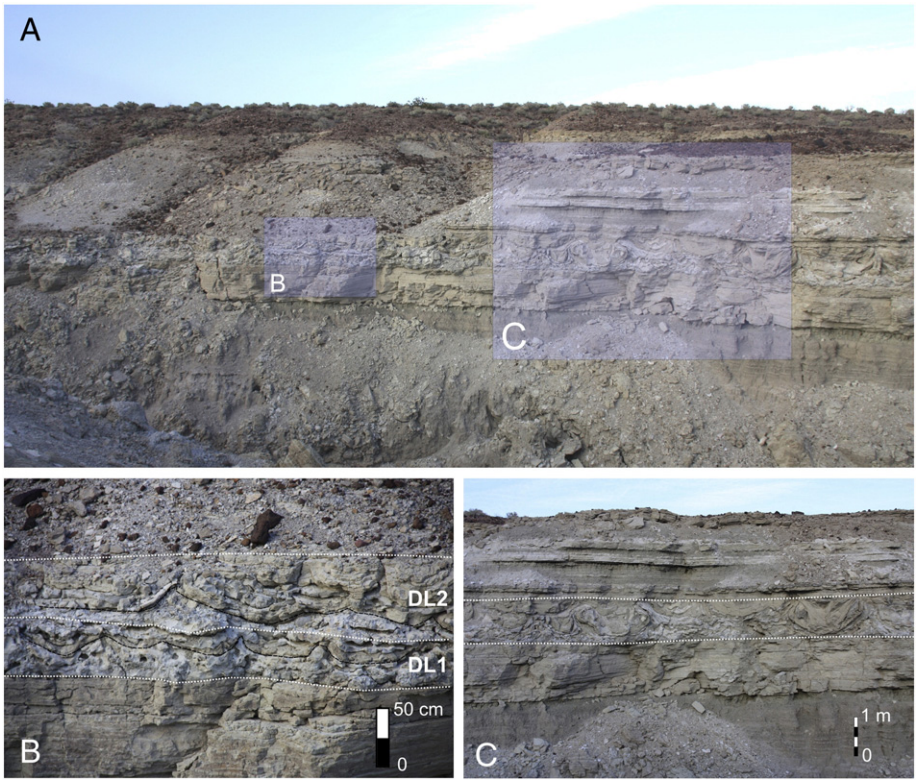

Fig. 9. A. Panoramic view of the Bishop Ash deformed bed where single large load structures pass to two superposed beds with smaller load structures. Note the different scales for comparison. B. The Bishop Ash bed is composed of two deformed layers with load casts structures with less than 1 m in size. Observe the presence of several intercalated laminas of light grey fine grain ash and mud. C. Transition zone between large load structures (more than 2 m in width) to the right and superposed load structures separated by unconformable surfaces, to the left.

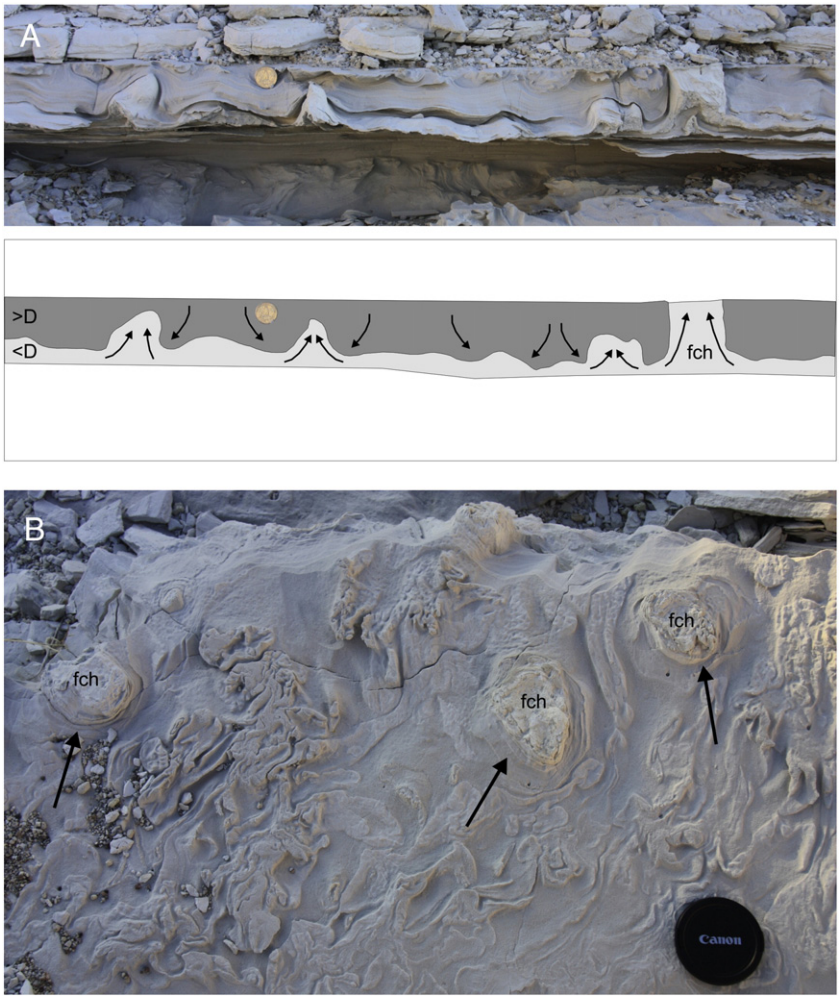

Fig. 10. Several examples of centimetric soft-sediment deformation structures in Lava Creek ash bed. A. Vertical section of a thin deformed layer (less than 10 cm in thickness, see coin for scale), with pillowy load structures. A narrow fluidization channel is observed to the right. B. Plan view of a deformed layer where is possible to recognize the circular to elliptical morphology of the fluid-escape structures (indicated by black arrows).

structures. The examination of the SSD structures in the Tecopa outcrops indicates that the original internal layering of a deposit can control the final morphology of the soft-sediment deformation structures produced. Within each ash-rich deposit, the number of superposed deformed layers containing detached structures depends on the number of heterogeneities. However, these thin layers can be completely engulfed and deformed if large parts of the lower deposit become liquefied and move upward. Lateral changes from simple to multilayered reverse density gradient systems could be responsible for these local variations in the number of superposed deformed beds (Fig. 14).

Lacustrine paleo-environments appear to be ideal places to record past seismic events. However, we would caution again a direct calculation of recurrence intervals using only a sequence of seismites.

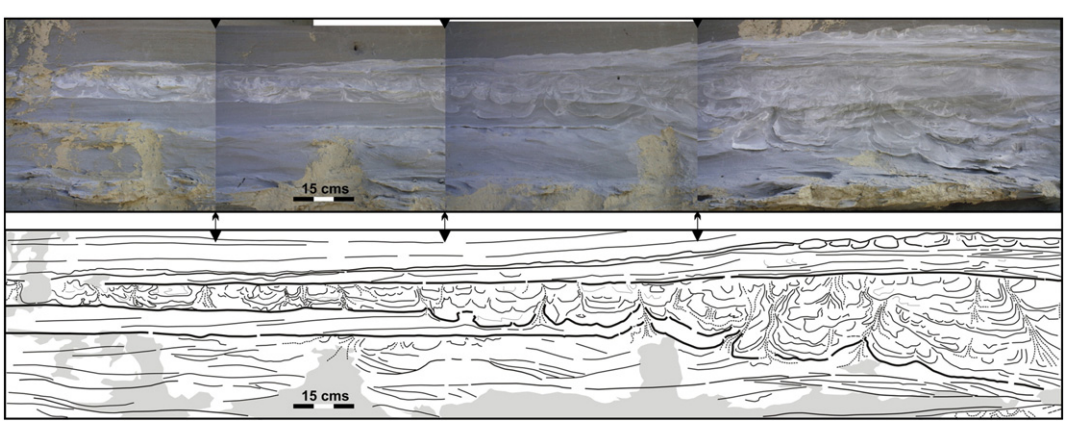

Fig. 11. Complex deformation of the Lava Creek ash bed (east wall of the quarry next to Shoshone with a N–S trend). Observe superposed load structures to the left with several non deformed layers and fluid-escape structures which cut all the deformed bed to the right.

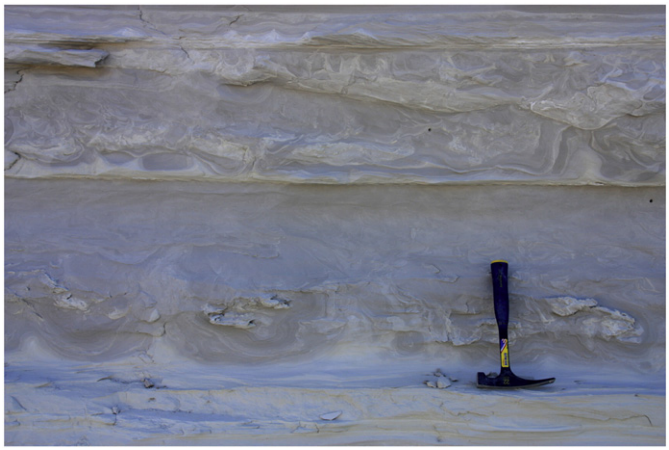

Fig. 12. Several superposed deformed layers with centimetric to decimetric load structures. Observe the horizontal thin layer approximately at the center of the picture which separates two main deformed beds. Laterally (see Fig. 12), this horizontal thin layer is cut by a large fluid-escape structure.

Firstly, seismites can only develop under conditions adequate for liquefaction in saturated sediments. In shallow lakes, temporarily non-saturated areas are common during dry seasons and this would allow many shaking events to go unnoticed. A second difficulty in calculating recurrence intervals is revealed in this report only because of the extensive exposures. We have shown that lateral changes in the internal layering of a single ash-rich deposit will produce lateral changes in the number of deformed beds.

A shaking event can produce a single deformed bed characterized by large soft-sediment deformation structures in simple reverse density gradient systems but, a few metres away, if the sedimentary column passes to a multilayered reverse density gradient system, this deformed bed is divided into two (Bishop Tuff outcrop) or more (Lava Creek Tuff) superposed deformed beds. This difficulty would be pronounced in studies with limited exposures (such as in drill cores) and could cause confusion in counting liquefaction events.

6. Conclusions

The Pleistocene sedimentary record of the Tecopa Basin is characterized by several ash-rich beds and other deposits showing pervasive soft-sediment deformation. Two deformed levels were examined in detail: those related to the Bishop and the Lava Creek Tuffs. These beds could be studied around and across the basin, where they were deposited in different paleo-environments, and subjected to different diagenetic processes. Where deformed, the Bishop Tuff is characterized by load-type structures up to 1.5 m thick. Similarly, the Lava Creek Tuff is characterized by load-type structures with local fluid-escape structures up to 1 m high.

We interpret these soft-sediment deformation structures as seismites, formed during seismic shaking events. Specific conditions favoured their widespread development in the Tecopa Basin: (1) the accumulation of ash and silts provided unconsolidated sediments with reverse density gradients; (2) the more central parts of paleo-Lake Tecopa provided a saturated groundwater environment necessary for liquefaction; and (3) the region had numerous fault zones that were active during the time of deposition.

Both ash-rich beds were traced laterally, where we observed changes in the extent of soft-sediment-deformation. A single deformed deposit merged laterally into multiple superposed deformed beds, before then merging into an undeformed deposit. A morphological and structural examination suggests that this lateral variation in deformation style was the result of a single shaking event (probably from a major earthquake, plus aftershocks). The presence of heterogeneities in the deformed ash-rich sedimentary deposit, especially thin layers of finer sediment (barriers to liquefaction), acted as partitions between deforming horizons. Laterally, where more extensive deformation occurs, large fluid-escape structures cut and fold these thin layers which act as passive beds.

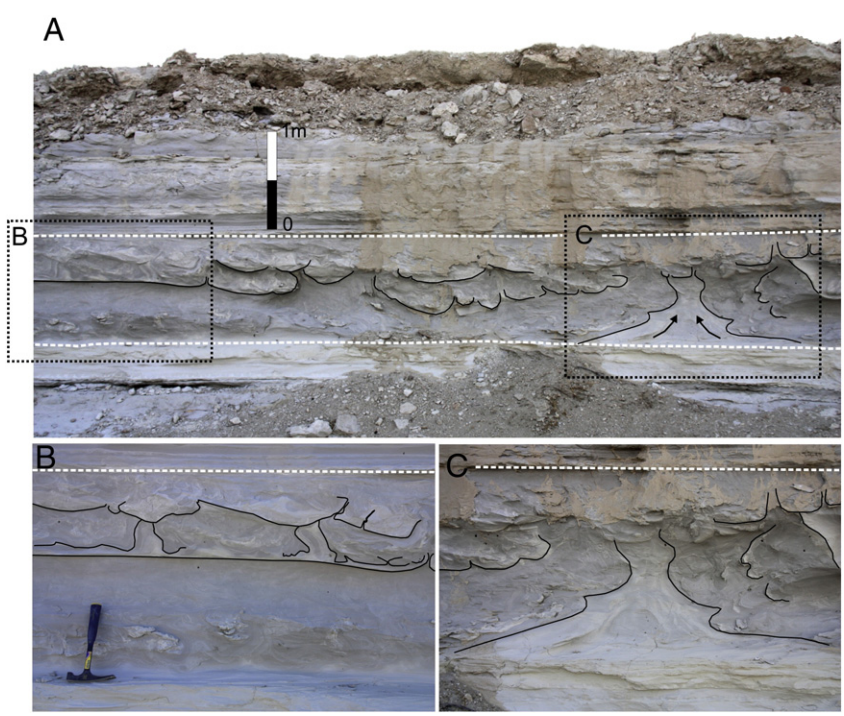

Fig. 13. A. Complex deformation of the Lava Creek ash bed (west wall of the quarry next to Shoshone), where small superposed deformed beds (B) pass to an area with large load and fluid-escape structures (C).

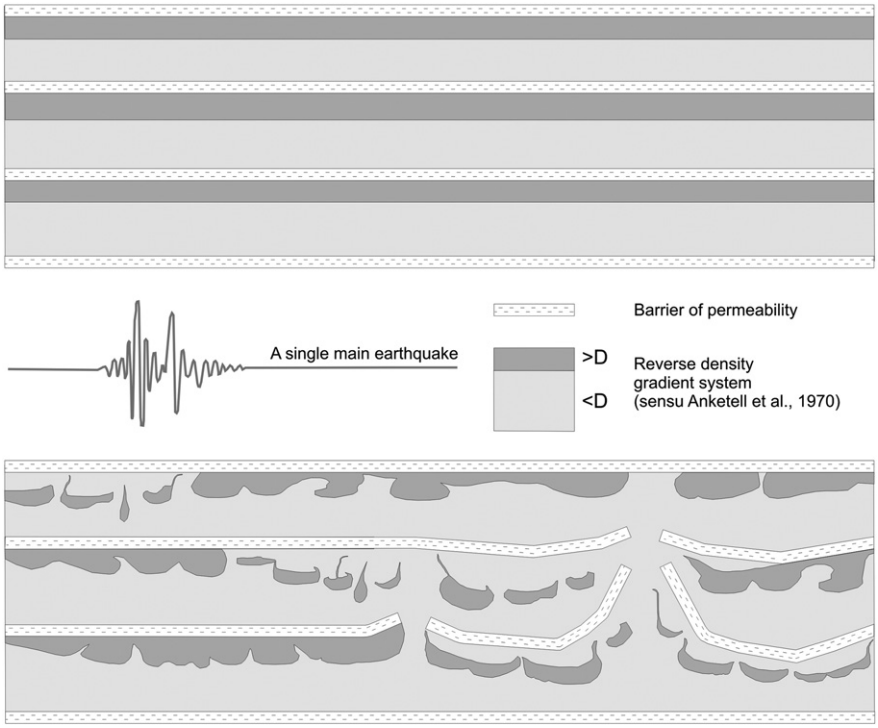

Fig. 14. Simple genetic sketch showing an initial sedimentary column with reverse density gradient systems separated by thin laminas of fine sediment. A single earthquake produces superposed deformed beds of load and fluid-escape structures. The thin layers of fine sediment favour the detachment of underlying load structures (load casts, pillows and drop structures).

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Seismically induced slump on an extremely gentle slope (<1°) of the Pleistocene Tecopa paleolake (California)

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ABSTRACT

A superbly exposed ~5 km² slump is developed in middle Pleistocene lacustrine deposits of the Tecopa Basin (California, United States). A subhorizontal detachment (<1°) with a single sheet ~8 m thick is displaced eastward nearly 100 m. The different parts of the slump are continuously exposed for more than 1 km, with extensional structures in the head zone and contractional features in the intermediate and toe zones. Meter-scale normal faults and thrusts coexist with meter- to decameter-scale open folds, all of them detached, on a clay bed rich in sepiolite. These well-exposed deformation structures were produced by liquidization of the clay bed, due to the thixotropic behavior of the sepiolite. Processes such as oversteepening or sedimentary overloading are discarded in light of sedimentological and paleoenvironmental analyses. Thixotropic behavior of sepiolites might have been caused by seismic shaking related to regional faults that were capable of producing large earthquakes during the middle Pleistocene (between 579 and 200 ka).

INTRODUCTION

Slumps are very frequent in subaqueous or subaerial lacustrine environments. Numerous examples have been studied from field observations (Gibert et al., 2005), cores (Bertrand et al., 2008), and/or seismic profiling investigations (Niemi and Ben-Avraham, 1994; Chapron et al., 2004), and studies have included structural analyses (Woodcock, 1976).

Several autogenic trigger mechanisms (directly associated with sedimentation processes and patterns, according to Owen and Moretti, 2011) can generate slumps in lakes, such as oversteepening or sedimentary overloading (Lewis, 1971; Pettijohn and Potter, 1964; Allen, 1982), but many slump deposits have been attributed to allogenic trigger mechanisms (externally acting agents), such as earthquakes. There is an extensive literature on seismically induced slumps, with examples in the geological record (Gibert et al., 2005; Karlin et al., 2004), related to historical earthquakes (Schnellmann et al., 2002), or recent earthquakes (Niemi and Ben-Avraham, 1994).

Studies of modern sediments show that slumps can occur on very gentle slopes (Wells et al., 1980; Field et al., 1982). In the geological record, however, the occurrence of slump sheets is usually taken to indicate the presence of a significant slope (Lewis, 1971; Allen, 1982). We report here an example of a slump in lacustrine deposits of middle Pleistocene age (579–200 ka) that moved over a surface sloping at <1°.

Although thixotropy has been considered as a potential mechanism of deformation, few examples have been documented from the geological record. This slump described in this paper occurred due to the thixotropic behavior of a sepiolite-rich horizon, triggered by seismic activity. The slump is also interesting because of the exceptional quality of exposures, which allow a detailed description of its internal structure and estimate of its displacement and transport direction.

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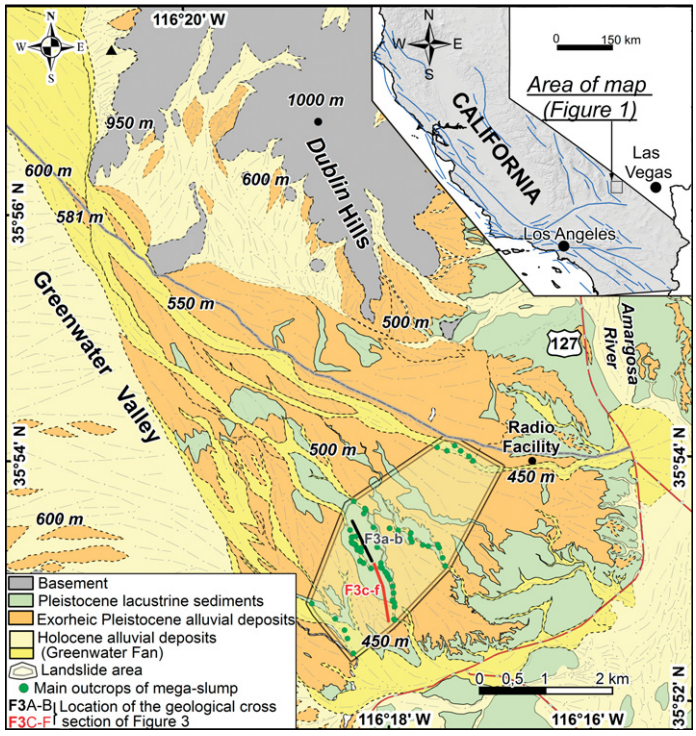

Figure 1. General geological and geomorphological map of Greenwater fan (western Tecopa Basin, California) (after Hillhouse, 1987) with indication of slump area, main outcrops, and geological section of Figure 3.

PLEISTOCENE TECOPA PALEOLAKE

The study area is in the Basin and Range region, 50 km southeast of Death Valley in the Amargosa Desert (California, USA; Fig. 1). The Tecopa Basin has an area of ~500 km² and is surrounded by the Dublin and Ibex Hills (west), the Resting Spring Range (east), and Sperry Hill (south), which are composed of Proterozoic and Cambrian rocks and late Cenozoic volcanic flows (Hillhouse, 1987). After the closed Tecopa Basin was captured ~0.2 m.y. ago (Morrison, 1999), subsequent erosion exposed a sedimentary infill >90 m thick. The stratigraphic sequence consists mainly of mudstones, carbonates, and alluvial deposits, with a general basinward dip of <1° (Sheppard and Gudde, 1968; Hillhouse, 1987) and numerous intercalated rhyolitic vitric ash layers. Two of these have been chemically correlated with ultra-Plinian eruptions in the western United States, the Bishop Tuff (759 ka), and Lava Creek Tuff (639 ka) (Izett, 1981; Sarna-Wojcicki et al., 2000) (Fig. 2). Starkey and Blackmon (1979) reported sepiolite as the dominant authigenic clay-type mineral, found mainly near ash beds and probably precipitated when silica became abundant, through dissolution of volcanic ash in the Mg-bearing lake waters.

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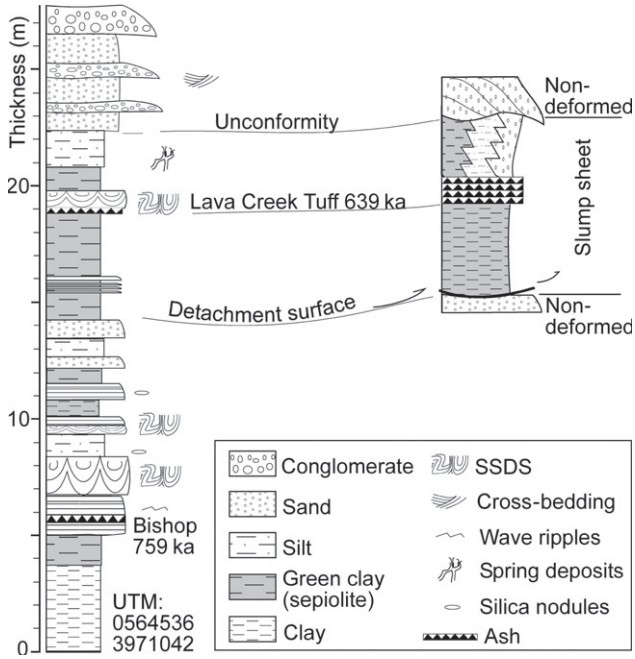

Figure 2. Upper Greenwater stratigraphic section showing location of Bishop and Lava Creek Tuffs (California). Detail of slumped deposits is also shown. Soft-sediment deformation structures (SSDS), interpreted as seismites in Gibert et al. (2011), that are incorporated into slump sheet relate to previous earthquake. UTM—Universal Transverse Mercator.

The Tecopa Basin has been interpreted as a central shallow saline lake surrounded by freshwater ponds located near springs and fluvial inlets (Gibert et al., 2011). Lake Tecopa underwent fluctuations in water level and salinity according to the hydrologic budget (Larsen, 2008). During dry periods, sediments accumulated in the central area by groundwater-fed discharge or occasional floods, while a shallow permanent lake was established during pluvial episodes.

TECOPA SLUMP

The slump sheet, in the Greenwater Valley, is well exposed along several subparallel streams (Fig. 1; see the GSA Data Repository[1]). We selected the main Greenwater stream section to construct a geological cross section (Fig. 3) because of its nearly continuous exposures.

The average thickness of the slumped unit is ~8 m and its sedimentary succession varies from head to toe of the slump. Features found throughout include (Fig. 2) (1) a nondeformed substratum composed of a competent bed, (2) a detachment layer mainly composed of sepiolite, (3) a main stratigraphic marker of deformation, the Lava Creek ash bed, (4) an upper deformed unit that varies laterally in grain size from gravel to clay, and (5) an undeformed alluvial unit that unconformably overlies and postdates the slump.

Geometry

The ~5 km^2 Tecopa slump has an ellipsoidal morphology, with a major axis of ~3.5 km and a minor axis of ~1 km (Fig. 1). It comprises a single slope failure and represents a discrete slump sheet. High-precision altimeter data indicate that the detachment surface has a very low angle paleoslope (<1°).

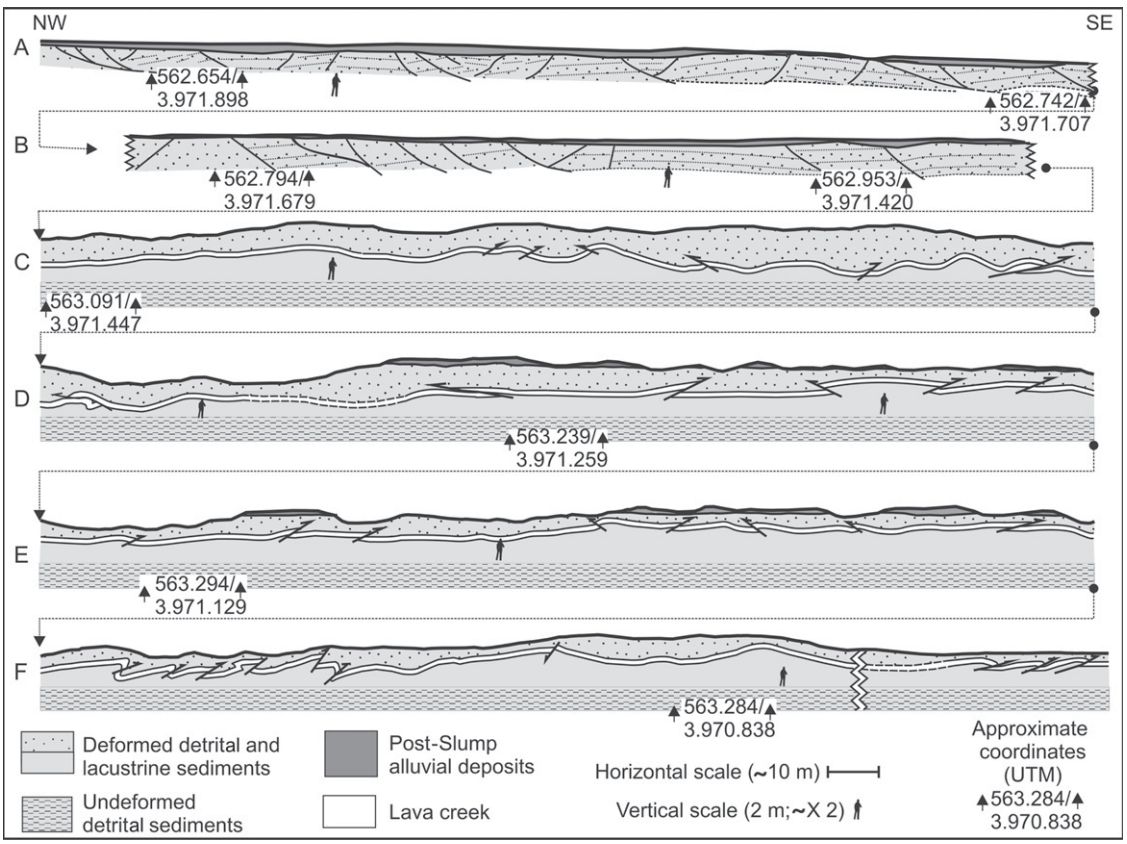

Figure 3. Geological cross section of Tecopa slump (California) based on field geological cross sections and panoramic pictures along main Greenwater stream (see location in Fig. 1). Geological cross section is divided into six segments (A, B, C, D, E, F). Outcrops are continuous from C to F, while minor discontinuities occur between A, B, and C. Georeferenced points are included, using Universal Transverse Mercator World Geodetic System (UTM WGS84) coordinates.

[1]GSA Data Repository item 2011309, supplemental information (Figures DR1–DR3) with various details of the slump, is available online at www.geosociety.org/pubs/ft2011.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

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From northwest to southeast, the slump can be divided into a head characterized by extensional features, a less deformed intermediate zone, and a toe characterized by compression. The slump head is a complex scar zone with numerous southeast-dipping high-angle normal faults, mostly with a listric geometry. The throw is variable, from a few centimeters to several meters. Antithetic fractures develop small horst and graben structures. There are many NE-SW–trending injection dikes filled with fluidized fine sediment. Friable tufa deposits record the discharge of regional groundwater directly into Lake Tecopa during wet climate periods (Nelson et al., 2001).

Southeast of the head zone, extensional structures pass into an intermediate zone with gentle folds (Figs. 3 and 4), and a toe zone with a wide variety of folds and thrusts. The Lava Creek bed forms an excellent deformation marker (Fig. 4). Most folds are asymmetric, with wavelengths to 30 m, and thrusts show a modest displacement of a few meters (Figs. 3 and 4). A series of thrusts repeats the Lava Creek bed in an imbricate structure (Fig. 4C). Folds in the central zone are northwest and southeast vergent, and thrusts and backthrusts predominate. In the toe zone, most compressional structures are southeast vergent.

Age

The Bishop (739 ka) and Lava Creek (629 ka) Tuffs give good chronological control to the upper part of the Greenwater Valley sequence. Interpolation of the two beds leads to estimated sedimentation rates of 0.1 mm/yr. Since the Lava Creek Tuff occurs 5 m below the top of the slump, we estimate an age of 579 ka (5000 mm, 0.1 mm/yr) for the youngest slumped sediments. Because the slump is overlain by alluvial fan sediments deposited when the basin was more open (exorheic), and assuming that the Tecopa Basin was captured ca. 200 ka (Morrison, 1999), the age of the slump can be constrained between 579 and 200 ka. Two highstands in Lake Tecopa during this period, related to glacial episodes (Larsen, 2008), could have favored slumping.

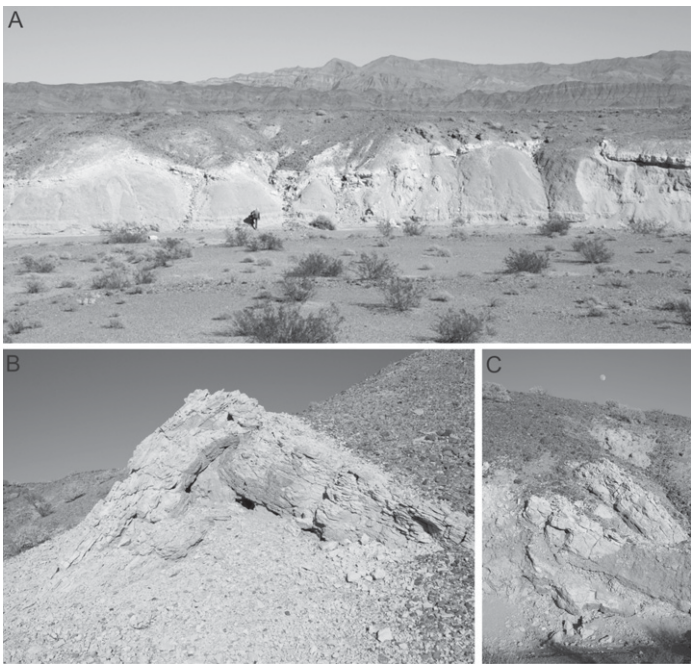

Figure 4. A: Panoramic view of toe (compression) zone, with large open folds and thrusts (Universal Transverse Mercator, UTM 563969, 3971741). B. Detail of fold-thrust deforming the Lava Creek Tuff (UTM 562590, 3970375). C. Detail of two thrusts that repeat Lava Creek Tuff in imbricate structure (UTM 562546, 3970502).

Reconstruction of the Slump

The continuous exposure of the compression zone allows the reconstruction of the original length of the slumped body (Fig. 3), and indicates a shortening of ~90 m.

Kinematic indicators of the transport direction include injection dikes, extensional joints, and striations. Injection dikes in the head zone with a dominant trend from N10E to N30E, and an extensional joint set with an average N25E strike indicate N100–120E extension, which is presumed to be parallel to the main transport direction. Striations in the Lava Creek ash bed with a dominant N115E trend may relate to shear stresses in the slumped body parallel to the main displacement of the slump. Although slight deviations are seen in the data, all indicators suggest southeast transport, which is consistent with the mapping of the slump (which indicates N130E transport), and with the basin geometry, with its depocenter southeast of the slump area.

The location of the scar zone of this slump could be related to the presence of friable tufas composed of disaggregated deposits. Some of these are characterized by multiple injection dikes that crosscut lake sediment (Nelson et al., 2001), and could have acted as a zone of weakness.

CAUSES OF SLUMPING

Driving Force and Mechanism of Deformation

The driving force for slumping is gravitationally induced shear, although in this particular case, the extremely gentle slope (<1°) generated low shear stresses. Nevertheless, studies in modern environments show that slumping can occur on subaqueous slopes as low as 0.25° (e.g., Field et al., 1982).

The detachment layer of the slump is a clay bed rich in sepiolite. Sepiolitic clays are thixotropic, undergoing liquidization when agitated (Strachan, 2002) and flowing as a plastic with high viscosity. When the sepiolite-rich clay bed liquidized, it allowed downslope movement of this multilayer slump, despite the gentle angle of the slope. During translation the slumped body behaved in a ductile-fragile manner, producing normal faulting at the head, and folding and thrusting at the toe.

Trigger Mechanism: Discussion

Several trigger mechanisms have been invoked to explain slumping in lakes, including those that are part of the depositional environment (autogenic, according to Owen and Moretti, 2011), or those that are external (allogenic). Among the most common autogenic processes in lacustrine environments are sedimentary, erosional, or tectonic oversteepening and sedimentary overloading (Pettijohn and Potter, 1964; Allen, 1982). Among allogenic causes, many slumps are attributed to a seismic trigger.

Postdepositional tectonic tilting can increase the slope angle, and therefore the shear stress on a surface, but the present subhorizontal geometry of the Tecopa slump substratum (<1°) leads us to discard tectonic oversteepening as a trigger.

Sedimentary oversteepening can act as a trigger through the continuous shear stress exerted by the downslope component of the weight of overlying sediment, related to progressive sedimentation. The magnitude of the shear stress increases with the slope angle and with the density and thickness of the overlying sediment (Lewis, 1971). Taking into account the low sedimentation rates (0.1 mm/yr) and the low paleoslopes of these lacustrine deposits, this slump is unlikely to have been triggered by sedimentary oversteepening or sedimentary overloading.

Field observations indicate that exorheic deposits are not involved in the slump. Therefore, it can be concluded that the slump occurred when the Tecopa was a closed basin. If this assumption is correct, erosional oversteepening can be discarded.

Once all of the possible autogenic trigger mechanisms have been discarded, we are left to interpret this slump as result of an earthquake

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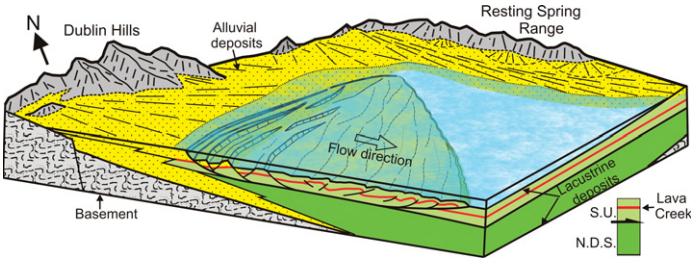

Figure 5. Three-dimensional sketch model for development of Tecopa slump. It is assumed that it was produced during highstand in lake that covered detrital subaerial deposits of most marginal area (according to Larsen, 2008, lake levels fell and rose several times after emplacement of Lava Creek Tuff). SU—slump unit; NDS—non-deformed substratum.

(Fig. 5). The presence of a sepiolite-rich layer with the potential for thixotropic behavior is consistent with this interpretation, allowing slumping on an extremely gentle detachment surface (<1°). Additional criteria support this interpretation: (1) seismically induced slumps have been documented from recent earthquakes (Niemi and Ben-Avraham, 1994); (2) there are several active faults in the region with enough seismic potential to have produced moderate- to high-magnitude paleoearthquakes, which could have generated liquidization of the sepiolite-rich layer; and (3) other middle Pleistocene paleoearthquakes have been recorded in Bishop and Lava Creek Tuffs (Gibert et al., 2011).

CONCLUSIONS

An extraordinarily well exposed ~5 km^2 slump occurs in Pleistocene lacustrine deposits of the Tecopa Basin (California). A single slump sheet ~8 m thick is characterized by extensional features at the head, a less deformed intermediate zone, and contractional features such as folds, reverse faults, and thrusts at the toe. The deformation mechanism is related to thixotropic behavior of a basal clay bed rich in sepiolite. From the present geometry of the contraction structures we estimate an approximate shortening of 90 m, which indicates transport of a few dozen meters. Mapping of the slump and kinematic indicators such as injection dikes, tensional joints, and striations indicate an average transport direction of N110–130E, consistent with the basin geometry.

Sedimentary and paleoenvironmental analysis of this extremely gentle sloping (<1°) slump leads to the rejection of autogenic sedimentary processes as a trigger mechanism. It is very likely that an earthquake triggered this slump between 579 ka, the age of the most recent deformed sediments, and 200 ka, the age of the exorheic deposits that postdate the slump.

ACKNOWLEDGMENTS

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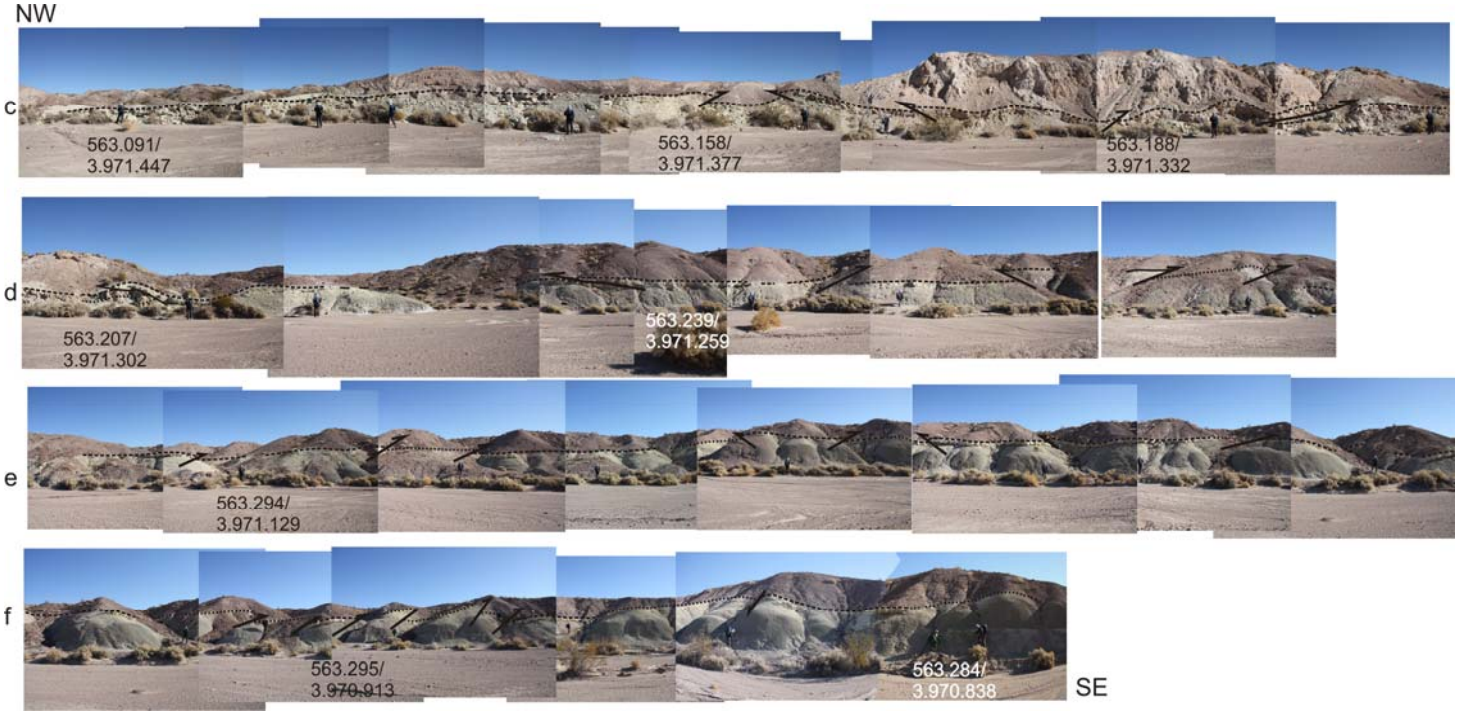


Figure DR1. Photo-mosaic along the main Greenwater stream. This stream offers a nearly continuous exposure of the slump. We have combined this information with field data to elaborate a geological-cross section (Fig. 3).

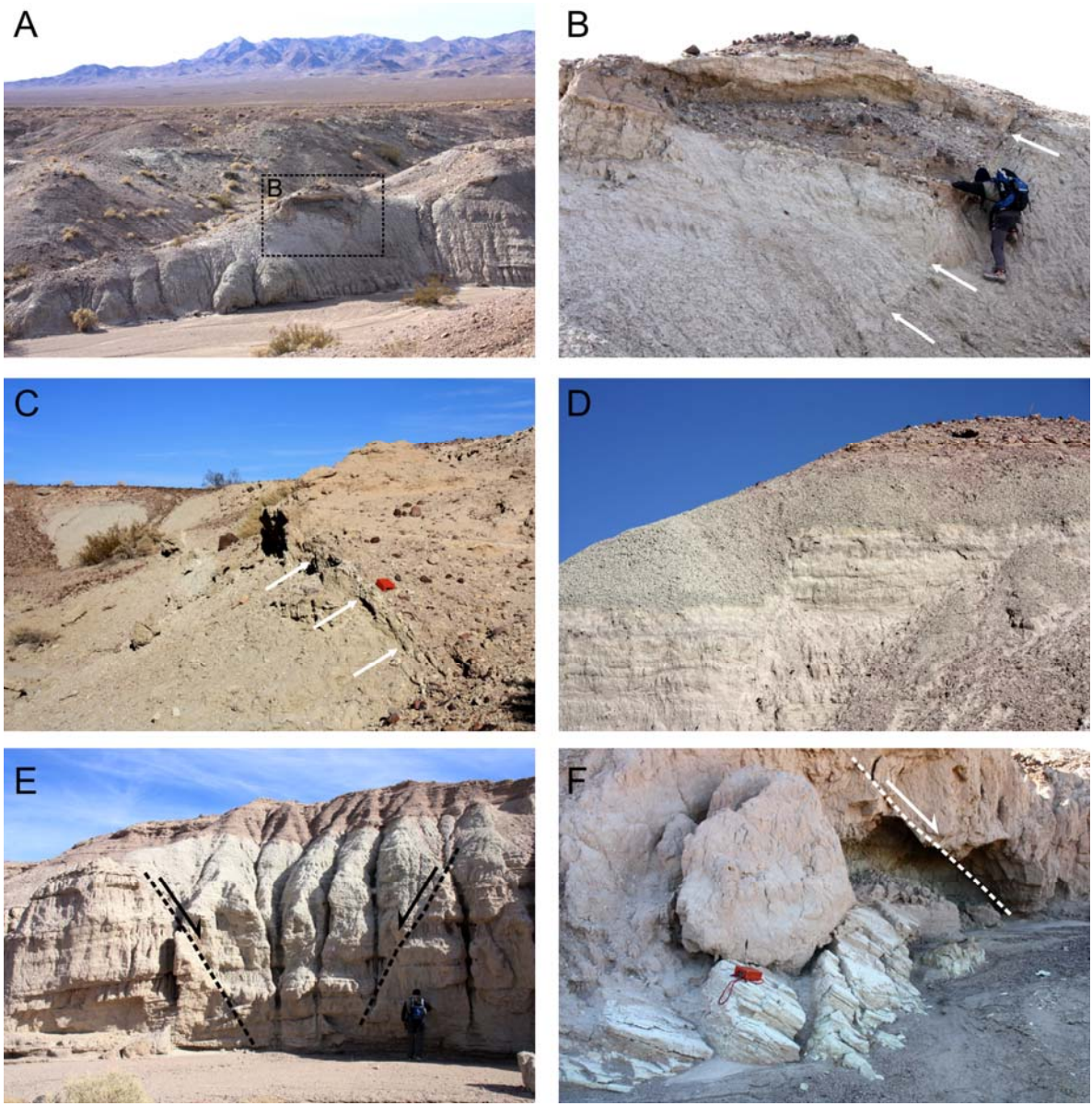


Figure DR2. Extensional features identified at the head zone. A. Listric normal fault. Square dashed line indicates location of figure 2B. B. Detail of the extensional fault of figure 2A. White arrows indicate the fault trace. C. Scar at the head zone indicated by white arrows; the hanging wall is located to the right of the scar. D. Normal fault with a metric throw. E. Graben formed by conjugate faults with metric throw. F. Antithetic (dashed white line) and synthetic normal faults deforming Lava Creek ash bed.

Figure DR3. Some kinematic indicators of the transport direction. A. Extensional joint set at the top of the Lava Creek ash bed. B. Striations in the Lava Creek ash bed, probably produced by shear stresses into the slumped body. C. Injection dyke, where red silt is fludized into white and grey lacustrine sediments.

Appendix A2.2 (Caskey)
Figures for purposes of discussions about the timing character of the Terminal Lake highstand and apparent Terminal lake still stand 430-435 m. Implications for seismic slumps?

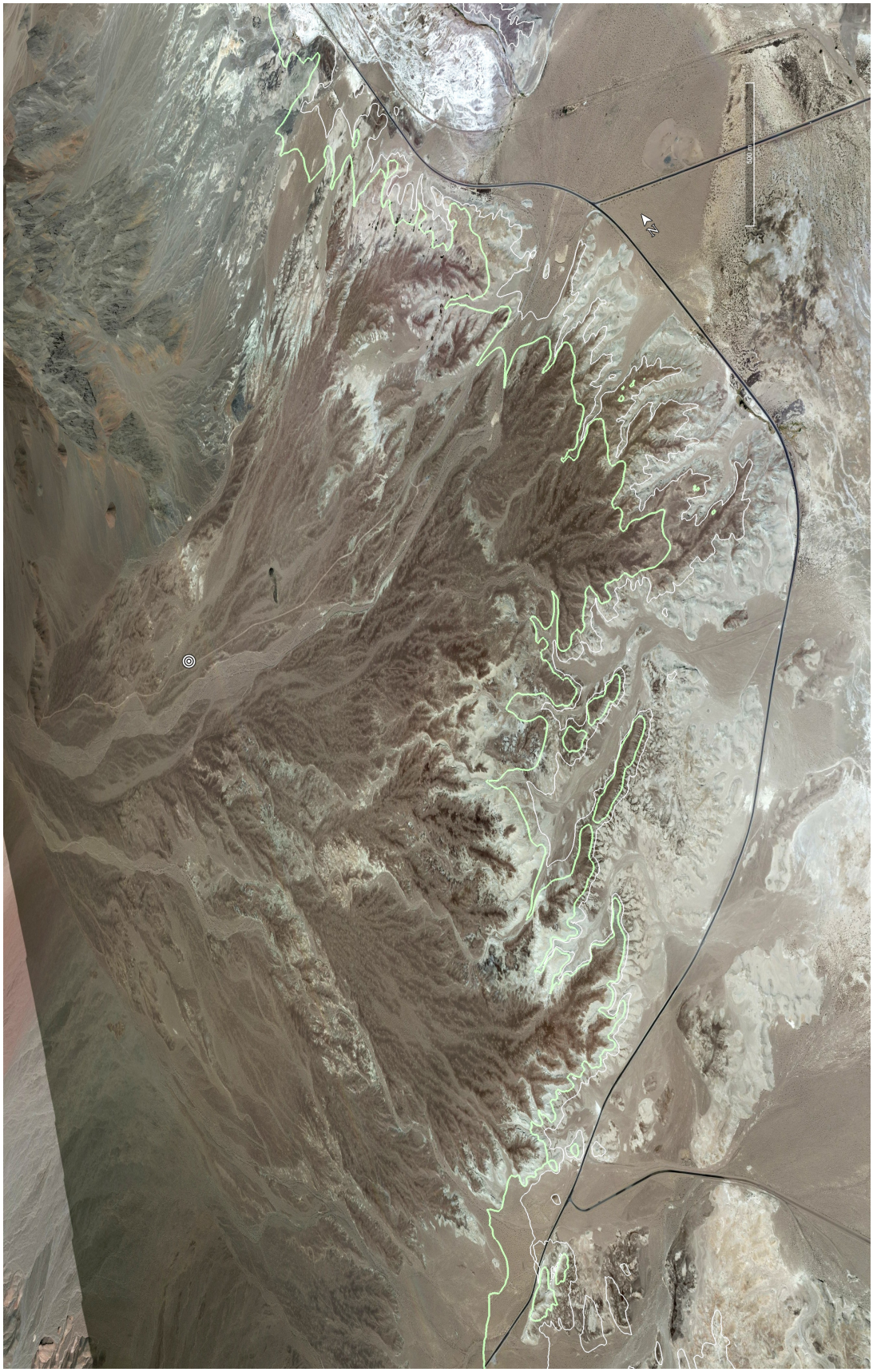


Fig. A2.2a

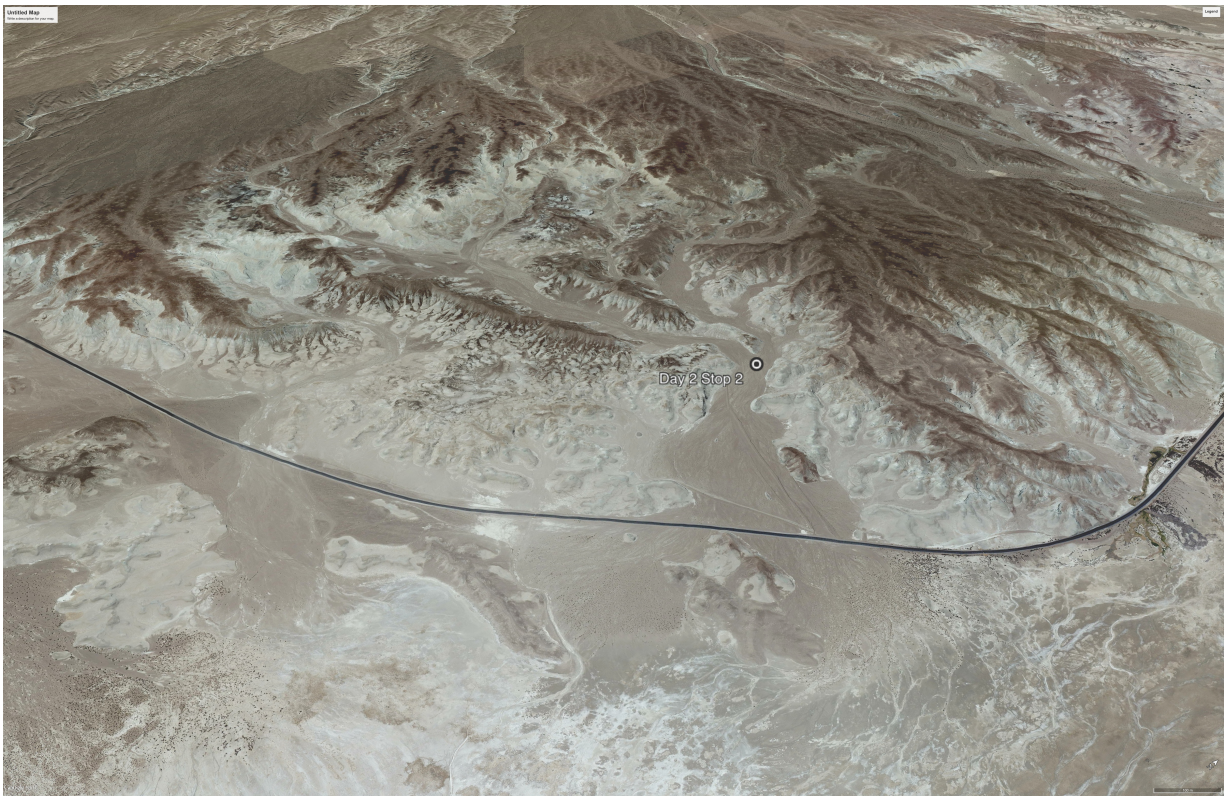


Figure A2.2a previous page and b– Large scale slumps at the toe of the Greenwater fan. The slumps appear to have formed on an older ,more dissected fan unit than the fan at Stop 2.1. The sill elevations for the High Lake and terminal lakes must have similar so a brief stand for terminal Lake Tecopa could reflect weaker threshold conditions for a breach. One initiated the cutting of the canyon could have occurred rapidly to only about an elevation of about 430-435 m, at which time a longer–lived terminal lake stand could occur, but at a much lower level. It may be that wave erosion at the time of the lower lake stand destabilized the lower Greenwater fan. Such a scenario would also have occurred when the ground was saturated or at least more wet immediately following lake withdraw. Ideas for discussion anyway.

Fig. A2.2b

Fig. A2.2c **Contour relief map of the Tecopa basin (Caskey)**

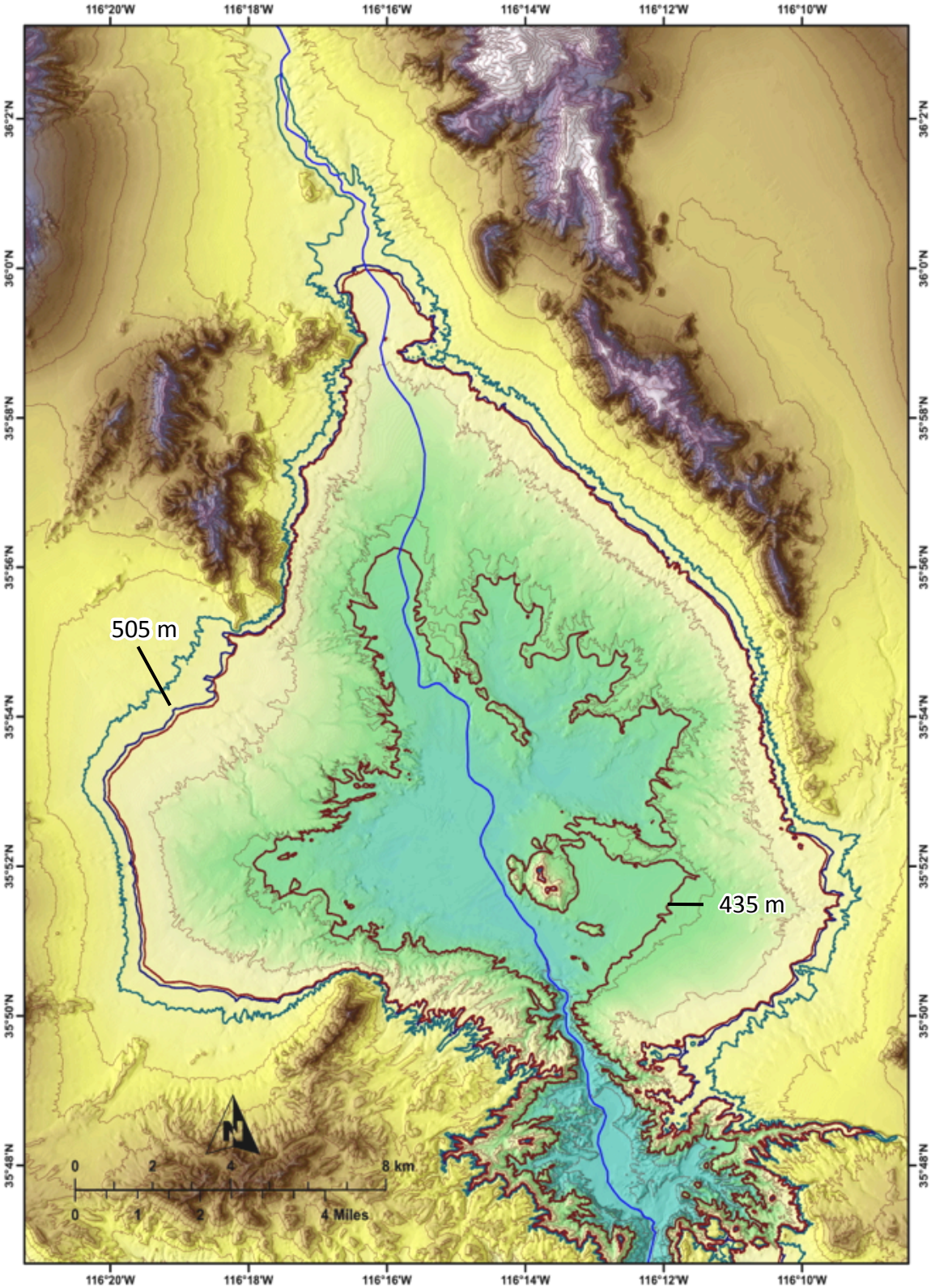


Image by John Niles

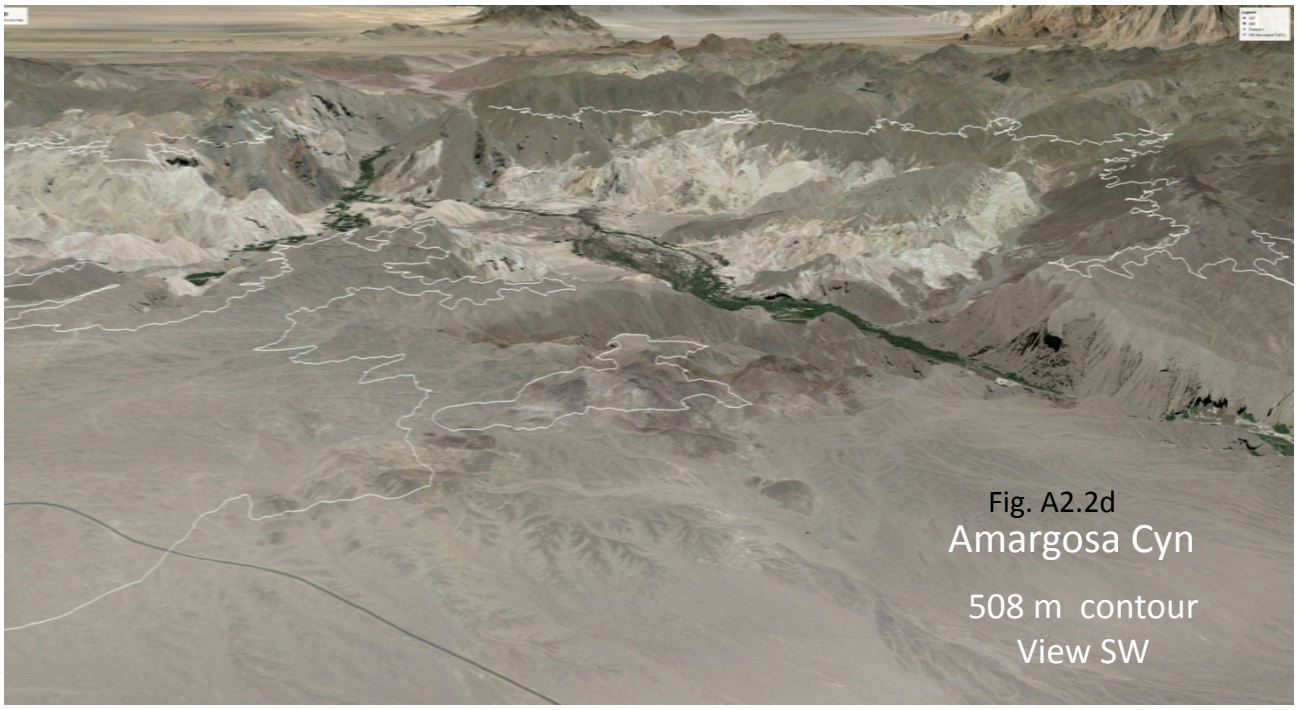


Fig. A2.2d
Amargosa Cyn
508 m contour
View SW

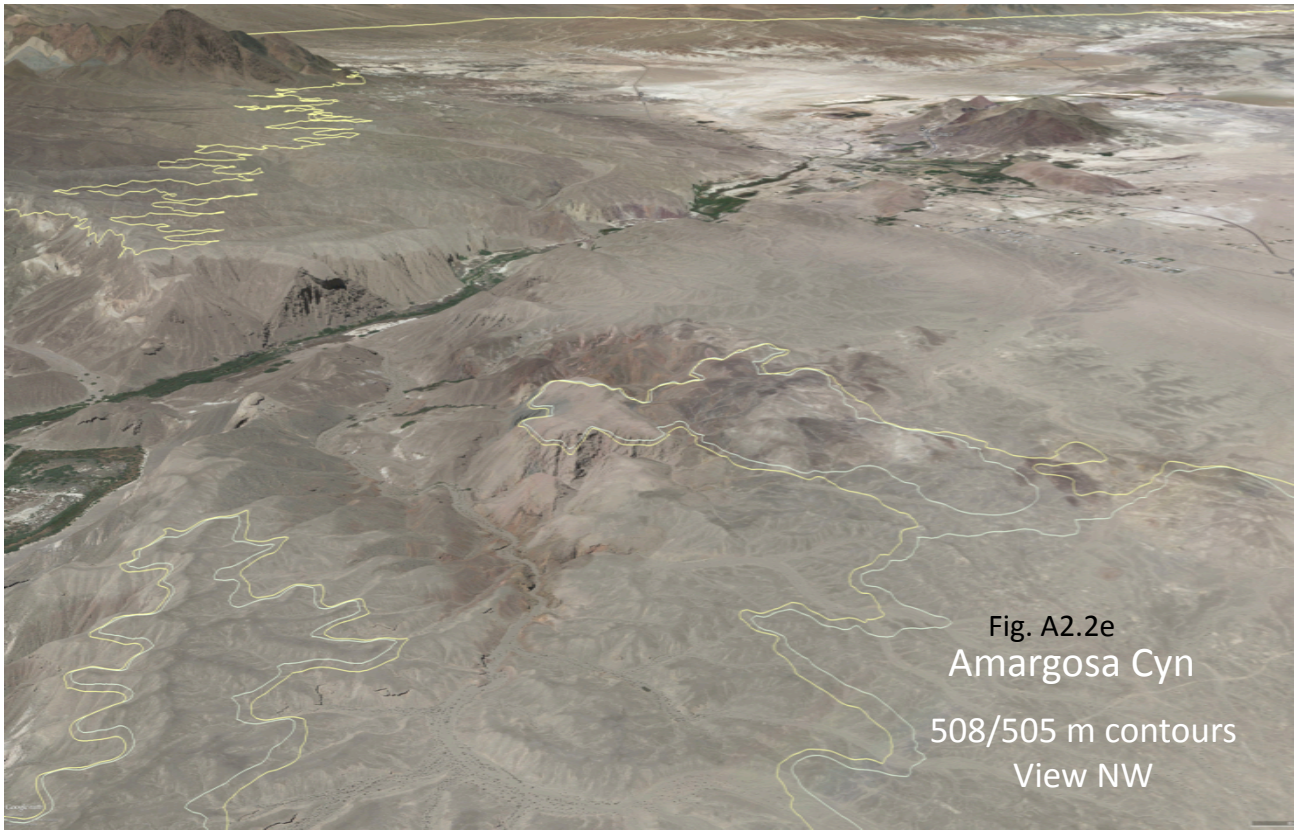
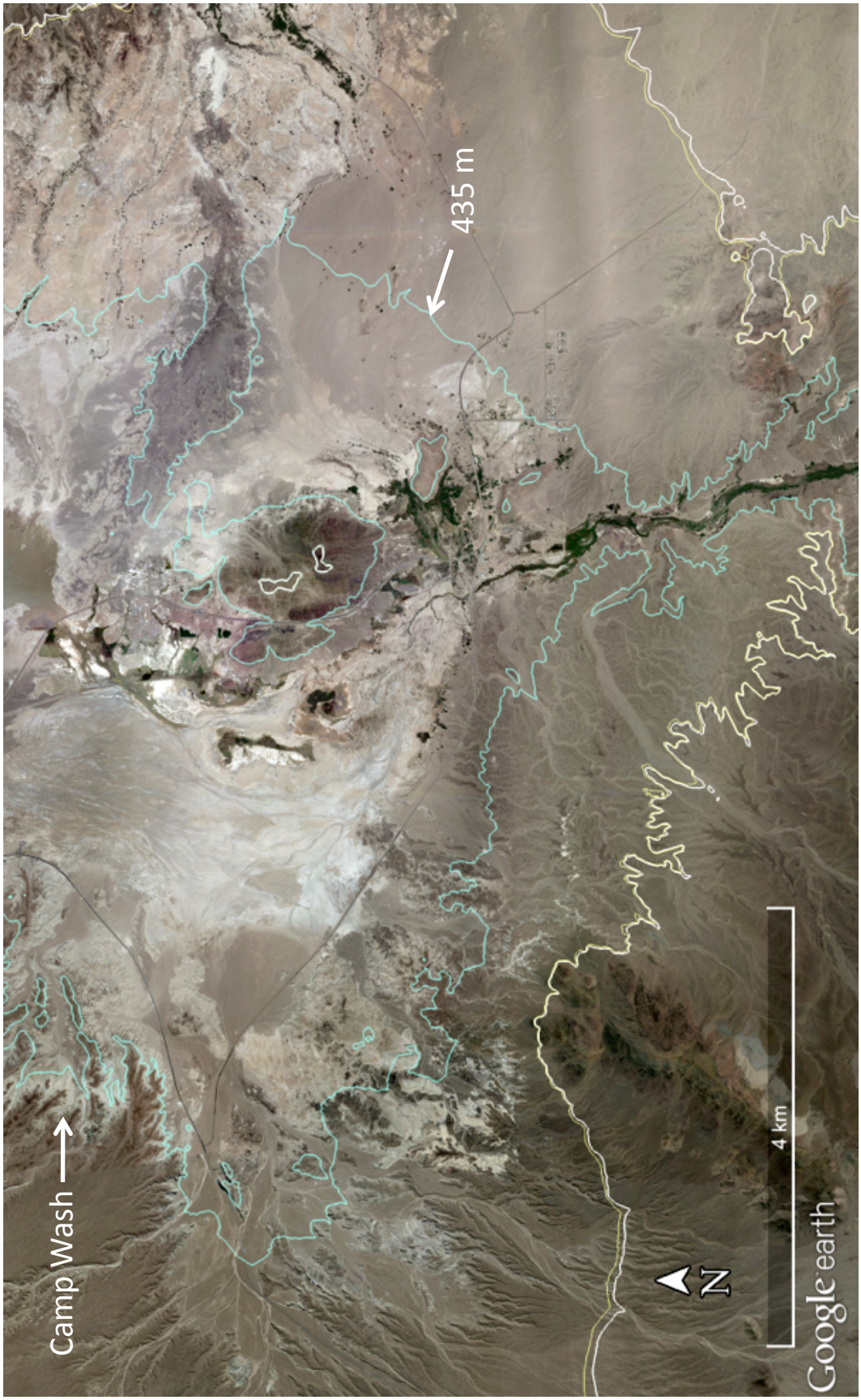


Fig. A2.2e
Amargosa Cyn
508/505 m contours
View NW


Camp Wash
435 m
N
4 km
Google earth

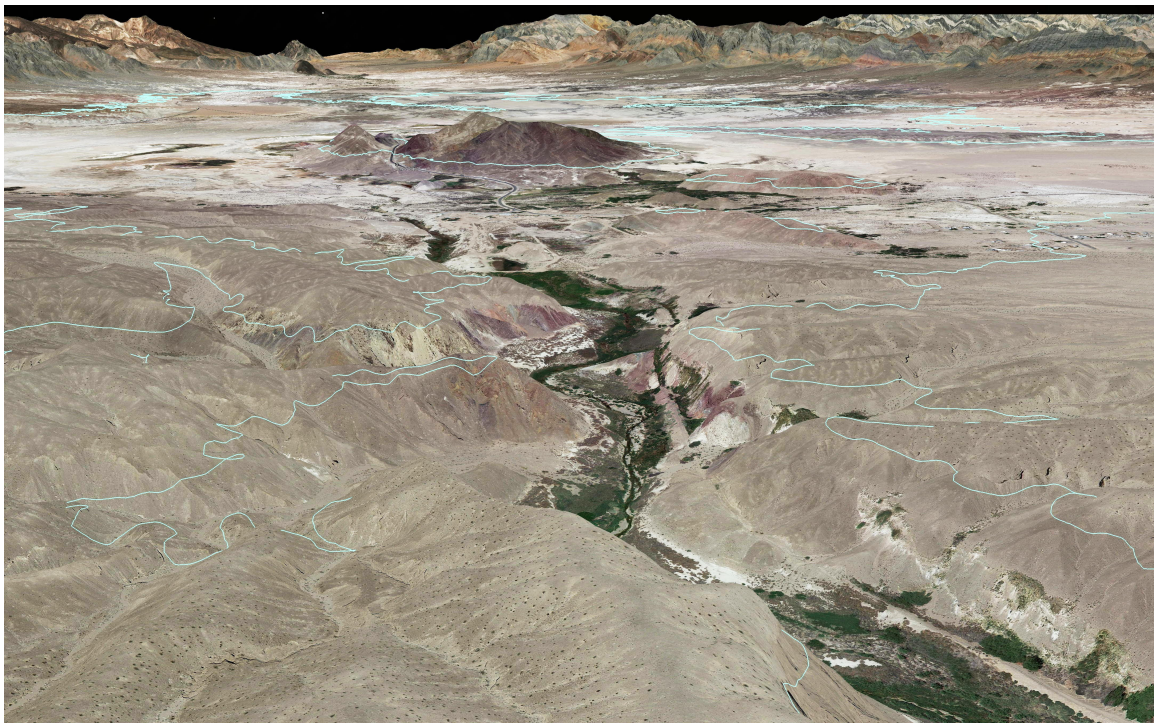


Fig. A2.2e Paleo-sill in the upper Amargosa Canyon. View is to the north.

Appendix A2.3 More Strat Columns (right out of Larsen, 2008)

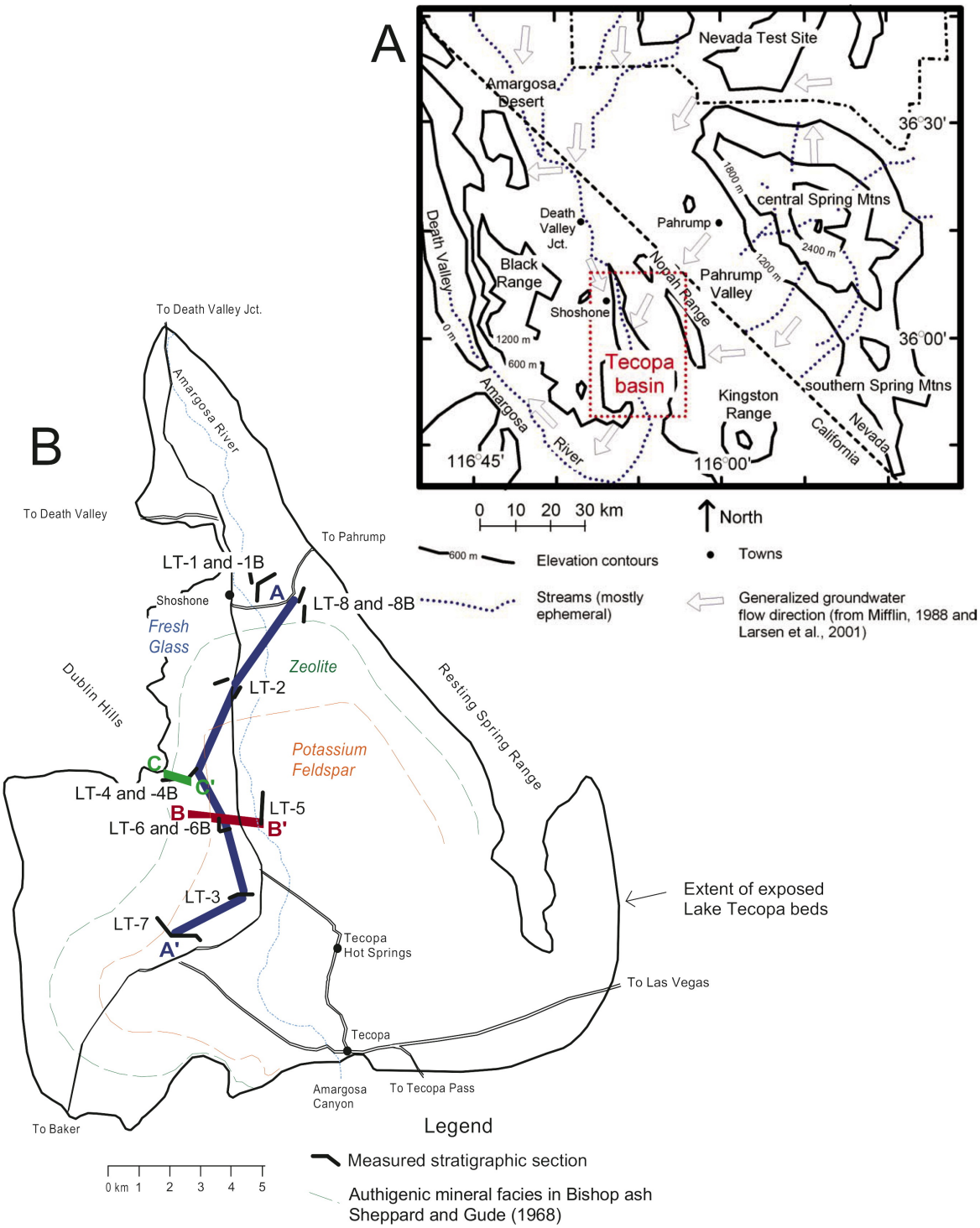
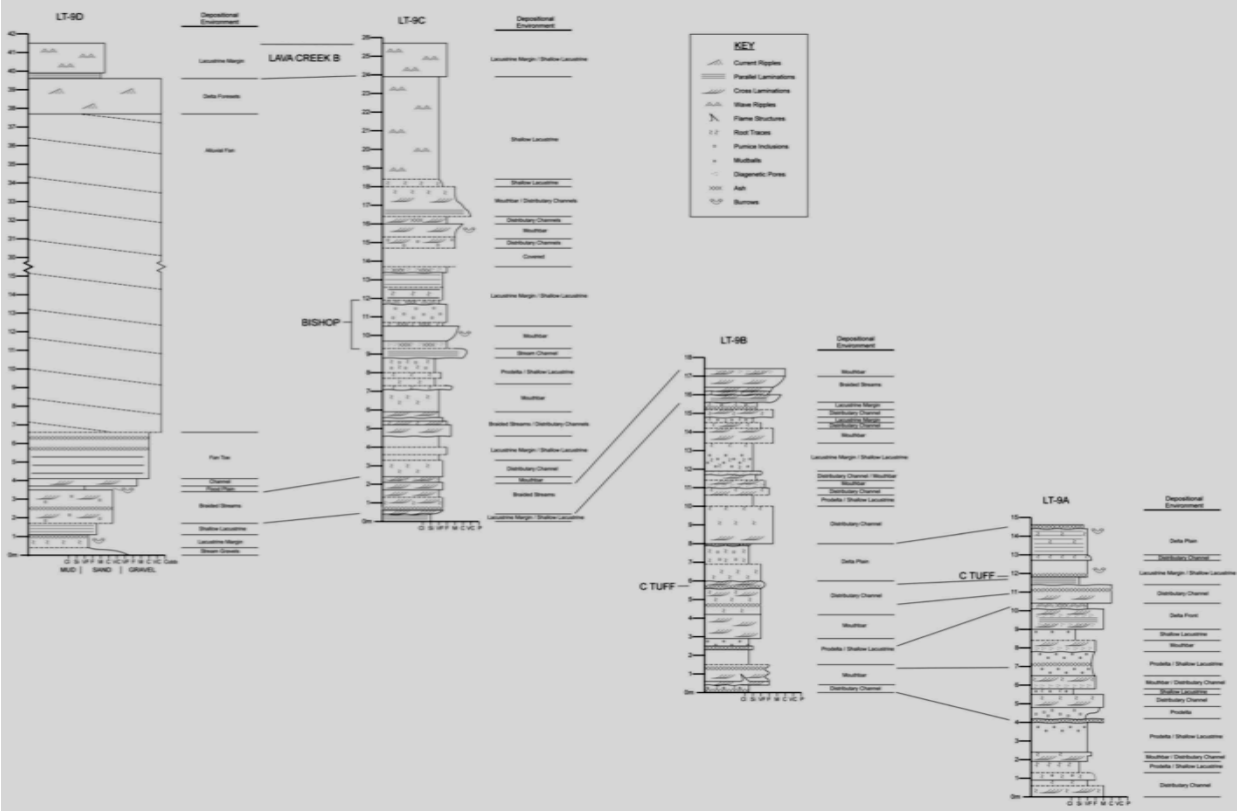


Figure 1. (A) Map of region surrounding the Tecopa basin, illustrating mountains, streams, and general groundwater flow directions. (B) Map of the Tecopa basin illustrating the locations of measured stratigraphic sections and correlated sections (line A-A' in Plates 1 and 2, line B-B' in Figs. 3 and 9, and line C-C' in Fig. 10). Note that line A-A' in Plate 2 includes section LT-1 rather than LT-8. Also shown are the authigenic mineral facies identified in the B tuff by Sheppard and Gude (1968).



Strat columns 9a-9d

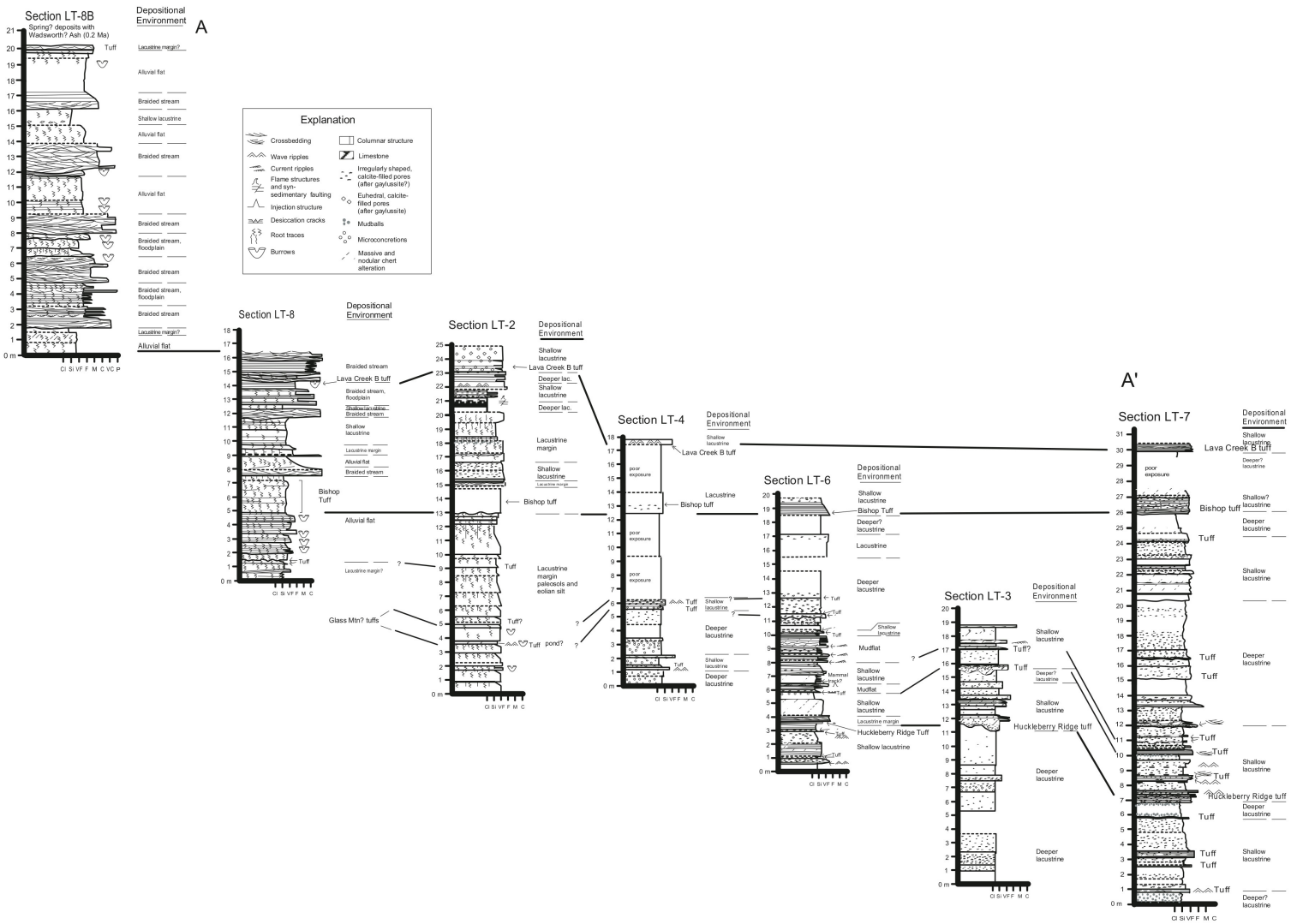


Plate 1. Correlated sections along line A-A' in Figure 1. Depositional environments are discussed in the text. Cl—clay; Si—silt; VF—very fine-grained sand; F—fine-grained sand; M—medium-grained sand; C—coarse-grained sand; VC—very coarse-grained sand; P—pebbles. If you are viewing the PDF of this paper or reading it offline, please visit http://dx.doi.org/10.1130/GES00152.SP1 (Plate 1) or the full-text article on www.gsajournals.org to view Plate 1.

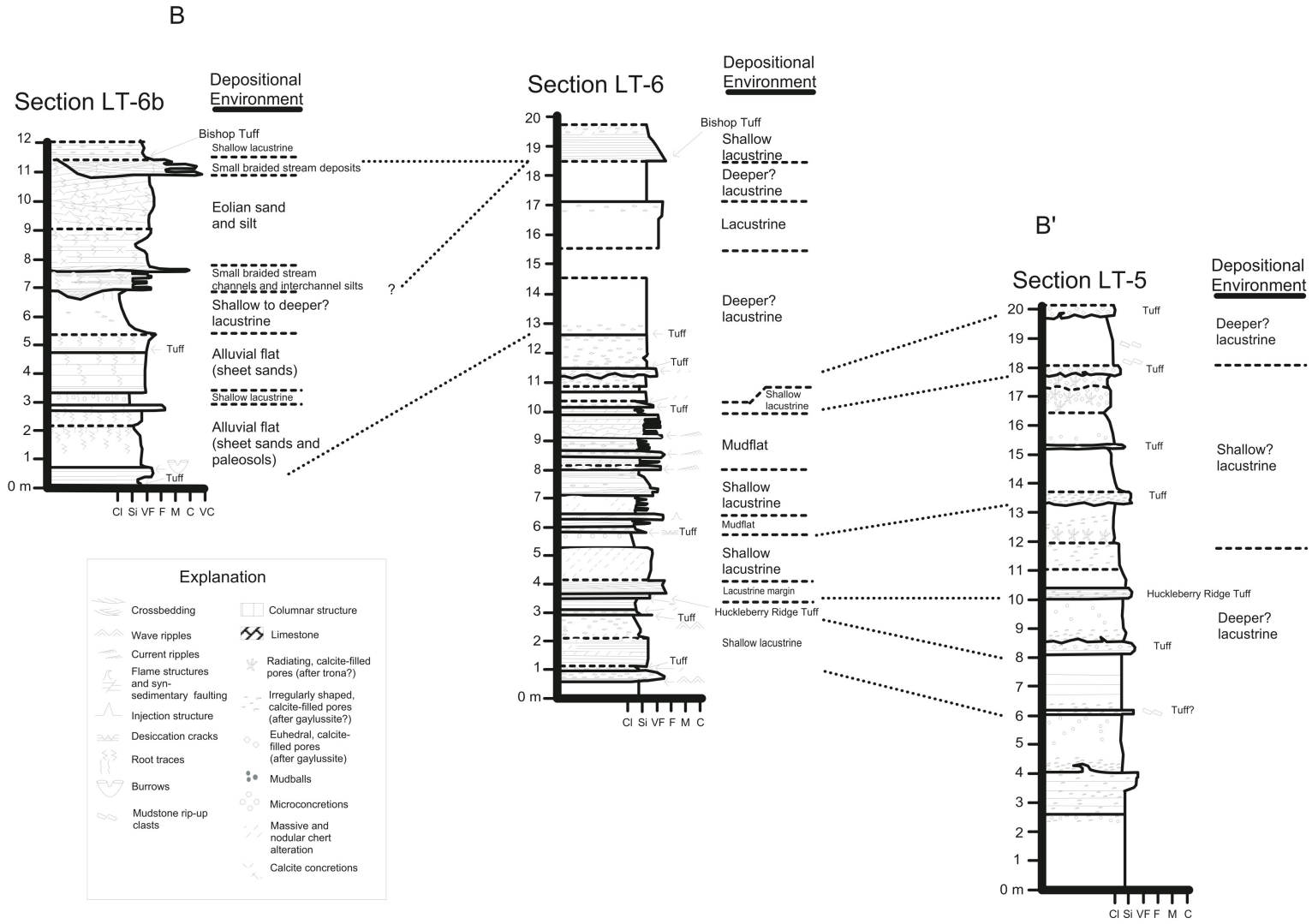


Figure 3. Correlated sections along line B-B' in Figure 1. Depositional environments are discussed in the text. Cl—clay; Si—silt; VF—very fine-grained sand; F—fine-grained sand; M—medium-grained sand; C—coarse-grained sand; VC—very coarse-grained sand.

Appendix 2.3.4 Chicago Valley Delta (Morrison, 2004)

EVIDENCE FOR A SHORT-LIVED LAKE RISE ~148 KA, LIKELY INDUCED BY FAULT-DAMMING OF THE AMARGOSA RIVER SOON AFTER THE DEMISE AND DRAINING OF LAKE TECOPA (CA)

MORRISON, Roger Barron, consultant, 7500 N. Calle Sin Envidia #8104, Tucson, AZ 85718, rbmorrison@earthlink.net

East of Tecopa Hot Springs 1 to 2.5 km, atop low mesa-like surfaces, are patches of post-Lake Tecopa gravel and sand that appear to be deltaic lake sediments from Chicago Valley drainage. They attain 445m altitude and >3 m thick locally, but thin rapidly to imperceptible. They likely record a short-lived lake rise that interrupted erosional history soon after Lake Tecopa's ~ 186 ka overflow. Calcareous cement (algal?) in deltaic gravel gave a U-Th age of 148±8 ka. These sediments overlie 1 to several m of cobbly alluvium from Chicago Valley, deposited after overflow of Lake Tecopa, after severe erosion, and late in development of the #1 post-Lake Tecopa alluvial surface. It is unlikely that climatic change caused this sudden lake rise; more likely it was fault-damming of the Amargosa River. I postulate that the damming was by a fault that crosses Armargosa River gorge 6 km (beeline) south of Tecopa general store and 1.2 km above the Willow (China Ranch) Wash junction. The fault strikes NNE and marks an abrupt narrowing of the river gorge, likely because its uplifted west side created a temporary sill. In the wider valley upstream, copious sidewash alluvium remains trapped and river terrace-gravel remnants are preserved. Possible confirmation of this interruption of through drainage from Tecopa Valley into Death Valley is from a core from Badwater: a short-lived desiccation of Lake Manly to a hypersalinity salt pan occurred ~148 ka.

2004 Denver Annual Meeting (November 7–10, 2004)
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Session No. 214--Booth# 40
Quaternary Geology (Posters) II
Colorado Convention Center: Exhibit Hall
8:00 AM-12:00 PM, Wednesday, 10 November 2004

Geological Society of America *Abstracts with Programs*, Vol. 36, No. 5, p. 497

Appendix 3A.1

LAKE MANLY AND PUPFISH DISPERSAL

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Two key conundrums of Lake Manly are its age and pupfish dispersal. Was Death Valley filled with water 30-10 ka (Last Glacial Maximum or Marine Isotope Stage 2; LGM or MIS 2) or at its maximum 180-120 ka? Blackwelder (1933, 1954) proposed that Lake Manly peaked and pupfish dispersed at the LGM. Better age control on both Lake Manly (e.g., Owen et al., 2011) and alluvial-fan deposits(e.g., Machette et al., 2008); the sediment core at Badwater (Forester et al., 2005; Lowenstein et al., 1999); and mitochondrial DNA work (Echelle, 2008; Smith et al., 2002) have clarified aspects of Lake Manly and pupfish dispersal. Although by no means unequivocal, the preponderance of data indicates that the deep lake that formed the prominent shoreline features (e.g., Badwater, Artists Drive, Beatty Road) were formed by a ~180-120 ka lake during MIS 6 (Owen et al., 2011). The Lake Manly at the LGM (MIS 2) was more likely a shallow lake that did not reach sea level (Knott et al., 2008; Machette et al., 2008). Of the many Lake Manly localities identified so far, none are unequivocally dated to coincide with the MIS 2 lake. A non-LGM age for Lake Manly is a hypothesis proposed by Blackwelder (1933).

The following is a summary from Knott et al. (2008). Mitochondrial DNA data indicate that pupfish that are found from Owens Valley southeast to the Colorado River dispersed from a common ancestor about 3 Ma (Smith et al., 2002). This may seem to make pupfish dispersal a Pliocene problem; however, a key component of dispersal is interconnection of pluvial lakes since 3 Ma. (2008). The hypothetical drainage limits of Death Valley today includes the Owens, Amargosa and Mojave Rivers. From 3-2 Ma, Death Valley was isolated and the Owens, Amargosa and Mojave Rivers terminated in Searles, Tecopa and Victorville, respectively. This hydrology persisted through ~500 ka. The Owens River flowed into Death Valley ~1 Ma (Jannik et al., 1991). By MIS 6, Death Valley was the terminus for the Owens (Jannik et al., 1991) and Amargosa (Morrison, 1999) Rivers. The incorporation of the Mojave River was in the late

Pleistocene. There is no hard evidence to support Blackwelder's hypothesis that Lake Manly overflowed into the Colorado River (Knott et al., 2008). Biological dispersal, like birds moving fish, is not addressed; other physical dispersal hypotheses are being tested.

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Special Paper 439
2008

A3.2 Knott et al. (2008)

Reconstructing late Pliocene to middle Pleistocene Death Valley lakes and river systems as a test of pupfish (Cyprinodontidae) dispersal hypotheses

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ABSTRACT

During glacial (pluvial) climatic periods, Death Valley is hypothesized to have episodically been the terminus for the Amargosa, Owens, and Mojave Rivers. Geological and biological studies have tended to support this hypothesis and a hydrological link that included the Colorado River, allowing dispersal of pupfish throughout southeastern California and western Nevada. Recent mitochondrial deoxyribonucleic acid (mtDNA) studies show a common pupfish (Cyprinodontidae) ancestry in this region with divergence beginning 3–2 Ma. We present tephrochronologic and paleomagnetic data in the context of testing the paleohydrologic connections with respect to the common collection point of the Amargosa, Owens, and Mojave Rivers in Death Valley

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during successive time periods: (1) the late Pliocene to early Pleistocene (3–2 Ma), (2) early to middle Pleistocene (1.2–0.5 Ma), and (3) middle to late Pleistocene (<0.7–0.03 Ma; paleolakes Manly and Mojave). Using the 3.35 Ma Zabriskie Wash tuff and 3.28 Ma Nomlaki Tuff Member of the Tuscan and Tehama Formations, which are prominent marker beds in the region, we conclude that at 3–2 Ma, a narrow lake occupied the ancient Furnace Creek Basin and that Death Valley was not hydrologically connected with the Amargosa or Mojave Rivers. A paucity of data for Panamint Valley does not allow us to evaluate an Owens River connection to Death Valley ca. 3–2 Ma. Studies by others have shown that Death Valley was not hydrologically linked to the Amargosa, Owens, or Mojave Rivers from 1.2 to 0.5 Ma. We found no evidence that Lake Manly flooded back up the Mojave River to pluvial Lake Mojave between 0.18 and 0.12 Ma, although surface water flowed from the Amargosa and Owens Rivers to Death Valley at this time. There is also no evidence for a connection of the Owens, Amargosa, or Mojave Rivers to the Colorado River in the last 3–2 m.y. Therefore, the hypothesis that pupfish dispersed or were isolated in basins throughout southeastern California and western Nevada by such a connection is not supported. Beyond the biologically predicted time frame, however, sparse and disputed data suggest that a fluvial system connected Panamint (Owens River), Death, and Amargosa Valleys, which could account for the dispersal and isolation before 3 Ma.

Keywords: Death Valley, Lake Manly, pupfish, tephochrology.

INTRODUCTION

Blackwelder (1933, 1954) hypothesized that during glacial (pluvial) climatic periods Death Valley was the final collection point for the Amargosa, Owens, and Mojave Rivers of southeastern California and western Nevada and that there may have been a hydrologic connection with the Colorado River (Fig. 1). Later, Hale (1985) hypothesized that the connection to the Colorado River was the result of overflow of Lake Manly,[1] which he believed occupied the Death Valley, Silver Lake, and Soda Lake Basins.

Blackwelder's (1933, 1954) geographic hypotheses were supported by Hubbs and Miller (1948), who found that pupfish (Cyprinodontidae) in the Owens, Amargosa, Mojave, and Colorado Rivers were morphologically similar. Working with Blackwelder's hypotheses and using pupfish morphology, Hubbs and Miller (1948) presumed that pupfish dispersal was facilitated by lake overflow along an integrated stream network. In their hypothesis, hydrologic isolation and the initiation of pupfish diversification began at the end of the Pleistocene, ~10,000 yr ago.

To test the overlapping hypotheses of Blackwelder (1933, 1954) and Hubbs and Miller (1948), Smith et al. (2002) examined Holocene and fossil mitochondrial deoxyribonucleic acid (mtDNA) of fauna in the Great Basin. By comparing the genetic change between living fauna and well-dated fossil fauna, a rate of genetic divergence among various species can be calculated. Focusing here on the pupfish, the molecular data support the divergence of pupfish in the Colorado, Owens, Amargosa, and Mojave Rivers from a common ancestor ~3–2 m.y. ago. Studies of mtDNA by Echelle et al. (2005) and Echelle (this volume) indicate isolation of Death Valley and the Lower Colorado River pupfish ca. 3–2 Ma, presumably by cessation of a hydrologic connection between these regions.

If the Owens, Amargosa, Mojave, and Colorado River systems were once integrated, then Death Valley is a key component because Death Valley is the only common connection point between the river systems (Fig. 1). In addition, the hypothesized back-flooding of Lake Manly into the Silver Lake and Soda Lake Basins ca. 180–128 ka (Hale, 1985; Meek, 1997; Hooke, 1999) provides an easier mechanism for pupfish dispersal to the Mojave River system than the relatively diminutive pupfish swimming upstream from lake to lake. Carrying the back-flooding argument further, if Lake Manly rose sufficiently to spill over into the Colorado River (Hale, 1985), then all of the southeastern California–western Nevada river systems would be integrated, which would allow mtDNA transfer in the most efficient manner.

Although the existence of Lake Manly deposits has been recognized for some time (Bailey, 1902), outcrops are discontinuous, and age control is sparse (Machette et al., 2001). As a result, the geologic debate of Blackwelder's original hypothesis regarding the integration of Lake Manly with the Owens, Amargosa, Mojave, and Colorado Rivers continues (Brown and Rosen, 1995; Enzel et al., 2002, 2003; Hale, 1985; Hooke, 1999, 2002; Meek, 1997). The 180–128 ka age of Lake Manly highstand deposits in Death Valley (Ku et al., 1998; Lowenstein et al., 1999) combined with the molecular data of Smith et al. (2002) and Echelle et al. (2005, this volume) increase the time frame for considering pupfish isolation, Lake Manly's existence, and hydro-

[1]In this paper we use Lake Manly in reference to Pleistocene-to-present Death Valley lakes when Death Valley had its approximate present configuration. Hence, we attribute Pliocene lake deposits in the Furnace Creek Basin and other areas to Death Valley lakes.

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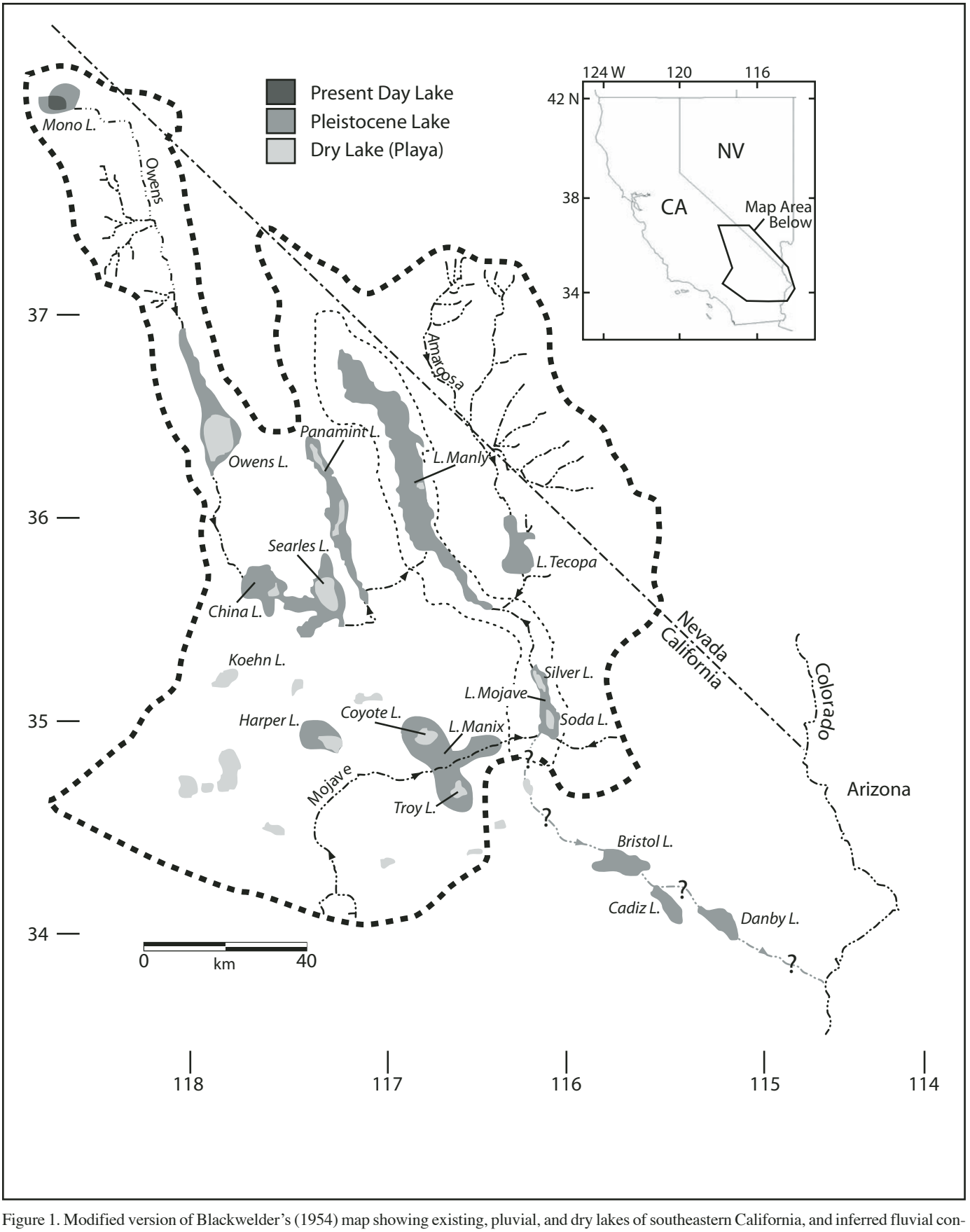

Figure 1. Modified version of Blackwelder's (1954) map showing existing, pluvial, and dry lakes of southeastern California, and inferred fluvial connections (dot-dash lines) during the late Pleistocene. The thick dashed line shows the approximate drainage basin limits for Death Valley (Pleistocene Lake Manly). The hypothesized connection to the Colorado River to the southeast is queried as Blackwelder showed. The thin dotted line shows the approximate size of a lake incorporating Death Valley and the Soda and Silver Lake Basins and overflowing toward the Colorado River.

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logic integration of southeastern California and western Nevada from the 10,000 yr suggested by Blackwelder to 10^5–10^6 yr. As a result, several key questions about Death Valley paleohydrology remain unanswered and should be considered from the perspective of both the biological and geological data and the expanded time frame:

(1) If the molecular data are correct (Smith et al., 2002; Echelle et al., 2005), what evidence is there in Death Valley for lakes and hydrologic connections to other basins and river systems during the Pliocene (4–2 Ma)?

(2) If Searles Lake, along the Owens River system, was overflowing ca. 1.2–0.5 Ma, possibly into Death Valley (Jannik et al., 1991), what evidence exists for Death Valley lakes during this same pluvial period?

(3) Did Lake Manly rise sufficiently to incorporate Lake Mojave ca. 180–128 ka (Hooke, 1999), providing a dispersal waterway and subsequent isolation?

In this paper, we present our initial findings and data regarding the age and extent of Pliocene lakes in northern Death Valley. The tephrochronology and lithofacies data indicate that a lake in the Furnace Creek Basin of northern Death Valley reached its maximum extent ca. 3.35 Ma. We have found no geologic evidence to suggest a surface hydrologic connection between Death Valley and the adjoining basins during the Pliocene. We also present tephrochronologic, paleomagnetic, and lithofacies evidence for early to middle Pleistocene Lake Manly. We use new tephrochronologic data from Soda Lake Basin core 1 to show that Lake Manly did not back-flood into the Silver Lake and Soda Lake Basins. Finally, these data and previous studies allow us to reconstruct Death Valley paleohydrology and to speculate on geologic solutions to pupfish dispersal.

METHODS

The results and interpretations presented herein are the result of several independent, but coordinated, studies of the Death Valley area. Geologic mapping was completed using maps, satellite imagery, and aerial photographs that varied in scale from 1:12,000 to 1:40,000. Tephrochronologic studies were completed using the methods described in Sarna-Wojcicki et al. (2005), and ash-bed correlations were made using the similarity coefficient of Borchardt et al. (1972), relative stratigraphic position to other tephra layers, and paleomagnetic data.

Paleomagnetic samples were collected from lacustrine sediments stratigraphically below the Bishop ash bed in a section known informally as Ash Canyon at Mormon Point (Fig. 2). Beginning at the top of the section, the sample sites are spaced vertically over a stratigraphic thickness of no more than ~12 m. The samples (cubes, 2.5 cm on a side) were measured and thermally demagnetized at the Paleomagnetism Laboratory of the University of California, Santa Cruz, using a 2-G cryogenic magnetometer.

For the Soda Lake Basin core, sediment samples, where available, were collected at depths where Muessig et al. (1957) recorded volcanic ash. Only core material from the even-numbered boxes of the Muessig et al. (1957) study is stored at the Desert Study Center at Zzyzx, California. The core storage was less than optimal, and there is evidence that other scientists have taken samples from the cores over the years. As a result, the cores are in poor condition.

RESULTS

Pliocene Tuffs and Deposits

The late Miocene–Pliocene Furnace Creek Basin consists of interbedded conglomerates, sandstones, mudstones, and basalts that are intermittently exposed from the northern Black Mountains to near the northeast corner of the Cottonwood Mountains on the north (Fig. 2; Hunt and Mabey, 1966; McAllister, 1970; Wright et al., 1999; Wright and Troxel, 1993). These sediments and volcanic flows were deposited in a down-dropping basin that developed between the Furnace Creek, Northern Death Valley, and Grand View fault zones (Wright et al., 1999). Reported Pliocene-aged formations of the Furnace Creek Basin include the Furnace Creek, Nova, and Funeral Formations. All of these formations contain basalt flows and rhyolite air-fall tuffs (Hunt and Mabey, 1966) and thus are datable by modern methods. The Nova and Funeral Formations are dominantly composed of breccias and conglomerates, whereas the Furnace Creek Formation includes fine-grained facies (Hunt and Mabey, 1966).

Hunt and Mabey (1966) and McAllister (1970) both recognized that the fine-grained facies of the Furnace Creek Formation represented playa-lake and/or playa depositional environments. Blair and Raynolds (1999) interpreted laminar, fine-grained sediments as indicative of lacustrine deposition in the northern portions of the Furnace Creek Basin. The only time control previously available for the Furnace Creek Formation was a single 4 Ma whole-rock K/Ar date on an overlying basalt flow (McAllister, 1970).

Several tephrochronologic studies (Knott et al., 1999; Sarna-Wojcicki et al., 2001) have provided age control for key Pliocene tephra beds and tuffs in the Death Valley region (Fig. 3). We expand this previous work to other localities in Death Valley in an effort to define the paleohydrologic conditions 3–2 m.y. ago (Figs. 2 and 3); sites are discussed next by area within Death Valley. The locations of tephra layers and tuffs analyzed for this study are also provided in Table 1.

Lake Rogers

The Lake Rogers area (Fig. 2) represents the northern portion of the ancient Furnace Creek Basin. The type locality of the informal Mesquite Spring tuff (Tables 1 and 2, JRK-DV-112) of Snow and White (1990) is located along the eastern Cottonwood Mountains piedmont, southwest of the Lake Rogers Basin (Fig. 2). Here, the ~10-m-thick, pumiceous, Mesquite Spring tuff has a $^{40}Ar/^{39}Ar$ age of 3.28 ± 0.07 Ma and is interbedded with alluvial-fan breccias and conglomerates (Snow and Lux, 1999). The pumiceous texture and age make this tephra bed a

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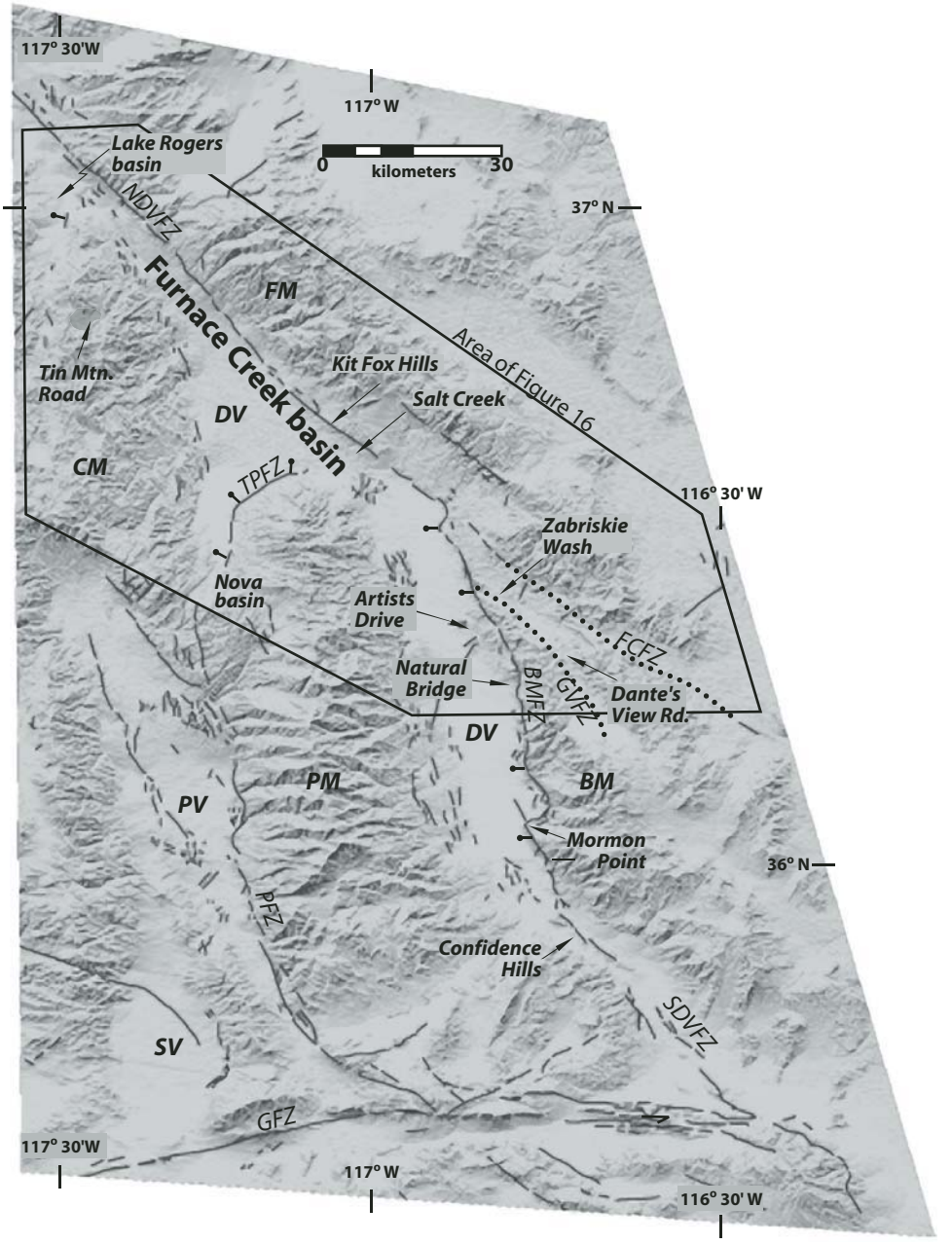

Figure 2. Shaded relief map of Death Valley area showing Quaternary fault zones (FZ—fault zone; BMFZ—Black Mountains; GFZ—Garlock; FCFZ—Furnace Creek; PFZ—Panamint; SDVFZ—Southern Death Valley; NDVFZ—Northern Death Valley; TPFZ—Towne Pass; GVFZ—Grand View), and the ancient Furnace Creek Basin, sample localities discussed in text, and main geographic features (BM—Black Mountains; CM—Cottonwood Mountains; DV—Death Valley; FM—Funeral Mountains; PM—Panamint Mountains; PV—Panamint Valley; SV—Searles Valley). Polygon shows approximate location of Figure 16.

likely correlative with the 3.29 Ma pumiceous tuff found at Zabriskie Wash.

South of the Mesquite Spring type locality, the older than 3.58 Ma, informally named, tuff of Artists Drive (Table 2, JRK-DV-113) is interbedded with alluvial-fan breccias and conglomerates. At Lake Rogers (Fig. 2), a 10-m-thick, air-fall tephra (REK-DV-3) interbedded with fine-grained lacustrine sediment is correlative with the 3.35 Ma Zabriskie Wash tuff (Table 2; Fig. 4). Here, the lacustrine deposits overlying the Zabriskie Wash tuff grade rapidly upward into massive conglomerates and breccias deposited in an alluvial-fan environment (Klinger and Sarna-Wojcicki, 2001). The tuff of Zabriskie Wash is also found west of Lake Rogers along Tin Mountain Road (Table 2, JRK-DV-71). Here, the Zabriskie Wash tuff is interbedded with alluvial-fan gravels and has a $^{40}Ar/^{39}Ar$ age of 3.35 ± 0.13 Ma (Snow and White, 1990).

Kit Fox Hills

Fine sand, silt, and clay exposed in the Kit Fox Hills were deposited near the center of the ancient Furnace Creek Basin (Fig. 2). These light-colored lacustrine beds (Fig. 3) are fine-grained and laminated and contain no evidence of nearshore wave effects such as ripple marks or cross-bedding. The lacustrine sequence is overlain by alluvial-fan conglomerates.

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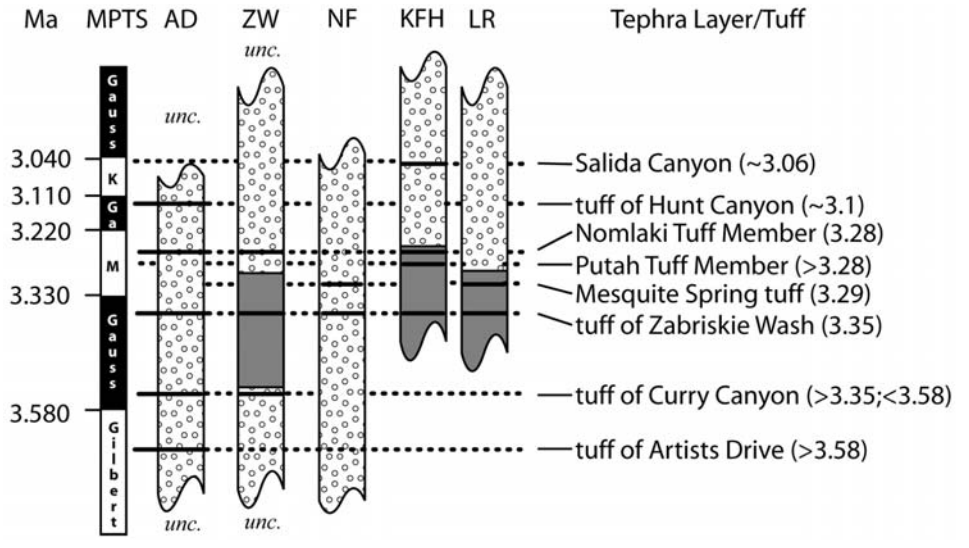

Figure 3. Diagram showing common tephra beds and tuffs in late Pliocene deposits of Death Valley (unc.—unconformity). Chemical analyses of the tephra beds and tuffs are given in Table 2. Ages of tephra beds and tuffs are in parentheses. Magnetic polarity time scale (MPTS) and chrons are shown at left (Ga—Gauss; M—Mammoth; K—Kaena). Death Valley areas are Artists Drive (AD), Zabriskie Wash (ZW), Nova Formation (NF), Kit Fox Hills (KFH), and Lake Rogers (LR). Solid shading of stratigraphic sections indicates lacustrine deposits; dot pattern indicates alluvial-fan deposits.

In the northern Kit Fox Hills, a 5.5-m-thick tephra bed is tentatively correlated with the 3.35 Ma tuff of Zabriskie Wash (Table 2, VME-KFH-07). Here, the Zabriskie Wash tuff marks the transition from fine-grained lacustrine deposits below to overlying coarse alluvial-fan deposits.

At the base of a section in the southern Kit Fox Hills, the 3-m-thick, 3.35 Ma tuff of Zabriskie Wash (Table 2, JRK-DV-204 and JRK-DV-219) is interbedded with lacustrine deposits (Fig. 5). The 3.28 Ma Nomlaki Tuff Member of the Tuscan and Tehama Formations (Table 2, JRK-DV-214) is found higher in the section, where the lacustrine deposits grade upward to alluvial-fan conglomerates. Within the overlying conglomerates, there is the informal tuff of Salida Canyon (Table 2, JRK-DV-226). An age of ca. 3.06 Ma for the tuff of Salida Canyon is estimated from sedimentation rates in the Burmester core from Lake Bonneville, Utah, another locality where this tuff also overlies the Nomlaki Tuff Member (Williams, 1994).

At Salt Creek, which abuts the southwestern Kit Fox Hills (Fig. 2), fine-grained lacustrine silt and sand deposits are exposed along the uplifting Salt Creek anticline. The ca. 3.29 Ma Putah Tuff Member of the Tuscan and Tehama Formations (Table 2, JRK-DV-209) is interbedded with these deposits. The Putah Tuff Member erupted from the Sonoma volcanic field just north of present-day San Francisco and is found underlying the Nomlaki Tuff Member in drill cores in northern California basins (Sarna-Wojcicki, 1976).

Nova Basin

The Nova Basin (Fig. 2) is located at the same latitude in the ancient Furnace Creek Basin as Zabriskie Wash, but to the west. The Nova Basin consists of alluvial-fan deposits with interbedded basalts and tuffs (Hunt and Mabey, 1966). Knott et al. (1999) inferred that the tuffs at Artists Drive (discussed later) were correlative with tuffs that Snow and White (1990) considered to be the Mesquite Spring tuff.

Tephrochronologic data confirm that tephra layers in the Nova Basin are equivalent to tephra layers found at Artists Drive, Zabriskie Wash, and Lake Rogers (Table 2). Thus, we consider the Nova Basin to be an alluvial-fan–filled embayment of the Pliocene (ancient) Furnace Creek Basin. The 3.35 ± 0.07 Ma tephra layer (Table 2, JRK-DV-80) in the Nova Basin (Snow and Lux, 1999) is correlative with the tuff of Zabriskie Wash at Artists Drive, Zabriskie Wash, and Lake Rogers. In addition, the older than 3.58 Ma tuff of Artists Drive (see Artists Drive section) is tentatively identified in the Nova Basin (Table 2, JT wall ash).

Zabriskie Wash

Deposits of the southern part of the ancient Furnace Creek Basin are well exposed at Zabriskie Wash (Fig. 2; McAllister, 1970). Here, alluvial-fan deposits surround fine-grained playa and lacustrine deposits (sandstones, mudstones, and marls). Near the base and top of the playa-lacustrine sequence are the informal Curry Canyon tuff (>3.35, <3.58 Ma) and Nomlaki Tuff Member (3.28 Ma), respectively. The 3.35 Ma Zabriskie Wash and 3.28 Ma Mesquite Spring tuffs (Fig. 6) occur between these two tephra beds. This suite of ashes crops out in basin-margin deposits for several kilometers to the northwest.

Southeast of Zabriskie Wash, along the Dante's View Road, the Zabriskie Wash tuff (Table 2, BTD-DV-01) is interbedded with locally derived alluvial-fan deposits. Here, the Zabriskie Wash tuff is ~1 m thick and extends for several meters along a steep stream bank. McAllister (1970) mapped these deposits as part of the Funeral Formation near the southeastern boundary of the ancient Furnace Creek Basin.

Artists Drive

At Artists Drive, southwest of Zabriskie Wash (Fig. 2), Knott et al. (1999, 2005) used glass-shard composition to identify two "Mesquite Spring–like" tuffs (Table 2, JRK-DV-41; JRK-DV-40). Knott et al. (1999, 2005) initially used lower and upper

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TABLE 1. SAMPLE LOCATIONS

Sample no.	UTM, Zone 11, NAD27	7.5′ Quadrangle	Description of sampled unit
JRK-DV-54	522943 m E 3989662 m N	Mormon Point	17-cm-thick, white to gray, weathering to tan tephra
JRK-DV-55	522842 m E 3989940 m N	Gold Valley	33-cm-thick, white weathering to tan tephra; upper part is reworked
JRK-DV-56	522842 m E 3989724 m N	Gold Valley	25–108-cm-thick, white weathering to tan, finely laminated ash
JRK-DV-57	523692 m E 3990096 m N	Gold Valley	19-cm-thick, white to gray, laminated to finely bedded ash
JRK-DV-70	522091 m E 3990153 m N	Mormon Point	140-cm-thick tuffaceous sand, poorly to moderately indurated
JRK-DV-77	522940 m E 3990463 m N	Mormon Point	15-cm-thick gray ash overlying JRK-DV-78
JRK-DV-78	522940 m E 3990463 m N	Mormon Point	15-cm-thick white ash
JRK-DV-58	523541 m E 3990373 m N	Gold Valley	61–80-cm-thick white biotite-sanidine lapilli to ash tuff
JRK-DV-42	521354 m E 4015385 m N	Devils Golf Course	White weathering to brown, biotite tuff
JRK-DV-43	521304 m E 4015385 m N	Devils Golf Course	40-cm-thick white tuff
JRK-DV-62	521303 m E 4015724 m N	Devils Golf Course	10-cm-thick well-indurated tuff
JRK-DV-93	521866 m E 3989813 m N	Mormon Point	2–3-cm-thick white biotite-phenocryst basal layer of 98-cm-thick tuffaceous sand
JRK-DV-99	521867 m E 3989228 m N	Mormon Point	180-cm-thick white biotite-phenocryst tephra
JRK-SL-407	583970 m E 3890713 m N*	Soda Lake North	Light brown to tan fine sand from Soda Lake core
JRK-DV-36	522269 m E 3989075 m N	Mormon Point	10-cm-thick white tephra
JRK-DV-8	522268 m E 3989199 m N	Mormon Point	7-cm-thick white tephra with oscillatory ripples
JRK-DV-90	521766 m E 3989783 m N	Mormon Point	15-cm-thick white tephra
JRK-DV-91	521766 m E 3989783 m N	Mormon Point	25-cm-thick white tephra
JRK-DV-10	521766 m E 3989937 m N	Mormon Point	White to tan, partially cemented ash
JRK-SL-599	583970 m E 3890713 m N*	Soda Lake North	Brown to tan fine sand
JRK-DV-226	496798 m E 4057478 m N	Stovepipe Wells NE	35-cm-thick, fine-grained white ash
JRK-DV-60	516049 m E 4024062 m N	Devils Golf Course	25-cm-thick, pinkish-white (5YR 8/1) tuff
JRK-DV-214	495358 m E 4057726 m N	Stovepipe Wells NE	20-cm-thick white biotite phenocryst ash
JRK-DV-209	499081 m E 4050268 m N	Grotto Canyon	Gray fine-grained ash
JRK-DV-112	463892 m E 4088441 m N	Tin Mountain	10-m-thick white pumice lapilli tuff
JRK-DV-71	461749 m E 4069457 m N	White Top Mtn	500-cm-thick pumiceous lapilli tuff
JRK-DV-80	486175 m E 4027694 m N	Emigrant Canyon	10-m-thick white biotite-phenocryst tuff
JRK-DV-204	495910 m E 4056936 m N	Stovepipe Wells NE	10-m-thick white massive tuff
JRK-DV-219	495010 m E 4057387 m N	Stovepipe Wells NE	190-cm-thick, finely bedded, pumiceous tephra
JRK-DV-41	516024 m E 4024093 m N	Devils Golf Course	37-cm-thick white tephra
REK-DV-3	460333 m E 4100617 m N	Ubehebe Crater	>1-m-thick pumiceous tuff
BTD-DV-01	529040 m E 4017203 m N	Ryan (CA)	32–36-cm-thick white biotite-phenocryst tephra
VME-DV-07	490621 m E 4061951 m N	Stovepipe Wells NE	350-cm-thick white to tan tephra
JRK-DV-39	516024 m E 4024000 m N	Devils Golf Course	60-cm-thick white to tan biotite-phenocryst tephra
JRK-DV-104	515924 m E 4024000 m N	Devils Golf Course	~50-cm-thick white ash
JRK-DV-113	467216 m E 4085673 m N	Tin Mountain	30-cm-thick light-gray biotite-phenocryst tuff
JT Wall Ash	482545 m E 4043815 m N	Stovepipe Wells Village	35-cm-thick white reworked tephra

*Estimated location of drill hole. Location of drill hole 1 from Muessig et al. (1957) is NE1/4 s. 1, T12N, R8E.

Mesquite Spring tuff to describe these tephra layers. As more data accumulated, the use of the lower and upper designations became impractical. The upper "Mesquite Spring–like" tephra layer has an estimated age of 3.1 Ma and is informally named herein the tuff of Hunt Canyon (informal location name) (Fig. 3). The lower "Mesquite Spring–like" tephra layer is correlated with the informally named tuff of Zabriskie Wash.

The "Nomlaki-like" tuff of Knott et al. (1999; Tables 1 and 2, JRK-DV-39) is stratigraphically below the 3.35 Ma tuff of Zabriskie Wash at Artists Drive and has reversed polarity. Knott et al. (2005) referred to this tephra layer as the lower Nomlaki tuff. In this paper, we informally name this tephra layer the tuff of Artists Drive. The first reversed polarity chron older than 3.35 Ma is the Gilbert (chron C2Ar; 4.180–3.580 Ma; Berggren et al., 1995), making the minimum age of the tuff of Artists Drive 3.580 Ma (Fig. 3). In Death Valley, the Artists Drive tuff crops out intermittently at Artists Drive (Table 2, JRK-DV-39 and JRK-DV-104) and is also found near the Cottonwood Mountains piedmont and in the Nova Basin (Fig. 2). The tuff of Artists Drive is also found to the southeast along the Colorado River (House et al., this volume).

In addition, we correlate a reversed polarity tephra layer 4 m stratigraphically above the Zabriskie Wash tuff (Table 2, JRK-DV-60) at Artists Drive with the 3.28 Ma Nomlaki Tuff Member. The glass-shard composition of the Nomlaki Tuff Member is very similar to the tuff of Artists Drive. This relation suggests

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TABLE 2. RESULTS OF ELECTRON-MICROPROBE ANALYSIS OF VOLCANIC GLASS SHARDS FROM PLIOCENE TEPHRA LAYERS IN DEATH VALLEY, AND FOR SIMILAR OR CORRELATIVE TEPHRA LAYERS ELSEWHERE

Sample*	Locality†	SiO_2	Al_2O_3	Fe_2O_3	MgO	MnO	CaO	TiO_2	Na_2O	K_2O	Total (o)§	SC#
Tuff of Salida Canyon (ca. 3.06 Ma)**												
JRK-DV-226	KFH	72.36	14.68	2.70	0.25	0.09	0.87	0.31	5.29	3.44	91.93	1.0000
C-83–6A††		72.50	14.47	2.73	0.27	0.11	0.89	0.29	5.39	3.36	N.A.§§	0.9772
BUR-915		72.02	14.46	2.67	0.26	0.11	0.87	0.33	5.66	3.63	N.A.	0.9817
Nomlaki Tuff Member of the Tuscan and Tehama Formations (3.28 Ma)												
JRK-DV-60	AD	76.90	12.76	1.21	0.24	0.04	1.13	0.21	3.99	3.52	93.00	1.0000
JRK-DV-214	KFH	77.11	12.92	1.06	0.18	0.04	0.91	0.21	3.45	4.11	89.72	0.9332
758–324(2)		77.50	12.90	1.14	0.17	0.03	0.95	0.22	3.80	3.40	94.60	0.9438
OP84–193A		76.80	13.00	1.15	0.22	0.04	1.08	0.23	4.10	3.40	94.40	0.9599
Putah Tuff Member of the Tuscan and Tehama Formations (ca. 3.29 Ma)												
JRK-DV-209	KFH	75.59	13.19	1.78	0.08	0.03	0.60	0.16	3.72	4.84	90.60	1.0000
759–287D		76.30	12.70	1.76	0.09	0.02	0.61	0.18	4.10	4.30	92.90	0.9630
758–292		74.90	13.90	1.76	0.09	0.03	0.72	0.14	4.10	4.40	92.90	0.9274
PVR-2		75.61	13.21	1.74	0.09	0.03	0.61	0.18	4.03	4.51	N.A.	0.9696
MRM-1–01		75.90	13.02	1.66	0.07	0.03	0.58	0.17	3.72	4.83	N.A.	0.9647
Tuff of Mesquite Spring (3.29 Ma)												
JRK-DV-112	CM	77.80	12.80	0.73	0.03	0.04	0.43	0.06	3.66	4.46	94.78	1.0000
MNM-DV-8-FC	ZW	76.28	13.01	0.72	0.04	0.09	0.46	0.05	3.59	5.76	95.57	0.9437
Tuff of Zabriskie Wash (3.35 Ma)												
JRK-DV-71	CM	77.25	12.67	0.73	0.04	0.08	0.48	0.05	3.48	5.21	93.86	1.0000
JRK-DV-80	NB	76.82	12.72	0.75	0.04	0.08	0.48	0.06	2.84	6.21	93.26	0.9594
JRK-DV-204	KFH	76.95	13.06	0.73	0.04	0.07	0.47	0.07	3.21	5.40	95.64	0.9319
JRK-DV-219	KFH	76.78	13.16	0.74	0.04	0.08	0.48	0.07	3.17	5.49	94.27	0.9315
JRK-DV-41	AD	77.18	13.23	0.75	0.04	0.07	0.48	0.07	4.60	3.57	93.10	0.9289
REK-DV-3	LR	76.99	12.82	0.71	0.03	0.09	0.48	0.06	3.70	5.13	94.94	0.9582
BTD-DV-01	FC	77.15	12.92	0.71	0.04	0.08	0.47	0.06	3.22	5.35	92.39	0.9529
VME-DV-07	KFH	77.45	12.73	0.74	0.04	0.04	0.45	0.05	3.17	5.34	93.54	0.9833
Tuff of Artists Drive (>3.58 Ma)												
JRK-DV-39	AD	75.41	13.82	1.20	0.24	0.06	0.93	0.22	4.55	3.56	93.27	1.0000
JRK-DV-104	AD	76.85	13.40	1.12	0.22	0.06	0.88	0.20	3.76	3.53	93.89	0.9479
JRK-DV-113	LR	76.70	13.52	1.12	0.23	0.07	0.91	0.20	3.79	3.47	93.69	0.9565
JT Wall Ash	NB	76.19	13.66	1.15	0.22	0.08	0.92	0.21	4.33	3.25	N.A.	0.9761
FLV-119-WW		76.88	12.96	1.07	0.16	0.05	0.89	0.19	3.40	4.41	92.30	0.9262
Natural Glass Standards used as monitor												
RLS-132 ($n = 18$)		75.4	11.3	2.12##	0.06	0.16	0.11	0.19	4.9	4.4	98.6	
$\pm 1\sigma$		0.1	0.2	0.04	0.01	0.01	0.01	0.01	0.1	0.1		
$\%\sigma$		0.2	1.4	1.9	17	6.3	9	5.3	2.7	1.4		
RLS-132		75.7	11.4	2.12	0.05	0.15	0.12	0.21	5.3	4.5	99.6	

Note: Values are given in mean wt% oxide, recalculated to 100% fluid-free basis; C.E. Meyer (1977–1996), U.S. Geological Survey, Menlo Park, California, analyst.

*Samples are grouped according to best matches (inferred correlations).

†Abbreviations for localities are as follows: AD—Artists Drive; CM—Cottonwood Mountains: FC—Furnace Creek; LR—Lake Rogers; KFH—Kit Fox Hills; NB—Natural Bridge; ZW—Zabriskie Wash (all in Death Valley).

§Original oxide totals before calculation are given to indicate approximate degree of hydration of volcanic glass. Original values for each oxide can be obtained by multiplying Total(o) by weight percent shown.

#Similarity coefficient (SC) of Borchardt et al. (1972) used to quantitatively compare the tephra samples, where 1.0000 represents a perfect match. All similarity coefficients were calculated using concentrations of Si, Al, Fe, Ca, and Ti oxides.

**See text for source of estimated or isotopic ages from other localities in the western United States.

††Data from correlative samples. C-83–6A—tuff of Salida Canyon, Salida Canyon, Kettleman Hills, California; BUR-915—Ash from 915 m depth in Burmeister core, Lake Bonneville core, Utah (Williams, 1994); 758–324(2) and OP84–193A—Nomlaki Tuff Member (of Tehama and Tuscan Formations), Sacramento Valley, California; 759–287D and 758–292—Putah Tuff Member (of Tuscan and Tehama Formations), southwestern Sacramento Valley, California; PVR-2—Putah Tuff Member type locality near Putah Creek, Sonoma Valley, California; MRM-1–01—Putah Tuff Member, Sonoma volcanic field, Sonoma Valley, California; FLV-119-WW—Black Hole Section, Willow Wash, Fish Lake Valley, California.

§§N.A.—not available.

##Iron reported as FeO for standards.

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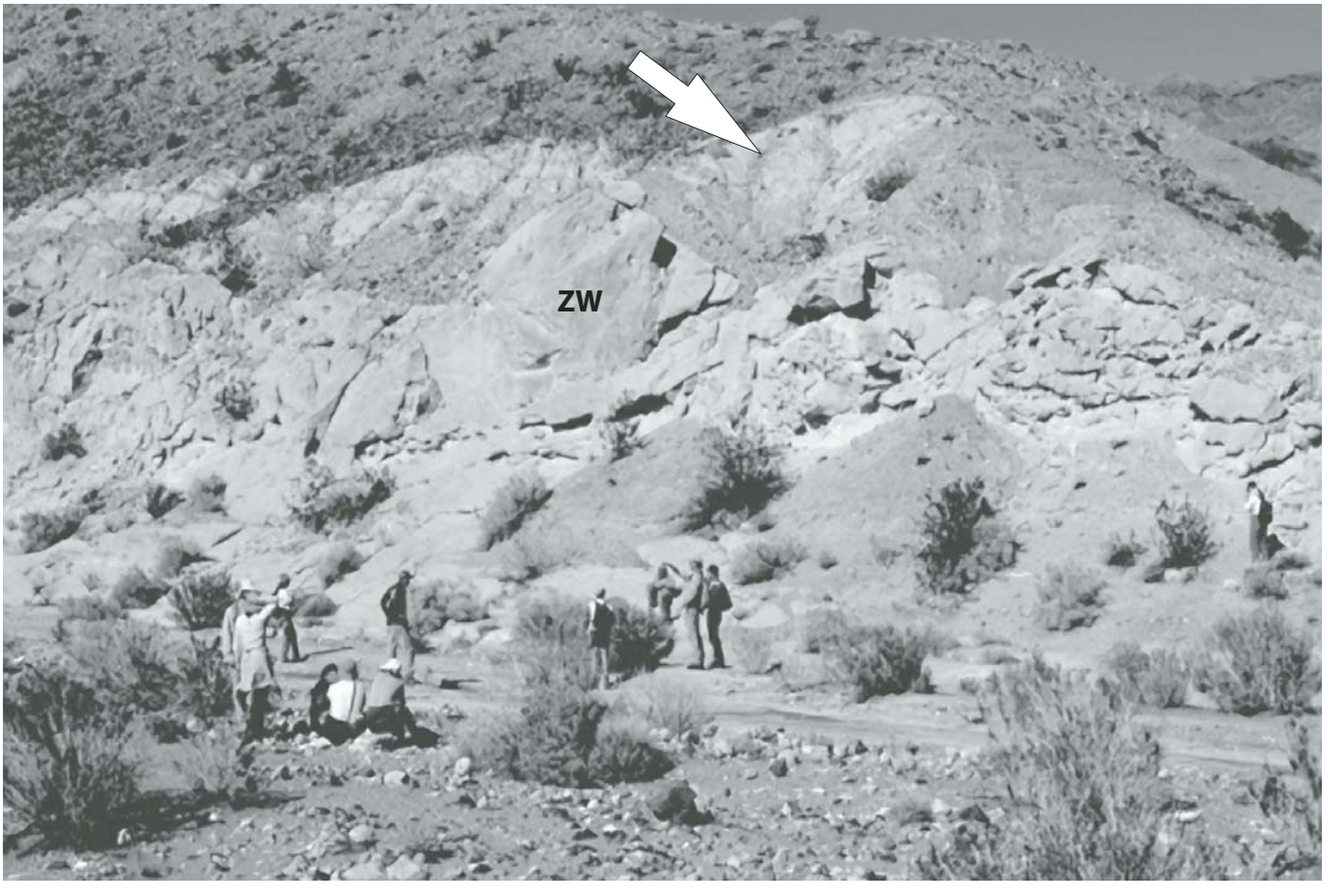

Figure 4. The tuff of Zabriskie Wash (ZW) exposed in Death Valley Wash where incised through the Lake Rogers Basin deposits. Tuff is ~10 m thick, is extensively faulted, dips ~45°E, and is overlain by fine-grained lacustrine deposits (white arrow).

that the source of the Artists Drive tuff is possibly in the southern Cascade Range of California, which is the source of the Nomlaki Tuff Member (Sarna-Wojcicki, 1976; Williams, 1994). The shard composition of the Nomlaki Tuff Member is heterogeneous (Williams, 1994) compared to the tuff of Artists Drive, which is observable in the CaO and Fe_2O_3 compositions (Fig. 7). All of the tephra beds at Artists Drive are intercalated with alluvial-fan sandstones and conglomerates of the Funeral Formation (Hunt and Mabey, 1966).

Lower to Middle Pleistocene Tephra Beds and Deposits

Mormon Point

Knott et al. (1999) correlated ash and tephra beds at Mormon Point along the Black Mountains (Fig. 2) with the 1.2–0.8 Ma upper Glass Mountain, 0.665 Ma Lava Creek B, and ca. 0.51 Ma Dibekulewe ash beds. Hayman et al. (2003) identified the Bishop ash bed at Mormon Point and Natural Bridge. These tephra layers are interbedded with both alluvial-fan and lake beds. Knott et al. (1999, 2005) named beds containing these lower to middle Pleistocene tephra layers the Mormon Point Formation (Knott et al., 2005).

With respect to Death Valley lakes, detailed mapping and analysis at Mormon Point (Tables 3 and 4) show that there are three lacustrine sequences of differing age there (Fig. 8). Each sequence is separated by alluvial-fan deposits (Fig. 9). The oldest lake sequence contains the 1.2–0.8 Ma upper Glass Mountain ash beds (Table 3, JRK-DV-8, JRK-DV-10, JRK-DV-36, JRK-DV-90, JRK-DV-91). The uppermost ash beds in two sections (JRK-DV-90 and JRK-DV-8) correlate well with the uppermost, reversed polarity ash bed near Long Valley (Glass Mountain D; Sarna-Wojcicki et al., 2005). Another ash bed found ~20 m lower in the section (Table 3, JRK-DV-10) at Mormon Point is correlative with the Glass Mountain E, F, or G ash beds, all of which have reversed polarity.

Fine-grained sand, silt, and clay with sparse massive breccia beds extend up the section from the upper Glass Mountain tephra layers to, but not above, the 0.759 Ma Bishop ash bed (JRK-DV-58), where the sediments change to alluvial-fan deposits (Fig. 9). A second lacustrine sequence is located just above the Bishop ash

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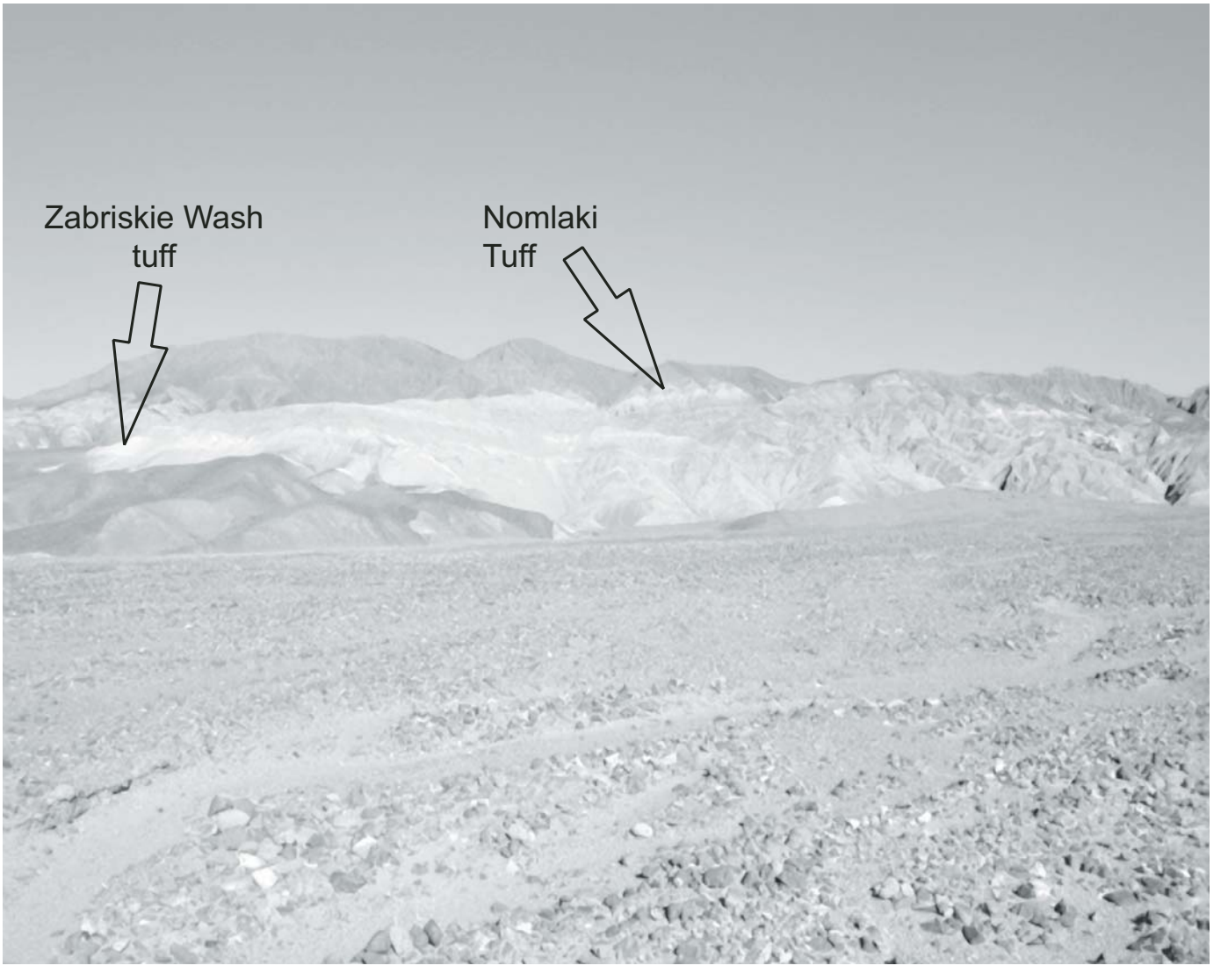

Figure 5. View east of the southern Kit Fox Hills showing light-colored lacustrine beds in the mid-ground overlain by dark-colored alluvial-fan deposits. The tuff of Zabriskie Wash is at the base, whereas the Nomlaki Tuff Member is found at the facies change near the skyline. Width of field of view is ~0.5 km.

bed, and this sequence terminates at the 0.665 Ma Lava Creek B ash bed. The Lava Creek B ash bed is well exposed in lacustrine sediments (Knott et al., 1999; Table 3, JRK-DV-70) to the west and in alluvial-fan deposits to the east (Table 3, JRK-DV-54, JRK-DV-55, JRK-DV-56, JRK-DV-57, JRK-DV-77, JRK-DV-78). The third lacustrine sequence, relatively thin and just above the ca. 0.51 Ma Dibekulewe ash bed, is dominated by a 2-m-thick marl (Fig. 8; Knott et al., 1999). Carbonate beds are common in all of the lacustrine sequences at Mormon Point (Fig. 9).

Paleomagnetic data for the Mormon Point section show that Bishop and Lava Creek ash beds are within a lengthy normal polarity interval. With only two exceptions, the sediment below the Bishop ash bed records reversed polarity, clearly indicating that this older part of the section is within the Matuyama (2.581–0.780 Ma; chron C1r.1r). The exceptions to reversed polarity are sites that are 1.2 m apart near sample locality 15 (Fig. 10). The paleomagnetic directions at those two sites are anomalous because they record transitional directions of northeasterly declination and negative inclination (10.0 m, D = 52.8°, I = −34°, VGP [Virtual Geogmagnetic Pole]: 15.9°N, 11.6°E) or southeasterly declination and positive inclination (11.2 m, D = 166.2°, I = 15.2°, VGP: 44.3°S, 262.6°E; Figs. 11 and 12). These data may indicate a transition to a normal polarity interval within the Matuyama reverse chron. The stratigraphic position of the transitional polarity interval relative to the ash beds (JRK-DV-90 and JRK-DV-91) that are correlative with the upper Glass Mountain D ash bed is consistent with the 1.070–0.990 Ma Jaramillo subchron. This interpretation is tentative, however, and requires additional data. Cumulatively, the Mormon Point tephrochronologic and paleomagnetic data define intermittent lake sequences that span the time period equivalent to marine oxygen isotope stages 34–16 (Shackleton et al., 1990).

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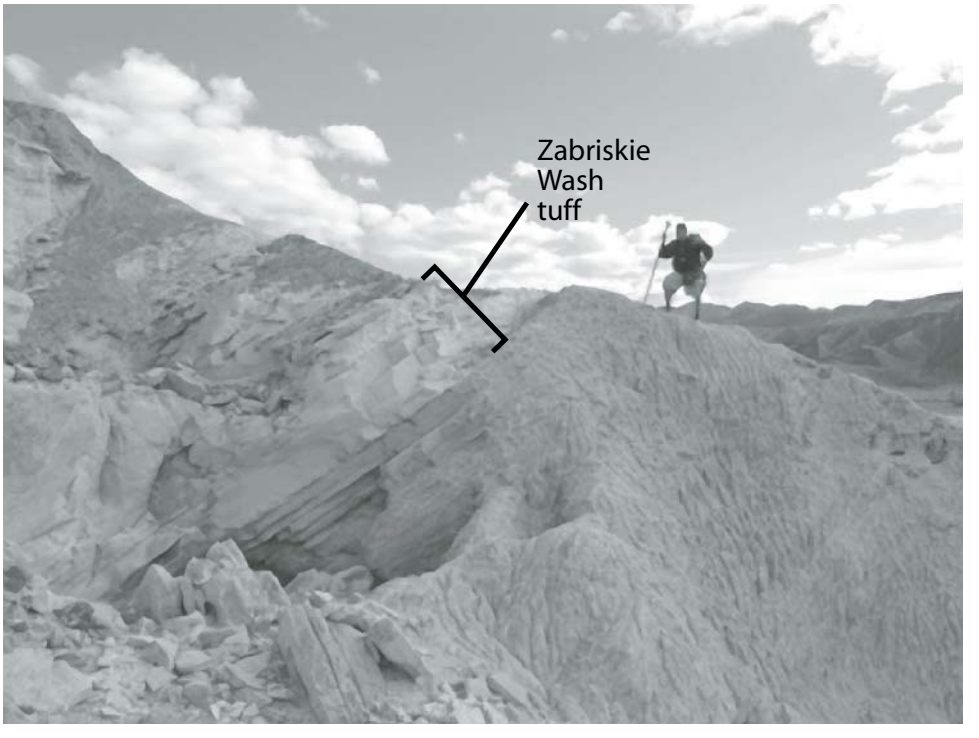

Figure 6. View of the tuff of Zabriskie Wash in Zabriskie Wash. Note the fine-grained, well-bedded lacustrine deposits above and below the tuff bed.

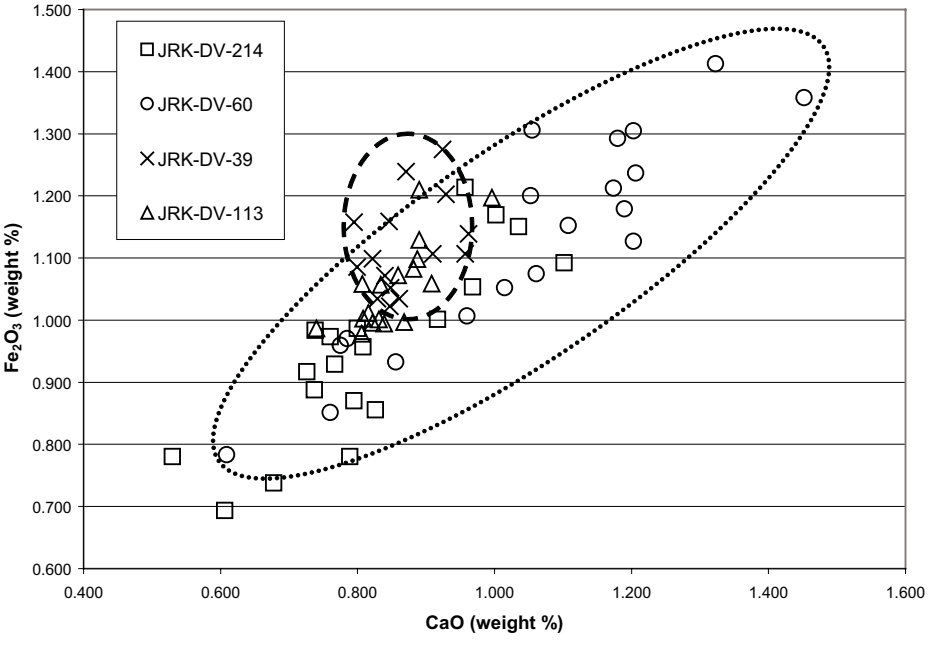

Figure 7. Scatter plot of CaO versus Fe_2O_3 (wt%) for representative Nomlaki Tuff Member (JRK-DV-214 and JRK-DV-60) and tuff of Artists Drive from Death Valley (JRK-DV-39 and JRK-DV-113). Dotted and dashed lines encompass the values of the Nomlaki Tuff Member and tuff of Artists Drive, respectively.

Natural Bridge

Natural Bridge is located ~6 km north of Badwater and 10 km south of Artists Drive (Fig. 2). Here, Hayman et al. (2003) correlated a highly faulted tephra layer with the 0.759 Ma Bishop ash bed. Additional mapping (Fig. 13) and tephrochronology (Table 4, JRK-DV-42, JRK-DV-43, and JRK-DV-62) show that the alluvial-fan deposits that include the Bishop ash bed extend substantially to the north and south along the Black Mountains. Unlike at Mormon Point, the mapping shows no evidence for a lake in the Natural Bridge area.

Kit Fox Hills

At the southern end of the Kit Fox Hills (Fig. 2), uplifted fine-grained beds were assigned to the Furnace Creek Formation by Wright and Troxel (1993). Although the exposures in this area are poor, the fine-grained texture suggests that the deposits are

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TABLE 3. RESULTS OF ELECTRON-MICROPROBE ANALYSIS OF VOLCANIC GLASS SHARDS FROM LOWER PLEISTOCENE TEPHRA LAYERS IN DEATH VALLEY AND SODA LAKE BASIN CORE, AND SIMILAR OR CORRELATIVE TEPHRA LAYERS ELSEWHERE

Sample*	Locality†	SiO_2	Al_2O_3	Fe_2O_3	MgO	MnO	CaO	TiO_2	Na_2O	K_2O	Total (o)§	SC#
Upper Glass Mountain (1.2–0.8 Ma) ash beds from Death Valley**												
JRK-DV-36	MP	77.16	12.71	0.78	0.04	0.03	0.43	0.07	3.68	5.11	95.20	1.0000
JRK-DV-8	MP	77.10	12.80	0.77	0.04	0.03	0.44	0.08	3.48	5.28	94.40	0.9663
JRK-DV-90	MP	77.50	12.39	0.76	0.04	0.04	0.43	0.06	3.44	5.35	94.90	0.9604
JRK-DV-91	MP	77.20	12.48	0.77	0.07	0.04	0.43	0.07	3.50	5.45	93.50	0.9937
BT-2A††		77.27	12.90	0.79	0.03	0.03	0.43	0.06	3.68	4.81	95.12	0.9657
BT-14A		77.21	12.78	0.75	0.03	0.04	0.43	0.06	3.77	4.92	95.36	0.9625
BT-15		77.18	12.77	0.77	0.04	0.03	0.46	0.06	3.85	4.85	94.90	0.9548
BT-1		77.32	12.91	0.80	0.03	0.03	0.43	0.05	3.85	4.57	95.56	0.9343
Upper Glass Mountain (1.2–0.8 Ma) ash bed from Death Valley												
JRK-DV-10	MP	77.20	12.70	0.76	0.04	0.04	0.43	0.06	3.50	5.25	94.90	0.9660
BT-14A		77.21	12.78	0.75	0.03	0.04	0.43	0.06	3.77	4.92	95.36	0.9961
BT-15		77.18	12.77	0.77	0.04	0.03	0.46	0.06	3.85	4.85	94.90	0.9832
BT-1		77.32	12.91	0.80	0.03	0.03	0.43	0.05	3.85	4.57	95.56	0.9531
Upper Glass Mountain (1.2–0.8 Ma) ash bed from Soda Lake core												
JRK-SL-599	SL	78.31	11.78	0.73	0.03	0.03	0.43	0.06	3.93	4.05	92.66	1.0000
BT-2A		77.27	12.90	0.79	0.03	0.03	0.43	0.06	3.68	4.81	95.12	0.9497
BT-14A		77.21	12.78	0.75	0.03	0.04	0.43	0.06	3.77	4.92	95.36	0.9612
BT-15		77.18	12.77	0.77	0.04	0.03	0.46	0.06	3.85	4.85	94.90	0.9446
BT-1		77.32	12.91	0.80	0.03	0.03	0.43	0.05	3.85	4.57	95.56	0.9165

Note: Values are given in mean wt% oxide, recalculated to 100% fluid-free basis; C.E. Meyer (1977–1996), U.S. Geological Survey, Menlo Park, California, analyst.

*Samples are grouped according to best matches (inferred correlations).

†Abbreviations for Death Valley localities are as follows: MP—Mormon Point, Death Valley; SL—Soda Lake Basin, Mojave Desert.

§Original oxide totals before calculation are given to indicate approximate degree of hydration of volcanic glass. Original values for each oxide can be obtained by multiplying Total(o) by the weight percent shown.

#Similarity coefficient (SC) of Borchardt et al. (1972) used to quantitatively compare the tephra samples, where 1.0000 represents a perfect match.

**See text for source of estimated or isotopic ages from other localities in the western United States.

††Data from correlative samples. BT-2A—ash of Glass Mountain (D), water deposited, south-central end of Volcanic Tableland, south of Long Valley, California; BT-14A—ash of Glass Mountain (E), air fall, underlying BT-2A; BT-15—pumice of Glass Mountain (F), reworked, underlying BT-14A; BT-1—ash of Glass Mountain (G), underlying BT-15 (Sarna-Wojcicki et al., 2005).

lacustrine or playa in origin. A 50-cm-thick tephra bed within the fine-grained deposits is correlative with Lava Creek B (Table 4, REK-DV-14). Thus, these deposits are the time-stratigraphic equivalent of the Mormon Point Formation and are not equivalent to the considerably older Furnace Creek Formation.

Soda Lake Basin Core

Pluvial Lake Mojave lay upstream along the Mojave River from Death Valley and Lake Manly (Fig. 1). Lake Mojave, which incorporated the present-day Silver Lake and Soda Lake Basins during pluvial periods (Wells et al., 1984), occupied a key geomorphic position relative to Lake Manly. Extremely deep water in Death Valley potentially could back up Lake Manly into the Silver and Soda Lake Basins of the Mojave River system and overflow toward the Colorado River system before inundating the topographically higher Amargosa or Owens River systems (Hale, 1985).

More than 19 different cores have been extracted from the Soda Lake and Silver Lake Basins (Brown and Rosen, 1995). Muessig et al. (1957) described five cores ranging in depth from 24 m to 324 m (Fig. 14). The Soda Lake Basin cores have green clays between 9 and 36 m depth that are interpreted as lacustrine (Muessig et al., 1957). Below ~36 m depth, the cored sediments consist of brown to orange-brown clastic deposits interpreted as nonlacustrine (Muessig et al., 1957). Core 1, which is near the center of the basin, contains three volcanic ashes at depths of 121, 149, and 182 m within the nonlacustrine interval (Muessig et al., 1957). The ash bed at the 121 m depth has a thickness of 0.3 m (Muessig et al., 1957).

Two samples, at 124 m and 182 m, yielded enough volcanic glass for electron microprobe analysis. The volcanic glass collected at 124 m has a chemical composition similar to tephras erupted from the Long Valley volcanic center in eastern California. Based on the morphology, thickness, and chemical composition of the vesicular shards, the 124-m-deep sample is correlated

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TABLE 4. RESULTS OF ELECTRON-MICROPROBE ANALYSIS OF VOLCANIC GLASS SHARDS FROM MIDDLE PLEISTOCENE TEPHRA LAYERS IN DEATH VALLEY, SODA LAKE BASIN CORE, AND SIMILAR OR CORRELATIVE TEPHRA LAYERS ELSEWHERE

Sample*	Locality†	SiO_2	Al_2O_3	Fe_2O_3	MgO	MnO	CaO	TiO_2	Na_2O	K_2O	Total (o)§	SC#
Lava Creek B ash bed (ca. 0.665 Ma)**												
JRK-DV-54	MP	76.80	12.20	1.53	0.02	0.03	0.53	0.11	3.49	5.26	94.90	1.0000
JRK-DV-55	MP	76.90	12.10	1.52	0.02	0.03	0.53	0.10	3.48	5.26	94.80	0.9786
JRK-DV-56	MP	76.90	12.10	1.53	0.03	0.03	0.53	0.10	3.52	5.29	94.00	0.9799
JRK-DV-57	MP	76.90	12.10	1.57	0.02	0.04	0.53	0.10	3.52	5.22	94.50	0.9748
JRK-DV-70	MP	76.80	12.20	1.57	0.03	0.03	0.52	0.10	3.45	5.34	95.30	0.9729
JRK-DV-77	MP	77.14	12.18	1.49	0.02	0.03	0.51	0.10	3.34	5.18	94.51	0.9678
JRK-DV-78	MP	77.24	12.13	1.49	0.02	0.03	0.52	0.10	3.32	5.15	93.67	0.9705
REK-DV-14	KFH	77.04	12.17	1.55	0.02	0.04	0.54	0.11	3.49	5.03	94.39	0.9926
TECO-30B-1††		76.40	12.50	1.57	0.02	0.04	0.53	0.12	3.52	5.30	94.40	0.9724
ORNA-1		76.80	12.30	1.58	0.02	0.03	0.53	0.12	3.61	5.07	95.00	0.9754
Bishop ash bed (0.759 Ma) from Death Valley												
JRK-DV-58	MP	77.50	12.30	0.74	0.03	0.04	0.44	0.06	3.46	5.39	93.20	1.0000
JRK-DV-42	NB	77.60	12.60	0.72	0.03	0.03	0.44	0.06	3.78	4.20	94.30	0.9896
JRK-DV-43	NB	77.30	12.90	0.76	0.03	0.03	0.44	0.06	4.20	4.34	93.90	0.9849
JRK-DV-62	NB	77.90	12.40	0.77	0.03	0.03	0.43	0.06	4.00	4.39	92.00	0.9850
JRK-DV-93	MP	76.80	13.24	0.70	0.03	0.03	0.43	0.05	3.54	5.21	91.90	0.9353
JRK-DV-99	MP	77.10	12.95	0.69	0.03	0.03	0.43	0.05	3.37	5.34	93.60	0.9375
BT-8		77.40	12.70	0.72	0.04	0.04	0.45	0.06	3.34	5.22	94.20	0.9935
OL92–1021		76.91	13.00	0.72	0.03	0.03	0.45	0.06	4.00	4.81	95.46	0.9779
Bishop (0.759 Ma) ash bed from Soda Lake Basin core												
JRK-SL-407	SL	78.11	12.74	0.74	0.03	0.03	0.43	0.06	3.92	4.04	92.90	1.0000
BT-8		77.40	12.70	0.72	0.04	0.04	0.45	0.06	3.34	5.22	94.20	0.9688
OL92–1021		76.91	13.00	0.72	0.03	0.03	0.45	0.06	4.00	4.81	95.46	0.9643

Note: Values are given in mean wt% oxide, recalculated to 100% fluid-free basis; C.E. Meyer (1977–1996), U.S. Geological Survey, Menlo Park, California, analyst.

*Samples are grouped according to best matches (inferred correlations).

†Abbreviations for Death Valley localities are as follows: MP—Mormon Point; KFH—Kit Fox Hills; NB—Natural Bridge (all in Death Valley); SL—Soda Lake Basin.

§Original oxide totals before calculation are given to indicate approximate degree of hydration of volcanic glass. Original values for each oxide can be obtained by multiplying Total(o) by weight percent shown.

#Similarity coefficient (SC) of Borchardt et al. (1972) used to quantitatively compare the tephra samples, where 1.0000 represents a perfect match.

**See text for source of estimated or isotopic ages from other localities in the western United States.

††Data for correlative samples. TECO-30B-1—lacustrine beds, Lake Tecopa, California; ORNA-1—lacustrine beds, near Oreana, Nevada; BT-8—Bishop tuff, basal pumice fall beneath ash flow, southeast of Volcanic Tableland, south of Long Valley, California; OL92–1021—Bishop ash bed from 303.94 m depth in Owens Lake core, Owens Lake, California.

with the 0.759 Ma Bishop ash bed (Table 4, JRK-SL-407). Volcanic glass separated from the 182 m sample (Table 3, JRK-SL 599) also has a chemical composition consistent with Long Valley volcanic-center eruptions. Based on its position below the Bishop ash bed, the 182-m-deep ash bed may be correlative with any of the 1.2–0.8 Ma upper Glass Mountain family of ash beds (Table 3).

DISCUSSION

Pliocene Lakes of Death Valley

The present southeastern California–western Nevada drainage system (Fig. 1) that includes the Owens, Amargosa, and Mojave Rivers was substantially different 3–2 m.y. ago (Fig. 15). The 3–2 Ma time period is when the mtDNA data imply that pupfish began diverging from a common ancestor due to severing of a hydrologic connection (Smith et al., 2002; Echelle et al., 2005; Echelle, this volume). During Quaternary pluvial periods, Lake Russell filled the Mono Basin at the Owens River headwaters. Before 1.6 Ma, however, Lake Russell flowed north into the Walker Basin (Fig. 15; Reheis et al., 2002a; 2002b), not to the Owens River. In the Searles Basin (Fig. 2), the KM-3 core shows that 3–2 Ma Searles Lake was perennial and overflowing into Panamint Valley (Fig. 15; Jannik et al., 1991; Smith, 1984). Unfortunately, there are no data regarding Pliocene lakes in Panamint Valley, so the evidence for a Pliocene connection of the Owens River to Death Valley ends in Searles Valley (Jannik et al., 1991).

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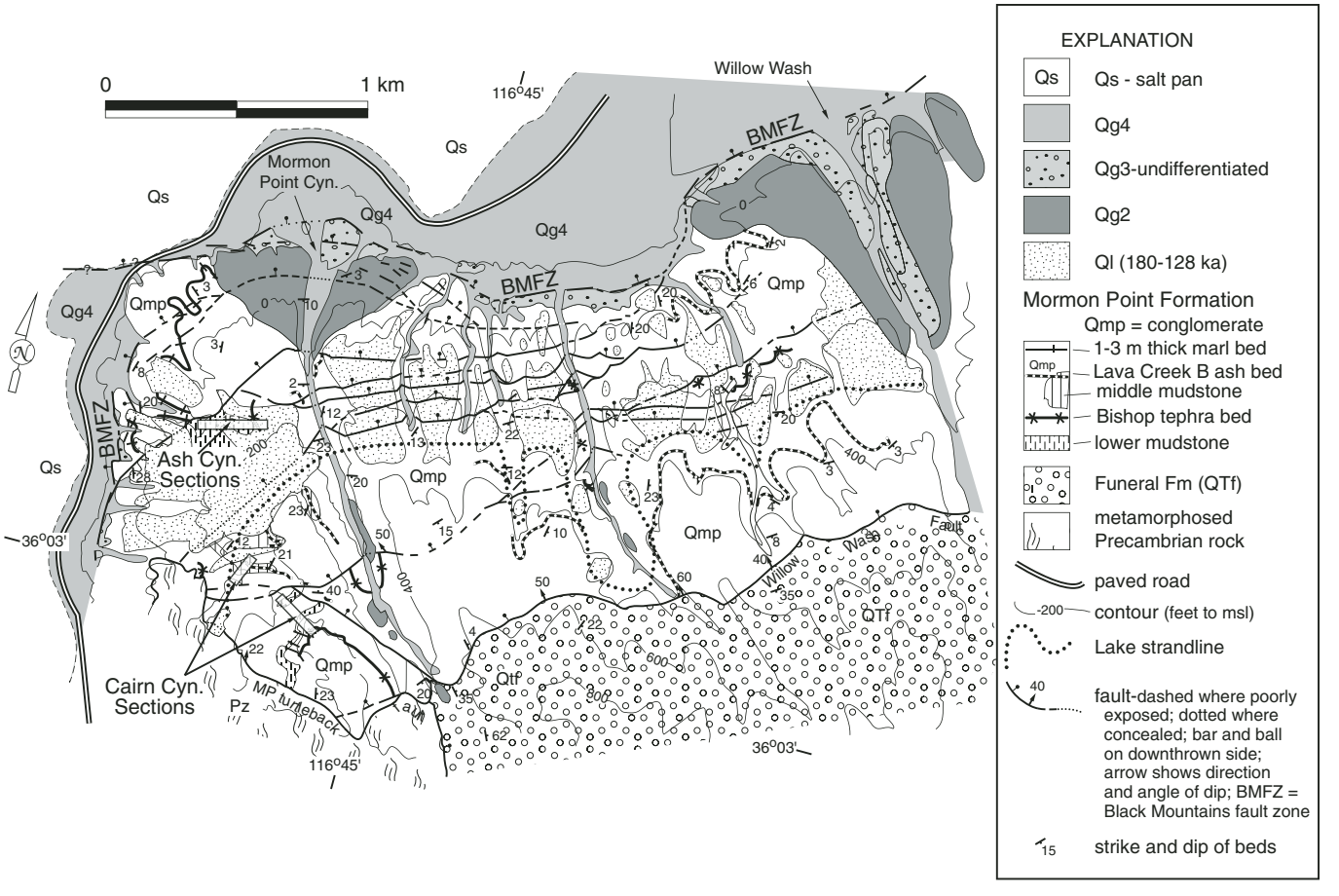

Figure 8. Geologic map of Mormon Point area (msl—meters above sea level). Rectangles show locations of measured stratigraphic sections in Ash and Cairn Canyons (after Knott et al., 2002).

The present fluvial connection from the Amargosa River to Death Valley was not present during the Pliocene. Deposits in the Lake Tecopa Basin (Fig. 15) show that the Amargosa River would have terminated there beginning at least 2 m.y. ago, and it did not have a surface hydrologic connection to Death Valley until some time after 600 ka and perhaps as recently as 160 ka (Hillhouse, 1987; Morrison, 1991). Thus, pupfish dispersal by surface-water systems along the Amargosa River is not a factor in Pliocene reconstructions.

The present-day Mojave River system was not present during the Pliocene either. Cox et al. (2003) showed that up until 3.5 Ma, an ancestral Mojave River flowed to the south through Cajon Pass and into the Los Angeles Basin (Fig. 15). There is no evidence of a downstream connection to the Soda Lake Basin during the Pliocene (Cox et al., 2003; Jefferson, 2003). Thus, there is no evidence of a Mojave River–Death Valley connection in the late Pliocene (from at least 3–2 Ma) as might be suggested by the pupfish mtDNA data.

In reconstructing hydrologic connections, outflow from a Death Valley lake and river system into ancestral Owens, Amargosa, and Mojave River systems must also be considered as a possible connection. Hunt and Mabey (1966) and Wright et al. (1999) described clast assemblages and facies within the 14–6 Ma Artists Drive Formation that are consistent with a northwest-to-southeast–flowing fluvial system during the late Miocene. This Miocene fluvial system originated in the Cottonwood Mountains and flowed southeast out of the Death Valley area roughly parallel to the Furnace Creek fault zone into the Amargosa Valley to the east (Wright et al., 1999; Renik and Christie-Blick, 2004). Clast assemblages and facies indicate that the fluvial system of the overlying Furnace Creek Formation, estimated to be 6–5 Ma by Wright et al. (1999), was more localized, having source areas in both the adjoining Funeral and Black Mountains and including fine-grained, laminated silts, and clays interpreted as lacustrine (Blair and Raynolds, 1999; Hunt and Mabey, 1966; McAllister, 1970; Wright et al., 1999). The ages of the Artist Drive and Furnace Formations were estimated from relatively sparse data. The tephrochronologic data presented here shows that the age of the upper Furnace Creek Formation is ca. 3.35 Ma, which is time correlative with deposits in the Nova Basin, Artists Drive, Kit Fox Hills, and Lake Rogers.

Using the 3.35 Ma Zabriskie Wash tuff, which is a widespread time-stratigraphic marker for deposits in the upper Furnace Creek Formation and coeval units, we can interpret the

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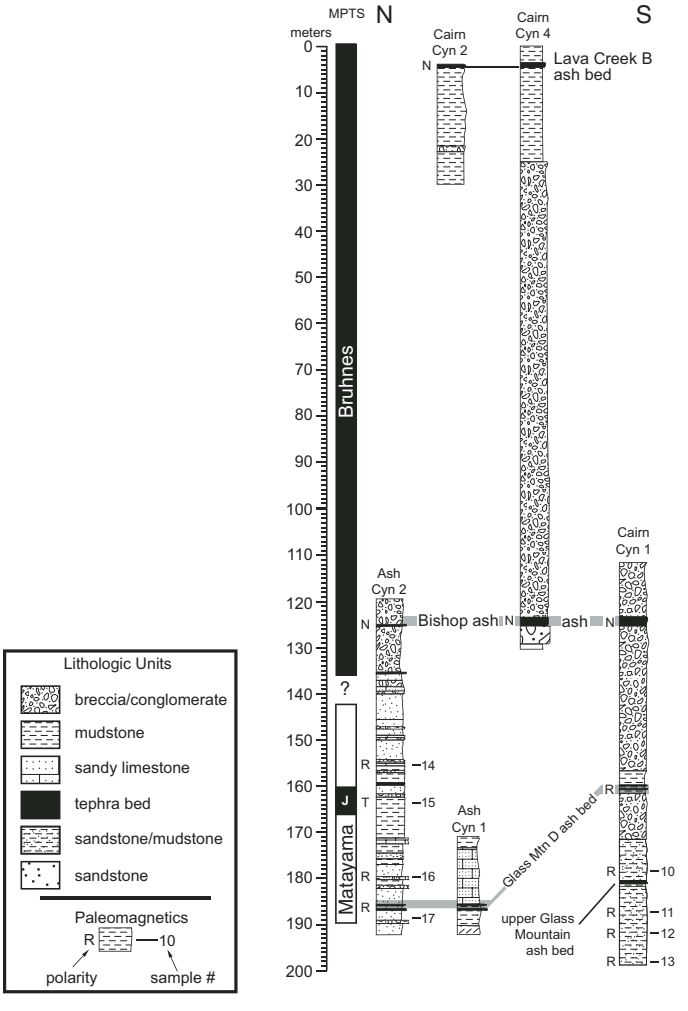

Figure 9. Measured stratigraphic sections in Mormon Point Formation at Mormon Point. Magnetic polarity time scale (MPTS) is shown at left with tentative location of Jaramillo subchron with respect to Ash Canyon 2 section.

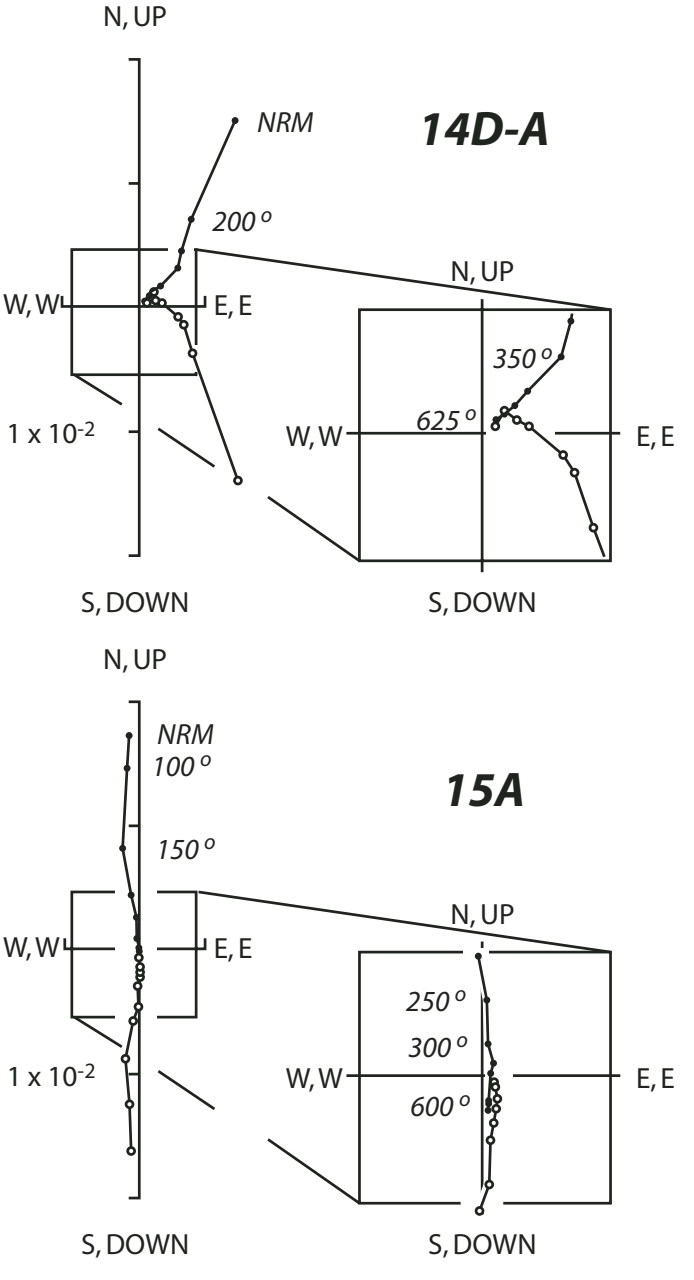

Figure 10. Vector end-point plots for samples from lacustrine deposits at Mormon Point, Death Valley, treated by thermal demagnetization. Temperatures shown are °C. Sample 14D is 1.2 m higher than sample 15. Note in the inserts for each sample, the paleomagnetic directions are anomalous at high temperatures of demagnetization (>500 °C). At those temperatures, the inclination of sample 14D-A is negative, and for sample 15A, the declination is southerly. Intensity is shown in Am^{-1} units. NRM—natural remnant magnetization.

paleogeography and configuration of the ancient Furnace Creek Basin (Fig. 16). The Zabriskie Wash tuff is interbedded with lacustrine clay and silt at Zabriskie Wash, the Kit Fox Hills, and Lake Rogers. At Dante's View Road, Artists Drive, Nova Basin, and the Cottonwood Mountains piedmont, the Zabriskie Wash tuff is interbedded with alluvial-fan deposits. This indicates that 3.35 m.y. ago, a narrow northwest-southeast–trending lake occupied 80 km of the ancient Furnace Creek Basin. Clast assemblages show that the alluvial-fan sources were local. For example, Dante's View and Artists Drive alluvial fans originated in the Black Mountains. McAllister (1970) showed that the locally derived alluvial fans, like those containing the Zabriskie Wash tuff along Dante's View Road (Fig. 16), encircled the southeastern end of the Furnace Creek Basin. Therefore, the Miocene through-flowing fluvial system of Wright et al. (1999) was not flowing southeast out of Death Valley 3.35 m.y. ago.

Over 90 km south of Artists Drive, Caskey et al. (2006) correlated tephra layers interbedded with lacustrine sediments near the intersection of the Southern Death Valley and Garlock fault zones with the Mesquite Spring family of tuffs. Between these localities, there are three additional outcrops of Mesquite Spring

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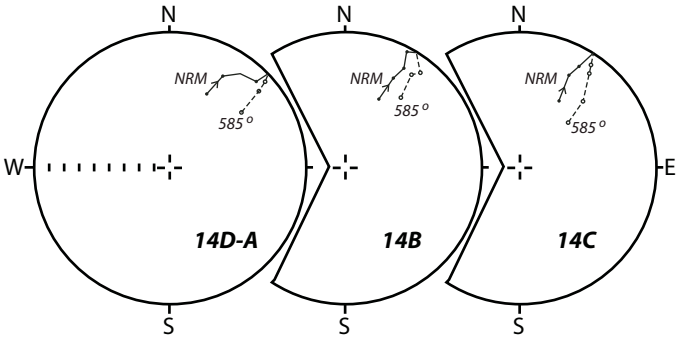

Figure 11. Equal-area plots of paleomagnetic directions for samples 14D-A, 14B, and 14C following thermal demagnetization to 585 °C. Note that for each sample, the inclination is negative when declination is northeasterly; these directions result in a VGP (Virtual Geomagnetic Pole) near the equator. Temperatures shown are °C; NRM—natural remnant magnetization.

family tephra layers, all of which are interbedded with alluvial-fan deposits (Knott et al., 1997, 1999). There are insufficient data to preclude or support the interpretation that a continuous lake extended south from the Furnace Creek area into central and southern Death Valley at 3.35 Ma; however, the alluvial fans at Artists Drive and elsewhere confine any southern arm of the ancestral lake to a relatively narrow area of the valley.

To the north, at Lake Rogers Basin, lacustrine deposition ceased just after deposition of the 3.35 Ma tuff of Zabriskie Wash (Fig. 17). By 3.29–3.28 Ma, the ancestral Furnace Creek Basin lake had receded sufficiently such that the 3.28 Ma Nomlaki Tuff Member is interbedded with alluvial-fan deposits at the Kit Fox Hills, Artists Drive, and Zabriskie Wash localities. Only at Salt Creek is the 3.29 Ma Putah Tuff Member found interbedded with lacustrine beds.

Data regarding a post–3.28 Ma lake of the Furnace Creek Basin are sparse. The 3.07 Ma tuff of Salida Canyon is found in the Kit Fox Hills interbedded with alluvial-fan deposits. At 3.12–3.06 Ma, unnamed tuffs from the Coso volcanic field (in the Owens Valley) were deposited in distal alluvial-fan deposits at Zabriskie Wash. Beratan et al. (1999) showed that 2.07–1.7 Ma beds in the Confidence Hills of southern Death Valley were deposited in semiarid to arid conditions, with no evidence of a large lake. Knott et al. (1999) found 1.9–1.7 Ma tephra beds at Artists Drive interbedded with alluvial-fan deposits and thin marl. These observations are consistent with no fluvial outlet south toward pluvial Lake Mojave and no large lake in Death Valley ca. 2.0–1.7 Ma.

In summary, the ca. 3.5 to 1.7 Ma tephra beds and facies distributions in Death Valley are consistent with a northwest-southeast–trending lake bounded by prograding alluvial fans with sources in the adjacent mountain ranges bordering a relatively narrow, but expanding Death Valley. By 3.35 Ma, the southeastern end of the Furnace Creek Basin, where streams flowed out of Death Valley during the Miocene (Wright et al., 1999), was cut off by relative uplift of the Funeral and Black Mountains and subsequent deposition of prograding alluvial fans. Alluvial-fan deposits interbedded with the tuff of Salida Canyon at the Kit Fox

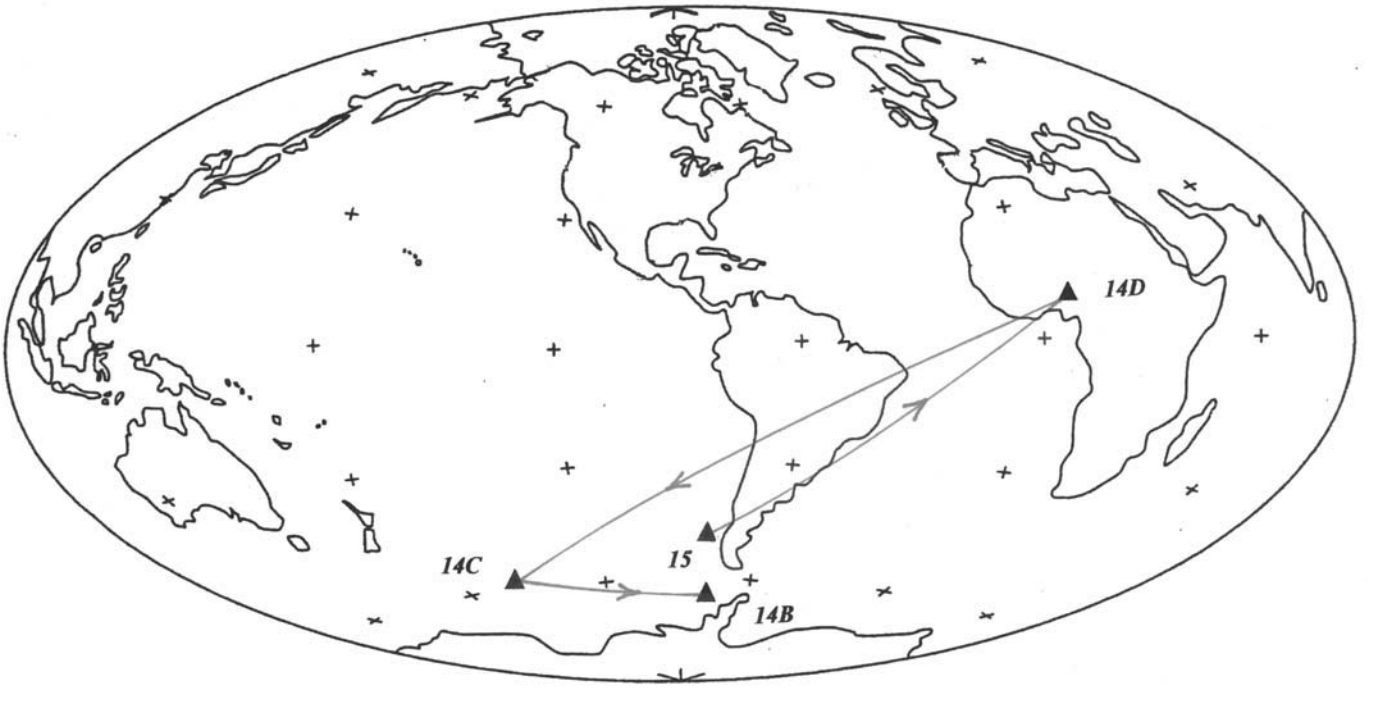

Figure 12. Map of Earth with Virtual Geomagnetic Poles (VGPs) for sites 14B, 14C, 14D, and 15. Site 14D has a VGP that is near the equator (15.9°N, 11.6°E) and is anomalous relative to samples 14C and 15, which bracket it.

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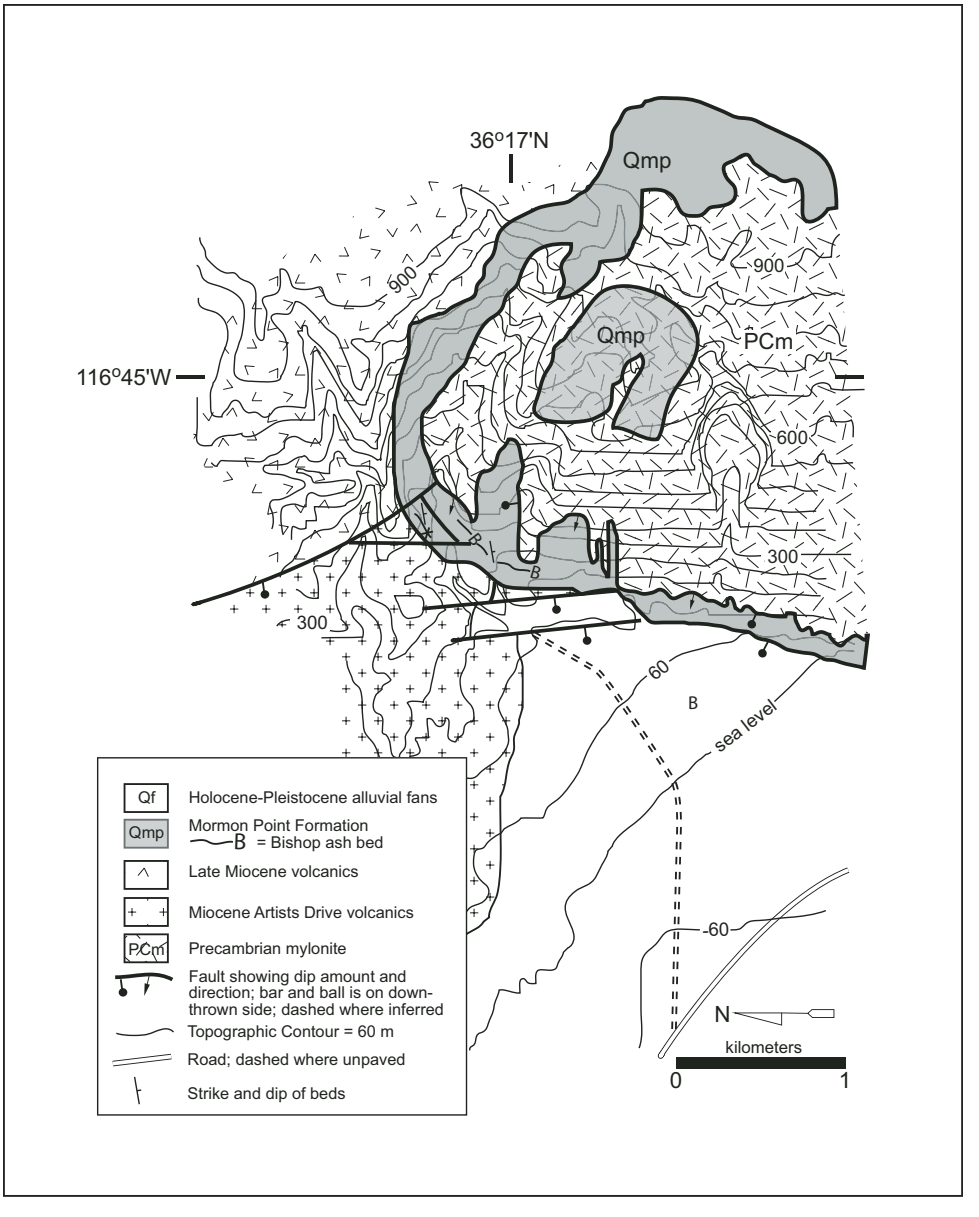

Figure 13. Geologic map of Natural Bridge, Death Valley. Mormon Point Formation conglomerates are in fault contact with Precambrian and Miocene rocks. Volcanic rocks were mapped by Hunt and Mabey (1966).

Hills and the unnamed Coso tuffs near Zabriskie Wash show that alluvial fans had prograded into the central part of the ancient Furnace Creek Basin by 3 Ma. Therefore, the ancient lake was gone by that time. Alluvial fans and playa deposits show that a dry alluvial fan and playa environment existed south of the ancient Furnace Creek Basin from 2.0 to 1.7 Ma with no fluvial connection to Lake Mojave and the Mojave River system.

The geologic data leave no outlet to pluvial Lake Mojave and the Colorado River beyond, no connection to the Amargosa River, and only a possible connection to the Owens River during the late Pliocene (ca. 3.5–3 Ma). The Amargosa River terminus was at Lake Tecopa (Morrison, 1991), and the incipient ancestral Mojave River was confined to an area adjacent to the San Gabriel Mountains (Cox et al., 2003) at this time. The thick, ca. 3 Ma evaporite deposits suggest that Searles Valley was the terminus of the Owens River at that time (Phillips, this volume). If overflow from the Owens River system occurred during the late Pliocene, as suggested by the mtDNA data (Smith et al., 2002; Echelle et al., 2005; Echelle, this volume), the overflow was not sufficient to produce an extensive lake in Death Valley.

Early to Middle Pleistocene Death Valley Lakes

Climate proxy records for the Mojave Desert region show that the early to middle Pleistocene (1.2–0.5 Ma) was one of the wetter climatic periods in the past 2 m.y. (Smith, 1991). Although 1.2 Ma is not the time frame of the molecular data of Smith et al. (2002), it is possible that nonsimultaneous connections among the

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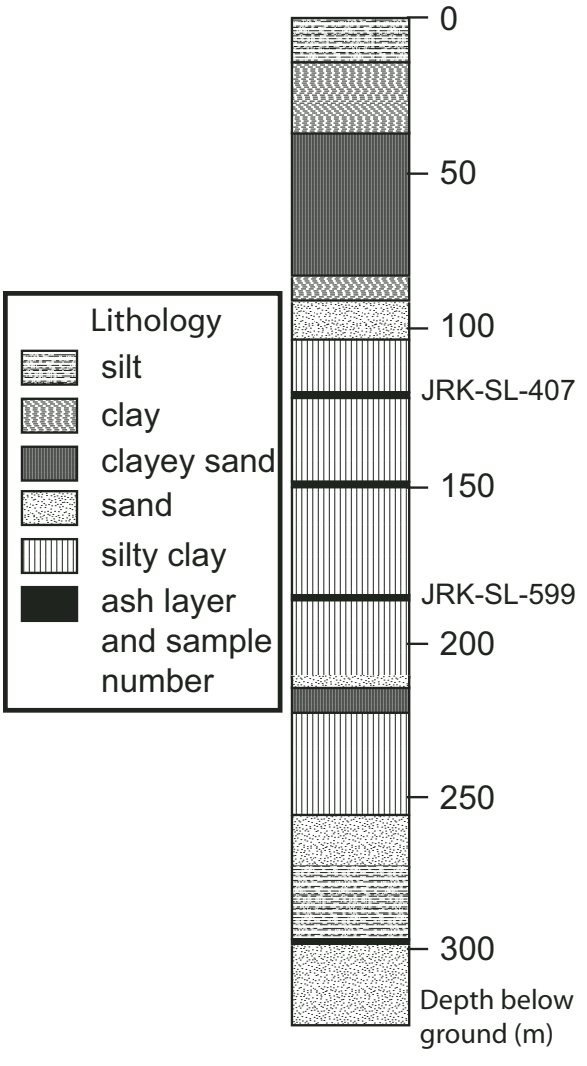

Figure 14. Composite lithologic log of Soda Lake Basin cores 1 and 2 as described by Muessig et al. (1957). Locations of samples JRK-SL-407 and JRK-SL-599 are ash layers logged at 121 and 182 m depth, respectively. The thicknesses of the ash layers are exaggerated so that their locations are apparent in the diagram (after Brown and Rosen, 1995).

Owens, Amargosa, Mojave, and Colorado Rivers either provided dispersal pathways for pupfish or isolated species as connections ceased. Thus, it is crucial to determine the timing and extent of hydrologic connections among the various rivers and lakes of southeastern California, especially during pluvial periods and after the hypothesized time of pupfish dispersal 2–3 m.y. ago.

During the early to middle Pleistocene, geologic data show that the Amargosa and Mojave Rivers terminated at Lake Tecopa (Hillhouse, 1987; Morrison, 1991; Menges, this volume) and Lake Manix (Jefferson, 2003) or farther upstream (Fig. 18). Thus, only the Owens River could have flowed into Death Valley. Interpretation of Searles Lake core KM-3 data indicates that between 1.2 and 0.6 Ma, Searles Lake spilled into Panamint Valley four different times (Jannik et al., 1991). Although core data from Panamint Valley are sparse (Jannik et al., 1991), outcrop data show only a small lake or marsh interval between 0.9 and 0.759 Ma, and no evidence of a lake capable of spilling into Death Valley (Larsen et al., 2003).

In Death Valley, the Bishop ash bed at Natural Bridge is interbedded with alluvial-fan deposits. The presence of the 0.759 Ma Bishop ash bed in nonlacustrine sediments at Mormon Point (Fig. 2) indicates that by 0.759 Ma, the lake had receded from there. This recession, which coincided with the Bishop ash bed, was followed by another rise in lake level; lacustrine deposits containing the 0.665 Ma Lava Creek B ash bed are preserved at Mormon Point and the southern Kit Fox Hills. At Mormon Point, the Lava Creek B ash bed is interbedded with lacustrine silt and clay to the west and south of Mormon Point. To the north and east of Mormon Point, the Lava Creek B ash bed is interbedded with alluvial-fan deposits derived from the east. Sedimentary facies within the Mormon Point section have rapid vertical transitions from silt and clay to carbonates with intercalated debris-flow deposits. The debris-flow deposits suggest that the area was proximal to the shoreline. The carbonate beds are interpreted as evidence of increased salinity in a shrinking lake. All of these data are consistent with Mormon Point being at the paleolake shoreline 0.6 m.y. ago.

Caskey et al. (2006) correlated upper tuffs of Glass Mountain (1.2–0.8 Ma) and Lava Creek B (ca. 0.665 Ma) layers that are interbedded with lacustrine sediments south of the Confidence Hills (Fig. 2) in Death Valley. The limited outcrops allow contrasting interpretations of the extent or depth of the Death Valley lake from 1.2 to 0.5 Ma. Early to middle Pleistocene Death Valley may have been inundated by either an areally extensive lake or several disconnected smaller lakes of limited areal extent. A lake of limited extent is consistent with the relatively small drainage area of Death Valley at that time (Fig. 18). A small Death Valley lake, however, contrasts with other basins in the western United States that contained substantial lakes at that time (Reheis et al., 2002a; Morrison, 1991).

Back-flow of Lake Manly during Marine Oxygen Isotope Stage (MIS) 6

Outcrop and geochemical data imply that by marine oxygen isotope stage 6 (MIS 6), between 180 and 128 ka (Fig. 19), the Amargosa and Owens Rivers flowed into Death Valley (Morrison, 1991; Lowenstein et al., 1999; Larsen et al., 2003). In contrast, the Mojave River terminated at Lake Manix and did not flow into pluvial Lake Mojave (Meek, 1989; Enzel et al., 2003; Jefferson, 2003). Lake Mojave covered both Soda Lake (southern end) and Silver Lake (northern end) basins during wet, cold periods of the latest Pleistocene and Holocene (Enzel et al., 1992; Wells et al., 2003). Radiocarbon ages and sedimentation rates from cores, along with soil development on adjacent alluvial fans, provide age estimates of ca. 22–8.5 ka (^{14}C yr B.P.) for pluvial Lake Mojave stands (Ore and Warren, 1971; Wells et al., 1984; Reheis

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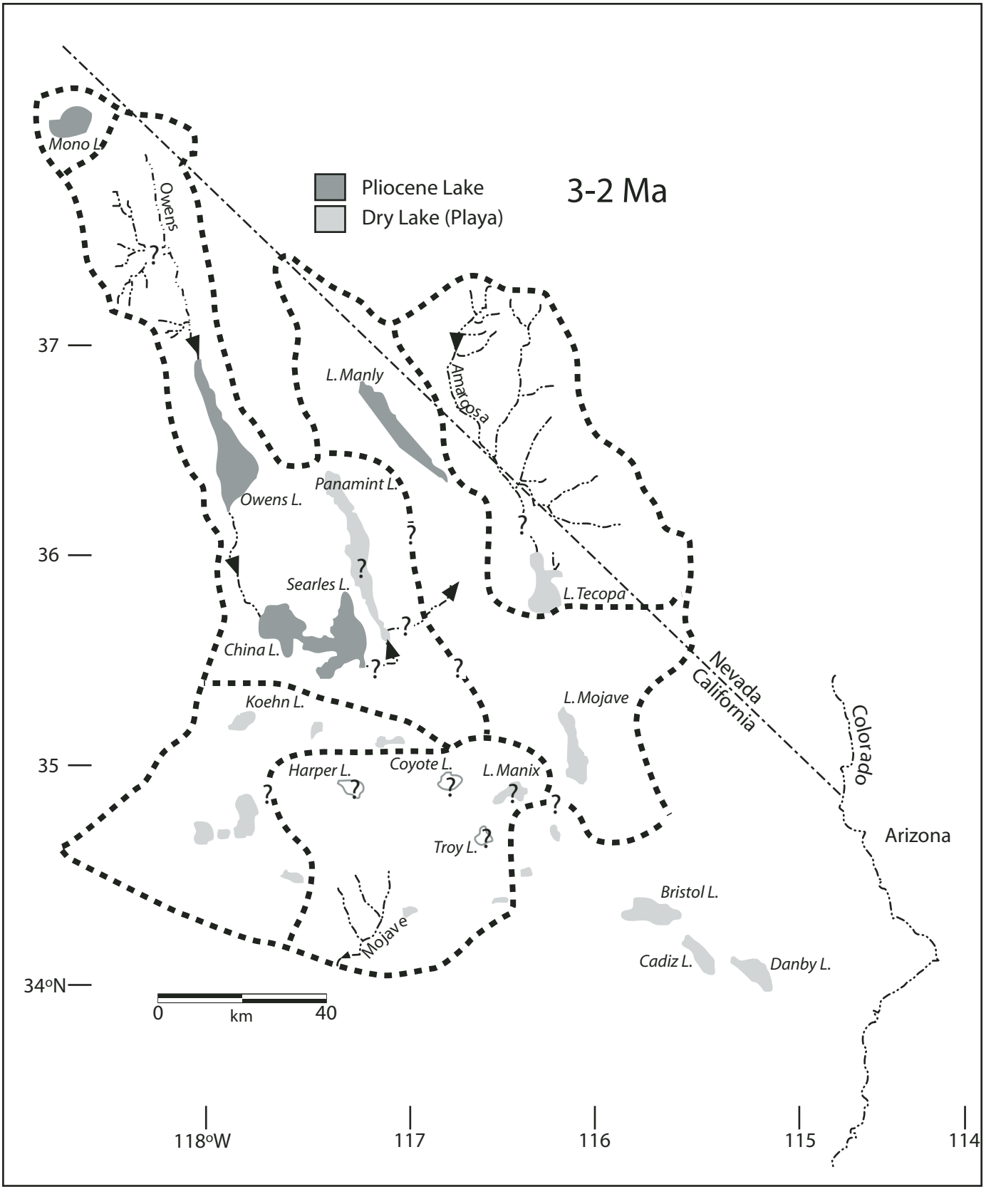

Figure 15. Late Pliocene (3–2 Ma) paleohydrology of southeastern California and western Nevada. Pluvial lakes are shaded gray, and dot-dash lines are inferred fluvial connections. The thick dashed line shows the approximate drainage basin limits for each drainage system. Question marks indicate no supporting data for lake or river. For reference, the outline of the present-day (Harper, Coyote, and Troy) playa is shown and queried. See text for details. Note that the Mojave River is flowing south into the Los Angeles Basin and that Mono Lake is not integrated with the Owens River.

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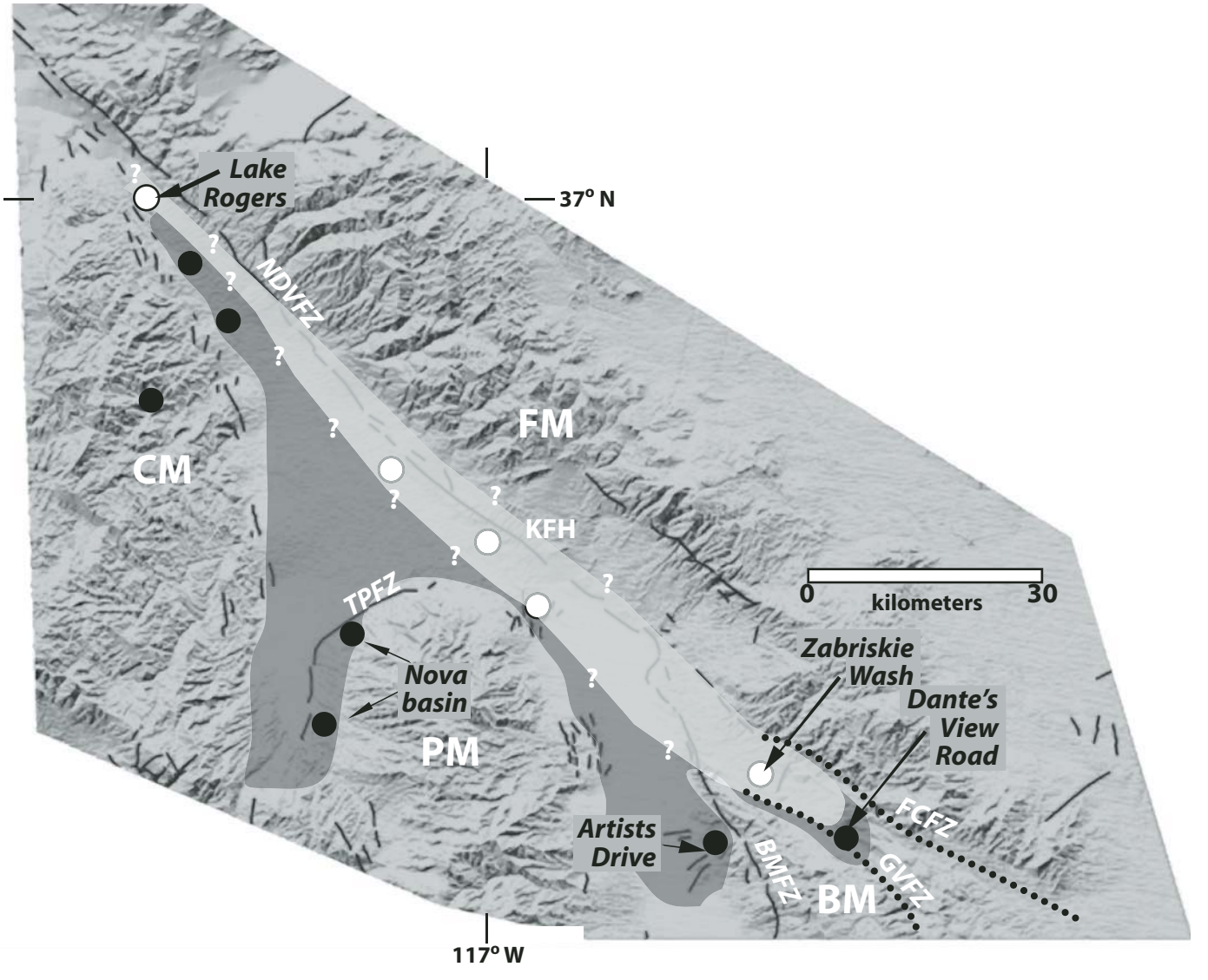

Figure 16. Estimated extent of late Pliocene (3–2 Ma) Furnace Creek Basin (dark shading) and Death Valley lake (light shading). Circles show known locations of 3.35–3.28 Ma tephra layers: black circles indicate alluvial-fan deposits; white circles are lacustrine beds. See Figure 2 for abbreviations of fault zones and geographic features.

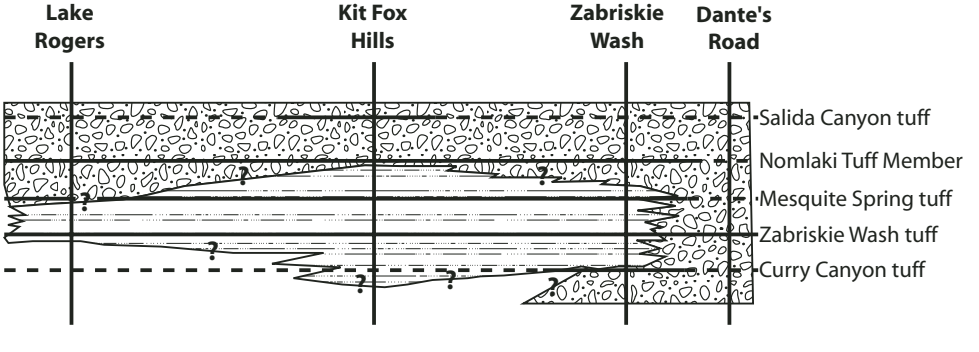

Figure 17. Schematic northwest-southeast–oriented cross section of the ancient Furnace Creek Basin showing key tephra beds. Solid lines indicate locations where tephra beds have been identified; dashed lines indicate that the tephra bed has not been identified at that particular location. Horizontal lines show the vertical and lateral extent of lacustrine deposits.

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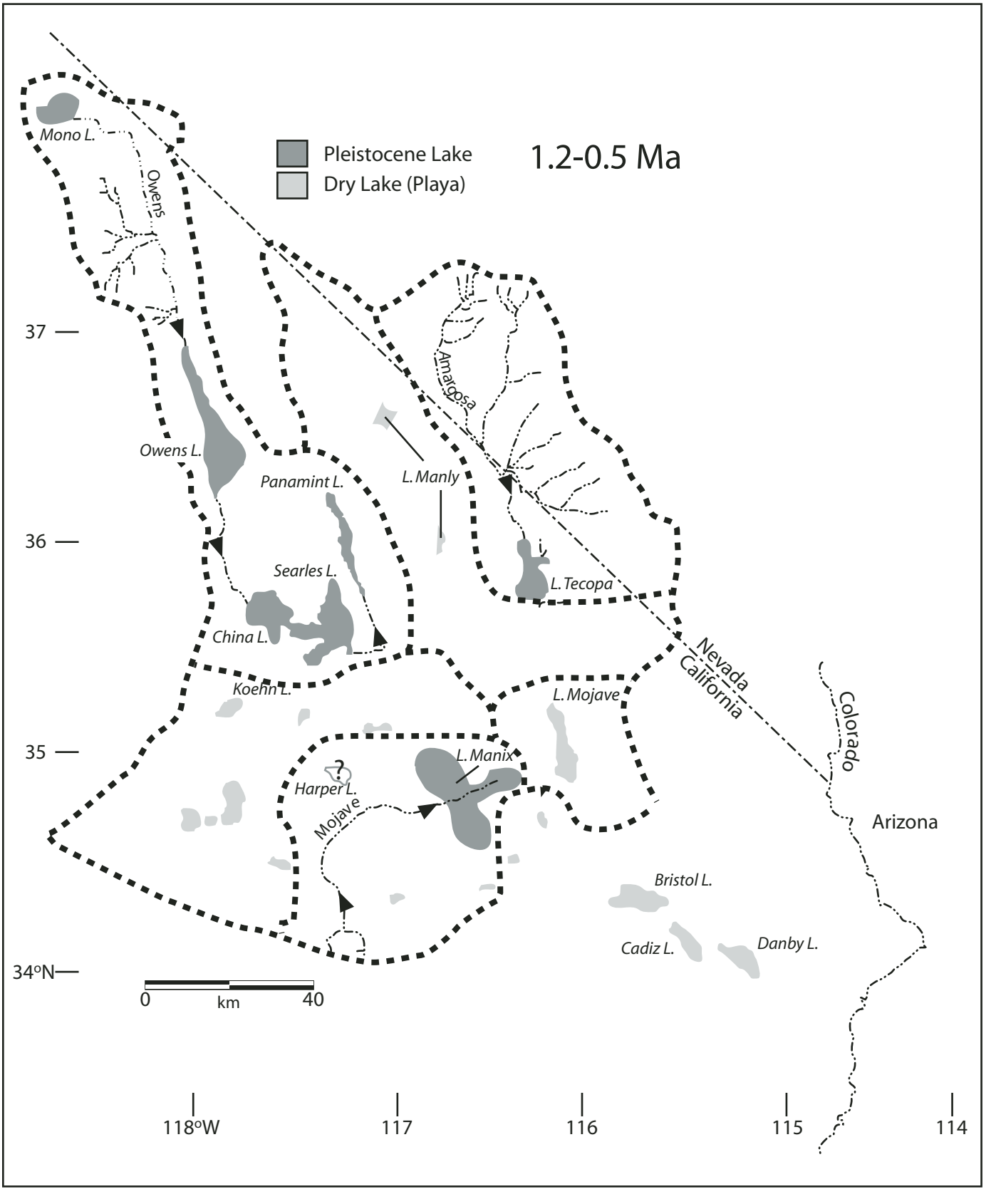

Figure 18. Early to middle Pleistocene (1.2–0.5 Ma) paleohydrology. The Mojave River is flowing north into Lake Manix by late in this time period (Cox et al., 2003; Jefferson, 2003). There is no connection between the Owens River and Death Valley, and only a small lake exists in Panamint Valley. Lake Mojave is not present either. The Amargosa River terminates at Lake Tecopa. Lake Manly is limited to a relatively small area in the lower parts of Death Valley.

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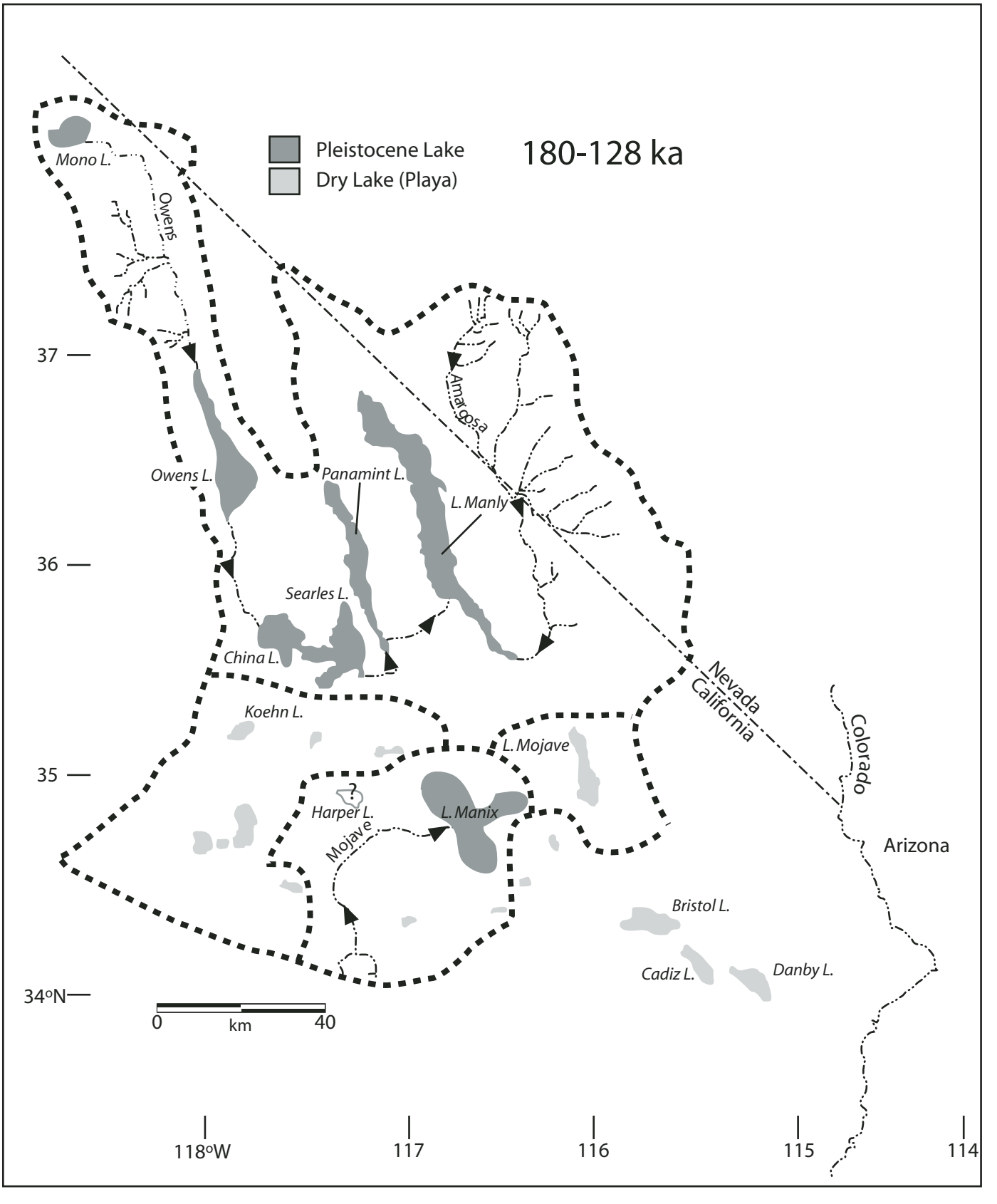

Figure 19. Middle Pleistocene (180–128 ka; oxygen isotope stage 6) paleohydrology. Mojave River terminates at Lake Manix, and pluvial Lake Mojave does not exist. The Amargosa and Owens Rivers reach Death Valley, forming a lake (Lake Manly) with a surface elevation of 90 m above present mean sea level. Lake Manly does not back up the proto–Mojave River. Lake Tecopa is no longer present, having been breached.

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et al., 1989; Wells et al., 2003). The cores from Silver Lake recovered by Wells et al. (2003) reached a maximum depth of 23.8 m or with a maximum age of 22 ka (^{14}C yr B.P.) based on sedimentation rates calculated from radiocarbon dating of the cores.

Although not integrated with the Mojave River system, the Soda Lake and Silver Lake Basins could have been hydrologically connected with Death Valley if Lake Manly rose to ~300 m above mean sea level. Geomorphic evidence led Hale (1985) to hypothesize that Lake Manly rose to an elevation of 595 m above mean sea level, encompassed the Soda Lake and Silver Lake Basins, and overflowed southward toward the Colorado River via Bristol Lake playa, possibly during MIS 6. Brown and Rosen's (1995) study of cores from the Bristol Lake playa, however, demonstrated that no substantial lake or fluvial system has existed in Bristol Lake playa since the late Pliocene.

More recently, Meek (1997) and Hooke (1999) hypothesized that a combined Pleistocene Lake Manly–Lake Mojave existed, but it did not spill into Bristol Lake playa. Hooke (1999) cited geomorphic evidence and coring data to support his hypothesis that a Lake Manly–Lake Mojave existed during MIS 6 (the Blackwelder stand). Hooke (1999) hypothesized that the orange-brown sediments at >100 m depth in the Soda Lake cores cited by Muessig et al. (1957) are evidence of the MIS 6 lake. Enzel et al. (2002) disputed this interpretation, based in part on the lack of age control below the lacustrine deposits at >30 m depth in the Silver Lake cores, and, as noted by Smith (1991), oxidized orange-brown sediments are inconsistent with lacustrine deposition.

A correlation of the Bishop and upper Glass Mountain ash beds at 124 and 182 m depth, respectively, in the Soda Lake core 1 shows that sediments at >100 m depth are early to early–middle Pleistocene in age and not correlative with the 180–128 ka MIS 6 Blackwelder stand. Assuming that the description and interpretation of the Soda Lake core by Muessig et al. (1957) are accurate, we infer that there was no lake in the Soda Lake Basin during MIS 6, nor during the preceding early to middle Pleistocene (MIS 20–16). The absence of a lake in the Soda Lake Basin during the early to middle Pleistocene, when many other basins in North America contained substantial lakes (Smith, 1991; Reheis et al., 2002a), likely reflects the relatively small drainage area of pluvial Lake Mojave at the time (Fig. 19). These data support the interpretation of Brown and Rosen (1995) that Lake Manly did not inundate pluvial Lake Mojave or the Mojave River and did not overflow into the Colorado River system during either MIS 20–16 or MIS 6.

SUMMARY AND CONCLUSIONS

The mtDNA data for morphologically similar pupfish inhabiting isolated streams and ponds of southeastern California and western Nevada, and in the fossil record, suggest a common late Tertiary ancestor and divergence of isolated populations beginning between 3 Ma and 2 Ma (Smith et al., 2002; Echelle et al., 2005; Echelle, this volume). Alluvial-fan deposits, which include a Mesquite Spring tuff on the southeast flank of the Furnace Creek Basin, show that the inferred Miocene-age river of Wright et al. (1999) that flowed southeast from the ancient Furnace Creek to the Amargosa Valley was disconnected by Pliocene time. Tephrochronologic data and lithologic facies show that a northwest-southeast–oriented lake existed in the ancient Furnace Creek Basin in middle Pliocene time, specifically from ca. 3.4 to <3.0 Ma. The lake receded and was sustained longest in the Kit Fox Hills area, near the central part of the ancient Furnace Creek Basin. Previous work by others (Hillhouse, 1987; Cox et al., 2003; Jefferson, 2003) has shown that from 3 to 2 Ma, the Amargosa and Mojave Rivers terminated upstream and not in Death Valley. A 3–2 Ma connection of the Owens River to Death Valley via Panamint Valley cannot be excluded by the geologic data.

The wetter/colder climates of the early to middle Pleistocene produced substantial high lake stands in the Tecopa, Owens, and Searles Basins (Morrison, 1991; Smith, 1984; Smith and Street-Perrott, 1983), and elsewhere in the Great Basin (Smith, 1991; Reheis et al., 2002a). Limited outcrop data from Death Valley show a perennial lake near Mormon Point older than 0.78 Ma and again at Mormon Point and Kit Fox Hills between 0.759 and 0.6 Ma. Carbonate facies at these sites suggest that the lake was likely saline and thus probably shallow. Data from upstream basins show that this early to middle Pleistocene lake in Death Valley did not receive inflow from the Owens (Larsen et al., 2003; Phillips, this volume), Amargosa (Hillhouse, 1987) or Mojave Rivers (Jefferson, 2003), but instead was fed by internal drainage waters (both springs and precipitation).

The identification of the Bishop and upper Glass Mountain ash beds in the Soda Lake Basin core provides age control beyond the radiocarbon-dated deposits at ~30 m depth. According to Muessig et al. (1957), there are no lacustrine deposits between the green clay at 30 m and the Bishop ash bed (0.759 Ma) at 128 m depth. These observations show that there was no lake in the Soda Lake Basin throughout the middle Pleistocene and thus that Lake Manly did not encompass Silver Lake and Soda Lake Basins during OIS 6 at 180–128 ka. Combined with the studies of Brown and Rosen (1995), there appears to be no evidence supporting a rise in Lake Manly to a height such that the lake would flow south through Bristol playa and connect with the Colorado River. Therefore, if pupfish dispersed from Death Valley into pluvial Lake Mojave, it was done by upstream migration during the latest Pleistocene.

In our estimation, the geologic data also highlight at least two possible untested hypotheses. One, pupfish may have dispersed from the Owens River to Death Valley via Panamint Valley. The paucity of data regarding the Pliocene (3–2 Ma) lakes of Panamint Valley keeps this hypothesis viable, although the receding lake in the ancient Furnace Creek Basin described herein does not support substantial overflow from Panamint Valley at that time. Assuming this Pliocene Owens River–Death Valley connection did exist, pupfish could have dispersed to the Mojave

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and Amargosa Rivers by upstream migration after ca. 180 ka and ca. 20 ka, respectively, when surface hydrologic connections began. Even with this hypothetical dispersal path, however, a physical connection to the Colorado River is not evident.

A second untested hypothesis is that the Owens, Amargosa, Mojave, and Colorado Rivers were connected earlier than 3–2 Ma. There is geologic evidence of a late Miocene (older than 11 Ma) fluvial system connecting northern Panamint Valley, Death Valley, and the Amargosa Valley that was defeated ca. 6 Ma (Wright et al., 1999; Niemi et al. 2001; Renik and Christie-Blick, 2004). Although the extent and termination of this Miocene fluvial system are unknown, this could be an unrecognized Miocene divergence route for pupfish. Further investigation of these hypotheses must also provide substantive evidence of a connection to the Colorado River, which is an unproven hypothesis in our minds at this time.

ACKNOWLEDGMENTS

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Appendix A3.3

Stop 1: Top of Tecopa Hump on China Ranch road. 1 hr. McMackin, Menges, Caskey, Knott

Tecopa Hump and Integration of Amargosa River through Amargosa Canyon (McMackin and Menges)

This stop provides a spectacular view of the highland area that forms the topographic and structural boundary of the southern and southwestern Tecopa basin behind us and the northeastern and northern margin of the southern Death Valley and Valjean Valleys (visible in distance through Amargosa Canyon). The skyline ridgecrest is the formerly continuous paleodivide between these two basins. This divide is located on a topographically high structural culmination of the Tecopa Hump, an antiform that approximately continues 15 km in both directions from this site along a WNW-ESE trending axis with an asymmetric cross-profile much higher and steeper on the southwest (southern Death Valley) side.

In contrast to a classical anticline, the Tecopa Hump has a core of complexly faulted bedrock overlain by Late Miocene Basin assemblage, including the China Ranch beds (light colored sediments in middle and foreground) and extensive fanglomerates shed from adjacent ranges or uplifted bedrock in the Hump. Late Tertiary and older Quaternary fanglomerates dip as much as 10o to 15o north or south on the flanks of the Hump. We consider the Tecopa Hump to have not only played a major role in the original development of the boundary and paleodivide of the Tecopa basin, but also exposed and structurally inverted the core of an even older late Miocene sedimentary basin. The Hump also appears likely to have influenced the location and timing of the initial breach through the paleodivide that first integrated the Amargosa River between Tecopa basin and southern Death Valley. Both flanks of the Hump are extensively dissected by persistent and locally extensive incision (obvious in this area) which both predates and postdates initial integration of the Amargosa River in late-middle Pleistocene time. Persistent dissection into Miocene to Proterozoic basement rocks in this section of the Hump highland is especially well recorded by an inset or stair-stepped suite of at least three major and up to five or six detectable levels of strath or strath-fill terraces on the main-stem and tributary drainages (including Willow Creek) of the Amargosa River. We propose that the topographic form and geomorphic characteristics of the Hump, when combined with incision history (ie., spatial patterns, timing, and divergence-convergence height variations of strath terraces) suggests a continued low-level of uplift on this structural boundary of southern Death Valley throughout at least mid-late Quaternary time.

Appendix A3.4

NEW SHORELINE AGES AND OIS-6 LAKE-LEVEL HISTORY OF LAKE MANLY, DEATH VALLEY

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Despite a long history of Lake Manly research, a well-dated core record of the last two major lake cycles [186-120 ka (OIS-6) and 35-10 ka (OIS-2)], and numerous attempts to date shoreline deposits, the pluvial history as recorded in the ages and elevations of Lake Manly shorelines has remained controversial and unresolved. To this end, we initiated a basin-wide survey of high-level shorelines in areas of relative tectonic stability. We see remarkable consistency in elevations of two distinct high-level shorelines spanning northern, central (west side), and southern Death Valley. The highest formed at ~60 m above sea level (masl). Slightly lower, and notably prominent shorelines mark lake levels at ~46 masl. Similarities in degree of preservation suggest shorelines at both levels formed during the same pluvial cycle. New U-series tufa ages from 46-m shorelines on Johnson Canyon fan and a basalt hill (BH) on the north side of Blackwater fan (BF) are best bracketed between 129.5 ± 1.6 ka and 122.4 ± 1.1 ka, and indicate a high lake level late in the OIS-6 pluvial. Further details of OIS-6 lake fluctuations are recorded on BF where a high beach ridge (57 m) built onto the south flank of the BH during (what we now take as) the OIS-6 high stand. Following lake regression from the high stand, the 57-m beach ridge was deeply incised on the flank of the BH and inset by a younger fan. The timing of lake regression isn't known. Remarkably, however, the fan that was inset during lake regression records a later transgression of Lake Manly in the way of shorelines and overlying beach deposits that can be traced on the fan to a maximum elevation of 46 m. Importantly, the relations show that the late OIS-6 lake at 46 masl was a late-transgressional phase, not a recessional lake level and thus implies an earlier OIS-6 age for the high stand. Our new observations and age constraints agree well with the lacustrine record from CORE DV93-1. We interpret that: 1) the 60 m shorelines mark the original, quasi-stable, highstand of Lake Manly reached in the earlier part of the OIS-6 pluvial (186-166 ka interval); 2) lake regression (recorded at BF) likely occurred during a time of higher lake salinities (166-144 ka interval); and 3) late OIS-6 transgression to 46 masl matches a deep lake interval at 122 ka, but perhaps more likely correlates to the deep lake interval ending at 128 ka.

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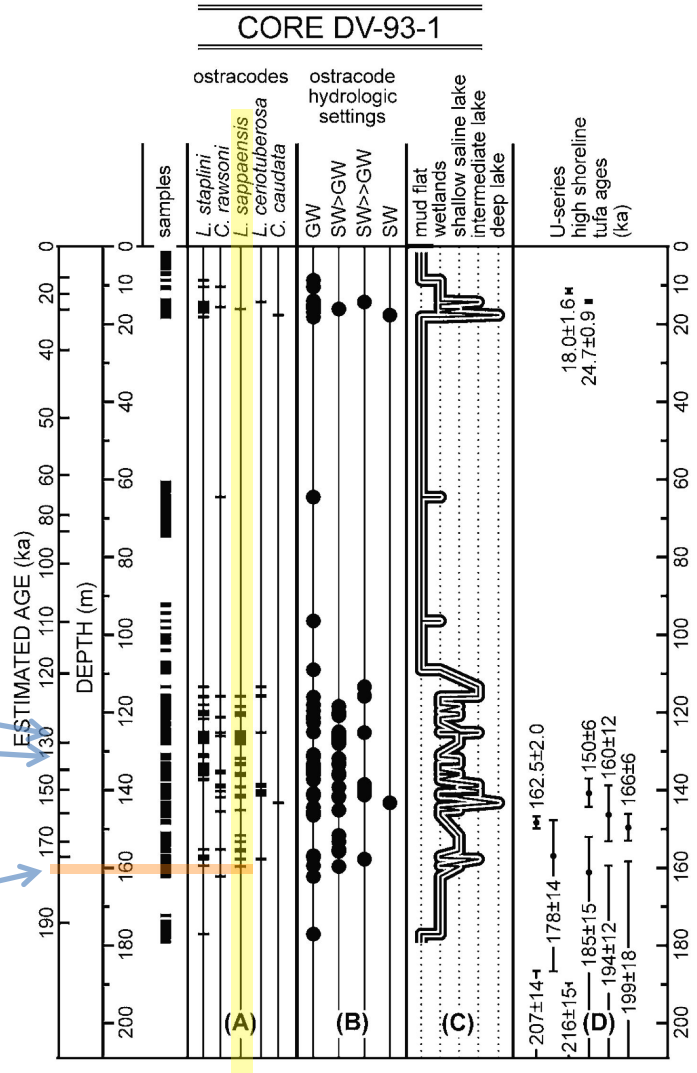
The interval from ~133 m (137 ka) to 127.5 m (ca. 129 ka) is dominated by *Limnocythere sappaensis,* implying that a stream-supported, alkaline-saline lake persisted through this stratigraphic interval. Streamflow implies a higher level of effective moisture that was due probably to wetter conditions. *Limnocythere staplini* and other ostracodes are also present, but they occur less commonly, indicating that marginal lacustrine environments are present but not common.

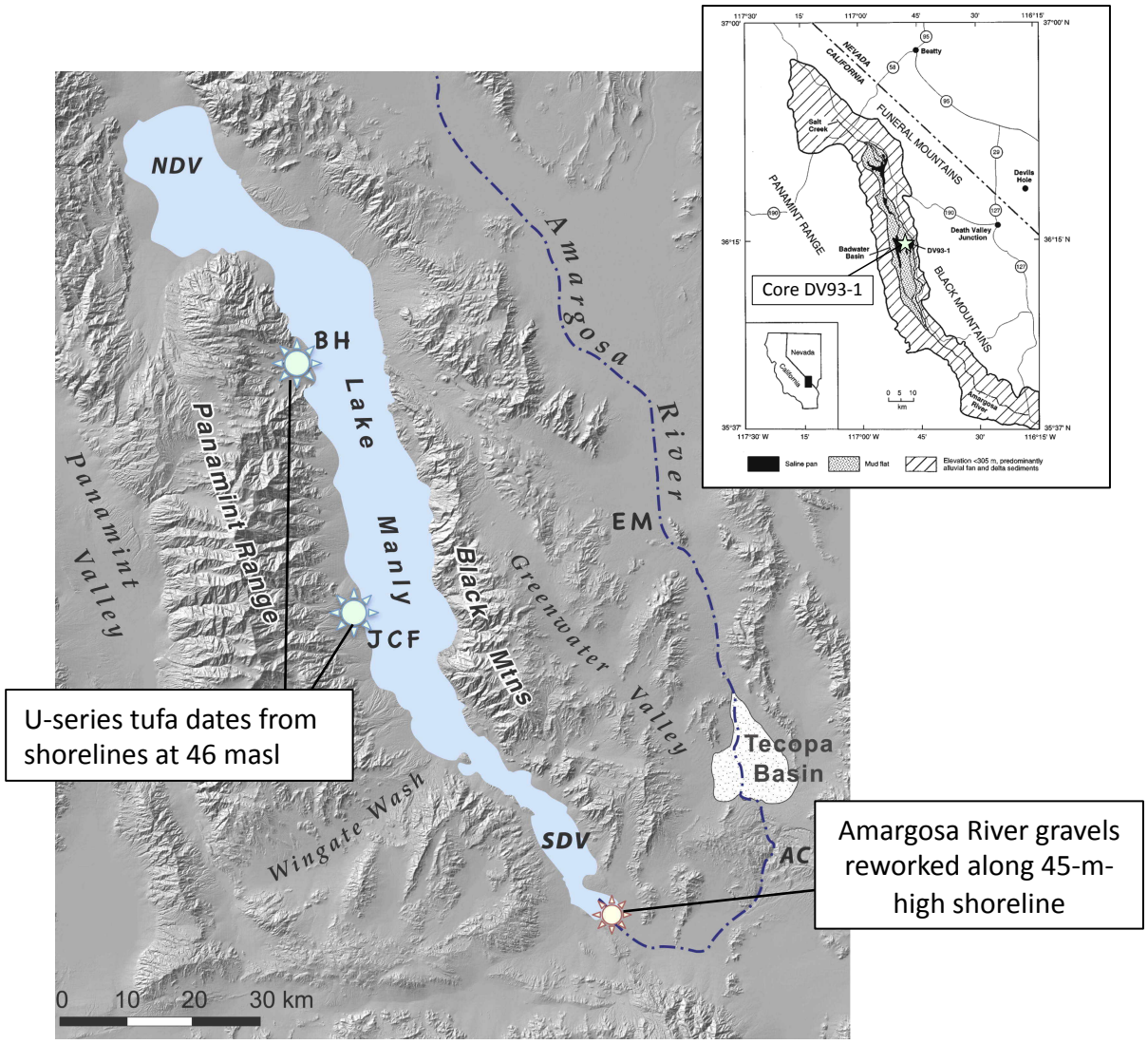
Significantly, the next ostracode to appear in the section is *Limnocythere sappaensis* at 159.6 m (184 ka; Fig. 4). Its presence implies water with a high alk/Ca solute value likely derived primarily from Amargosa streamflow. The timing of the shift from *C. rawsoni* to *L. sappaensis* suggests that the chemical transition from low to high alk/Ca water took >2000 yr, assuming a constant rate of sediment accumulation, although the poor preservation of ostracodes creates uncertainty in the timing of the chemical change. Nonetheless, the appearance of *L. sappaensis* indicates substantial stream inflow into the basin, thus initiating the lakes and wetlands that characterize most of MIS 6.



From Forester et al. (2005)

Limnocythere sappaensis
184 ka

Forester et al. makes a strong copmment in his 2005 paper regarding the first appearance of Lsapp in the core



U-series dating results; Denver Radiogenic Isotope Lab
June 28, 2010
Mojave Paleohydrology Project

Sample ID	Sample type	[U], µg/g	[Th], µg/g	$^{230}Th/^{232}Th$ AR[a]	$^{230}Th/U$ Age (ka)	±2s	Initial $^{234}U/^{238}U$ AR	±2s
PNT10-3B1a	Layered tufa rind	18.9	0.14	414	134.1	1.2	1.519	0.004
PNT10-3B1b	**Layered tufa rind**	**25.9**	**0.30**	**249**	**124.7**	**1.2**	**1.509**	**0.004**
PNT10-3B1c	Layered tufa rind	50.4	0.10	1612	142.9	1.4	1.559	0.004
PNT10-3B2a	Layered tufa rind	30.6	0.12	825	145.0	1.4	1.550	0.004
PNT10-3B2b	**Layered tufa rind**	**65.5**	**0.04**	**5284**	**122.4**	**1.1**	**1.509**	**0.004**
JCF-6a	**Charophytic tufa**	**80.8**	**0.02**	**11740**	**129.5**	**1.6**	**1.500**	**0.004**
JCF-6b	Charophytic tufa	77.8	0.01	21064	139.0	1.5	1.517	0.004
LM-UTh-1a-1	Carbonate clast rind	24.8	0.14	563	225	9	1.248	0.007
LM-UTh-1b-1	Carbonate clast rind	17.7	0.91	83	Post-depositional U mobility			
Mnx_lower B	Ostracod shells	7.5	2.09	12	488	145	1.46	0.17

Values shown in boled are least likely to have been impacted by post-depositional U mobility

Lakes of the Western Great Basin: Waucobi Lake Beds

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Ancient Waucobi

The Waucobi Lake Beds located near Big Pine, CA record a major change from a saline-alkaline to fresh water facies (Hay, 1964; Bachman, 1974), while other eastern Sierra Nevada lake systems underwent intense desiccation (Smith, 1984; Bradbury, 1992; Reheis et al., 2002). Recent work on the Waucobi Lake Beds, focusing on the interplay of climate and tectonics at the Plio-Pleistocene boundary, has been completed at Cal State Long Beach with assistance from the Berkeley Geochronology Center. We completed $^{40}Ar/^{39}Ar$ dating on tuffs located throughout the lake beds (Figure 1, De Masi, 2013), along with exhaustive mineralogical analysis which helps to confirm and refine stratigraphic interpretations by previous authors. Our hydroclimate from ancient Waucobi was compared to other lake systems in the Owens Valley region (Figure 2) to aid in determining if the facies change recorded in the Waucobi Lake Beds at ~2.5-2.4 Ma followed regional climate change or if the facies change was related to a hydrologic shift influenced by tectonic activity.

Radiometric dates in the Waucobi Lake Beds span 2.63 to 2.05 Ma. Samples older than 2.5 Ma are comprised mostly of phillipsite, and contain evidence of evaporite deposits, as noted by Hay (1964). The presence of phillipsite requires alkaline waters and the presence of evaporites implies a shallow intermittent lake, suggesting the lake had higher salinity as well as higher alkalinity before 2.55 Ma. Then from 2.55 to 2.35 Ma, the clays indicate a transition from alkaline-lower moisture conditions to more freshwater conditions. Montmorillonite becomes increasingly abundant in tuffs dated 2.35 Ma and younger. The transition from phillipsite to montmorillonite shows environmental change opposite of global and regional climate patterns during the late Pliocene and early Pleistocene. Based on clay mineralogy, the Waucobi Lake Beds record a relatively dry period from 2.63 Ma until 2.55 Ma. Between 2.35 to 2.05 Ma the mineralogy of the lake beds continue to shift to mostly montmorillonite, continuing to indicate presence of fresher, possibly deeper water.

Comparing paleoclimate indicators from Waucobi to other lacustrine systems in the region shows that Waucobi was moistening, although many climate records demonstrate a decrease in effective moisture across the Plio-Pleistocene boundary. Mineralogy from both Tule Lake and Searles Lake, which form a north-south transect east of the Sierra Nevada, suggest a dryer climate during the latest Pliocene. Searles Lake is interpreted as experiencing increasing dryness at 2.6 Ma followed by complete desiccation from 2.50 to 2.0 Ma (Smith, 1984), whereas Waucobi appears to have a stable amount of fresh water. Tule Lake is also interpreted as drying from 2.65 to 2.35 Ma (Bradbury, 1992), similar to that of Searles Lake.

Tectonics and Regional Hydrology

It has been suggested that several lakes in the Western Great Basin were hydrologically connected during the Pliocene and Pleistocene (Phillips 2008; Reheis et al., 2002). Lakes on the eastern side of the Sierra Nevada surrounding the Owens Valley region were possibly linked through the paleo-Owens River (Figure 3), including Waucobi and Searles lakes. Re-routing of the paleo-Owens River could feed one lake while depriving another. Co-seismically deformed sediments in Waucobi, constrained radiometrically to have occurred between 2.5-2.4 Ma, indicate tectonics were important in controlling the geometry of the Waucoba Embayment. This seismite layer (Figure 4) is contemporaneous with the increase of moisture recorded in Waucobi sediments, and the desiccation of several local basins.

Continued extension near the Waucoba Embayment during the late Pliocene and early Pleistocene, would likely have allowed for a longer residence time for water within ancient Waucobi. The lacustrine record clearly demonstrate the presence of water for at least 1 Ma, implying that active faulting provided sufficient subsidence to keep the basin in existence.

Ancient Owens River was probably a major source of water input for Waucobi Lake. Phillips (2008) postulates that one possible path of the Owens River flowed through the Waucoba Embayment and eastward over into Saline Valley (Figure 3). Evidence supporting this idea was found within the lake beds by Bachman (1974) and De Masi (2013), in the form of interference-patterned water ripples preserved in carbonate sediment underlying ~2.5-2.35 Ma tuffs. The interference patterned water ripples imply a change in current direction, from a south-easterly flow to a north-westerly flow. This north-western flow is captured as current ripples in the overlying layer. It is possible that an outflow of the Owens River went through the embayment and through a gap in the proto-White-Inyo Mountains during this time period. A change in the direction of that outflow would have deprived Searles Lake from river input, hastening its desiccation while accounting for the freshwater of Waucobi Lake.

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Figures

Figure 1. Sediments in Duchess Canyon, Waucobi Lake Beds. Dated tuff layers shown in black dashed lines.

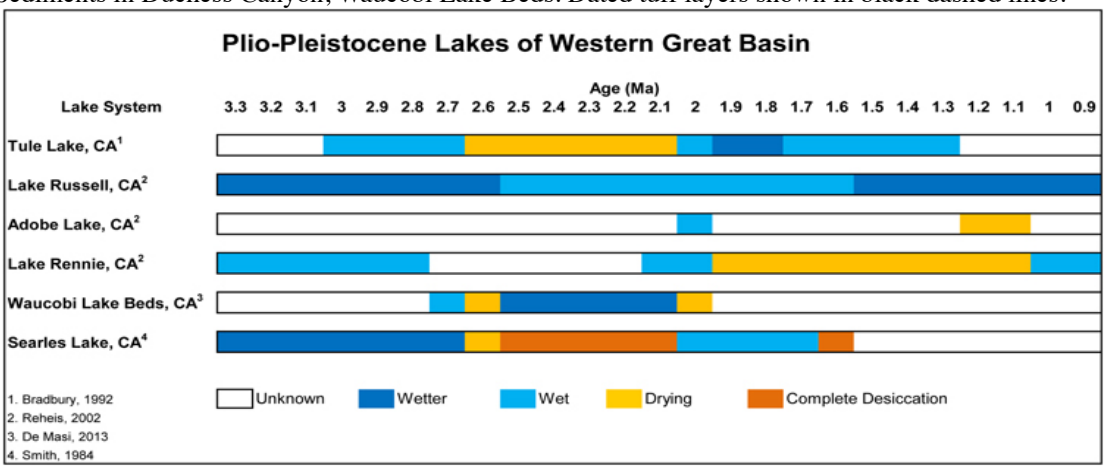


Figure 2. Environments of Plio-Pleistocene lakes in the Western Great Basin

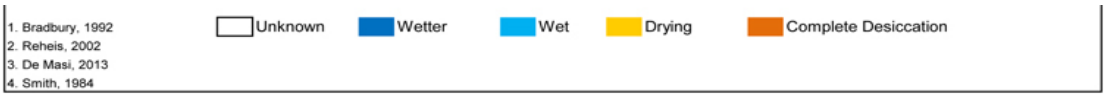


Figure 2. Environments of Plio-Pleistocene lakes in the Western Great Basin.

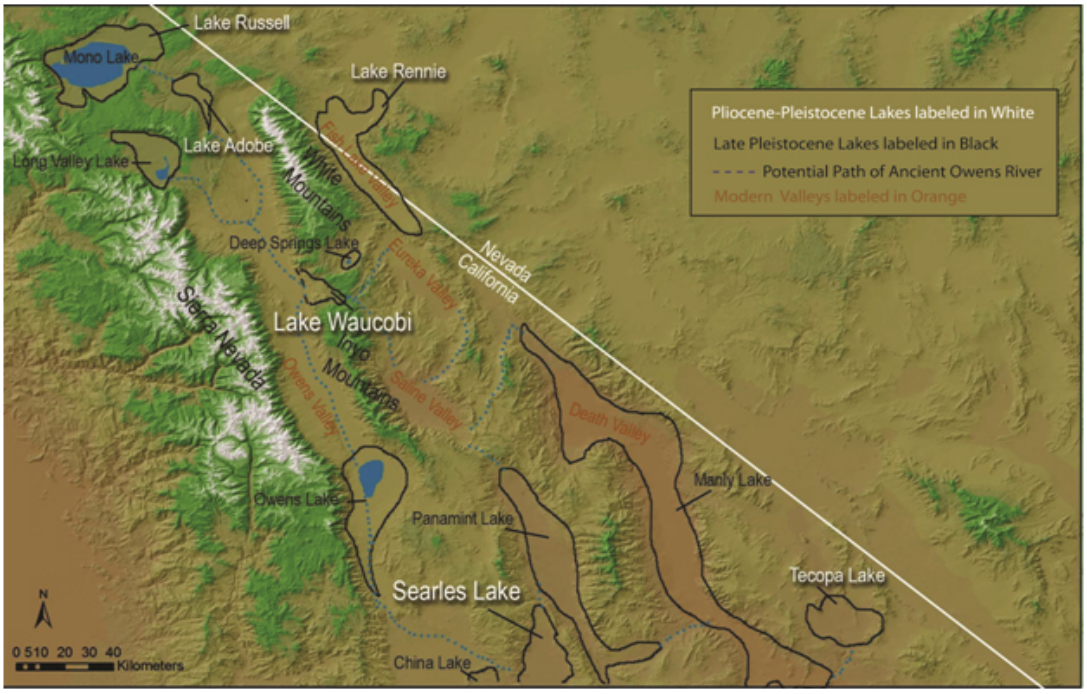


Figure 3. Map of paleo-Owens River with regional lake systems.

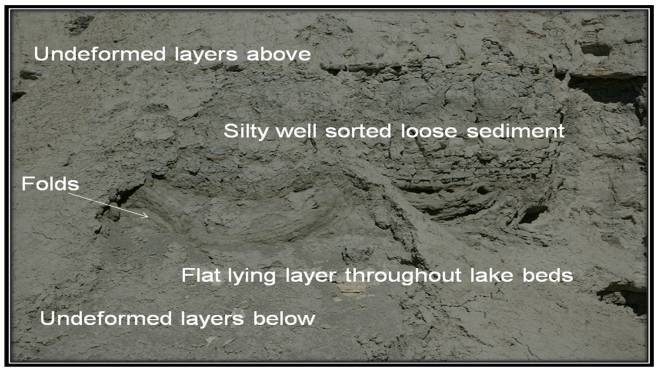


Figure 4. Co-seismically deformed sediments in Duchess Canyon.

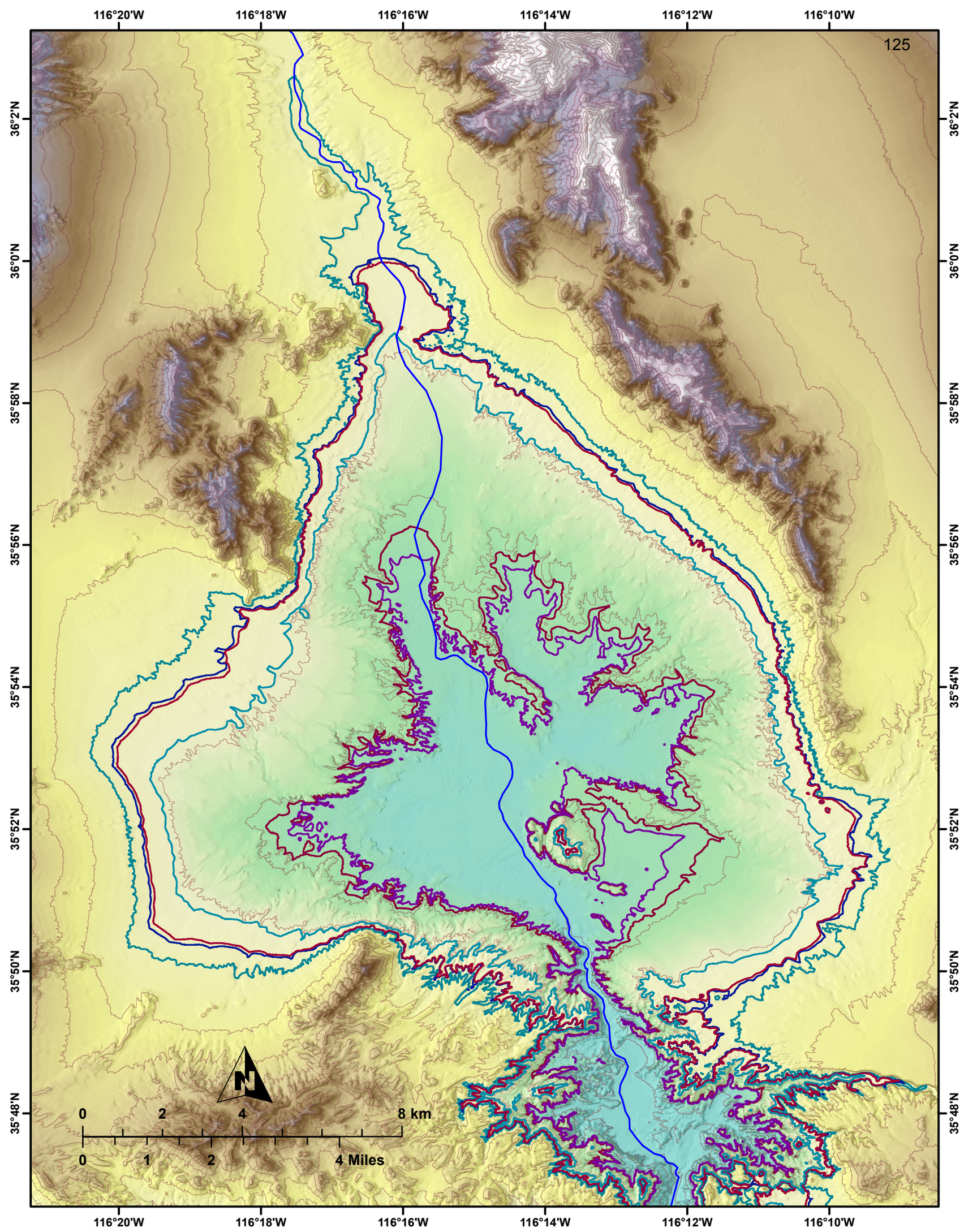
Appendix A3.6

Eagle Mountain Narrows (Menges)

Stop 3.3- Optional (Menges) At this stop we will examine stratigraphic and geomorphic evidence for a relatively recent late Pleistocene integration event in this central reach of the Amargosa River. This event linked a very low-relief, unincised, fine-grained playa margin reach of the river in the southern Amargosa Dester to the north (north of Eagle Mountain on the right), with a steeper, gravel-floored incised reach that continues into Tecopa Basin (below and to the south). These two subbasins were separated fluvially from one another by a formerly continuous, mixed alluvial and bedrock paleodivide until integrated by establishment of the present continuous mainstem channel of Amargosa River within the 5-km-long narrow confined canyon at this stop. The exact mechanism for breaching the divide likely included overtopping of shallow lake in the playa on the north and (or) groundwater sapping from above, with no evidence for any active tectonic factors in this area. This integration linked two formally isolated subbasins at different local altitudes, which imposed a sudden local base-level fall across the newly integrated reach that induced a discrete pulse of incision along the main channel and adjacent tributary drainages. The process and timing of this integration and subsequent canyon down-cutting is recorded in a complex stair-stepped suite of 4 major inset river terraces and (or) laterally continuous tributary terraces and fans on either side of the axial canyon reach. These terraces dramatically decrease or converge in height upstream and downstream from a maximum incision depth of 30-m where the channel crosses the location of the integration breach in the former paleodivide.

Four Cl^{36} cosmogenic ages (depth profiles), one from a pre-dissection section of paleodivide near the canyon and three inset terraces in the canyon or accordant tributary fans provide age control for the initial integration event and subsequent fluvial dissection. These ages crudely bracket the integration event between 210ka (age of pre-breach paleodivide) and the 50-45 ka age of the highest dated terrace level . However, the ages of both high-level strath-terraces are essentially overlapping between 50 and 40 Ka, whereas the age of a low terrace, only 2 m above the present channel, is 32 ka. These relations further suggest that (a) initial drainage integration through the paleodivide probably occurred no earlier than approximately 50 ka, and (b) a subsequent very nonlinear, single major pulse of canyon incision occurred within a primarily 20 ka interval (between 50 and 30 ka), with over 80-85% of the incision constrained to the first 10 ka. The heights of the 30 ka and all the younger fill terraces are 2 m or less above and subparallel to the modern channel, suggesting establishment of an essentially at-grade reach of the modern channel through the canyon by 32 ka. Thus the essentially modern channel profile followed the earlier period of rapid dissection of the channel that developed as a fluvial adjustment to the initial integration event . The subsequent post 30-ka fluvial history

in this canyon reach consists primarily of three climate-controlled latest Pleistocene to late Holocene fill terraces superimposed on 1-2 m of net channel incision.


116°20'W
116°18'W
116°16'W
116°14'W
116°12'W
116°10'W
36°2'N
36°0'N
35°58'N
35°56'N
35°54'N
35°52'N
35°50'N
35°48'N
N
0
2
4
8 km
0
1
2
4 Miles