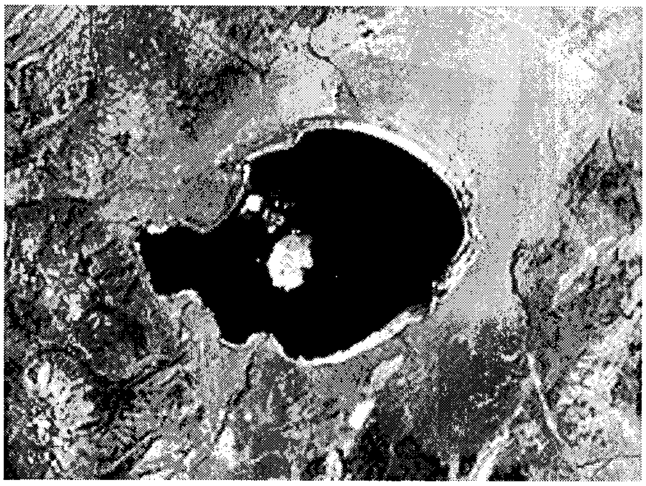
Third International Conference on Early Mars

Mono Basin Fieldtrip

Wednesday May 23, 2012



Field Trip Guide and Road Log

Field Trip Leader:

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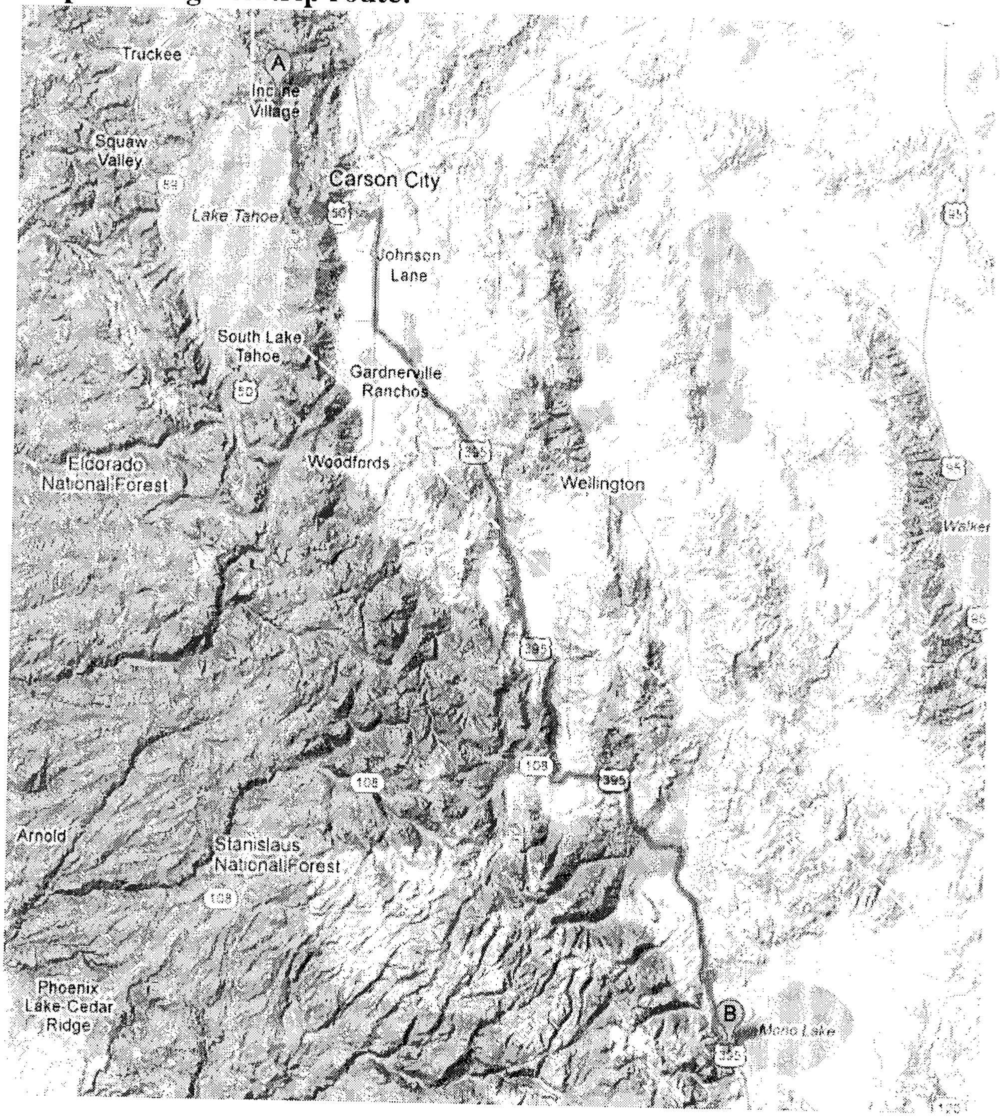
roles played by microorganisms in tufa genesis, particularly with regard to mineralogy and microtexture, biosignature preservation and the impacts of diagenesis.

Map showing fieldtrip route:

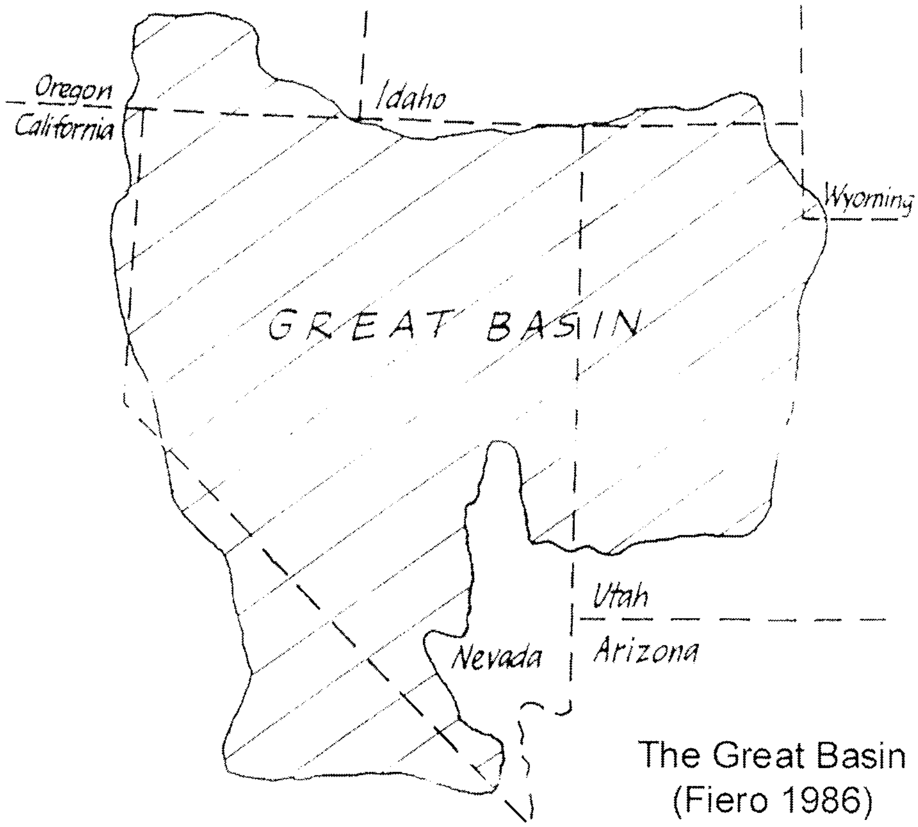


To place the field trip in a broader context, in the following sections, I provide some general geological background to facilitate discussion.

WHAT IS THE GREAT BASIN?

The Great Basin a major geomorphic province in the western United States that is bounded on west by the Sierra Nevada Mountains (highest range in the Continental US), on the east by the Wasatch Range and high plateaus of central and southern Utah, to the north by the Snake River Plain of southeastern Oregon, southern Idado/southwestern Wyoming (track of the Yellowstone hot-spot), and to the south (less distinctly) by the Colorado River and its northern tributaries (Figure 1).

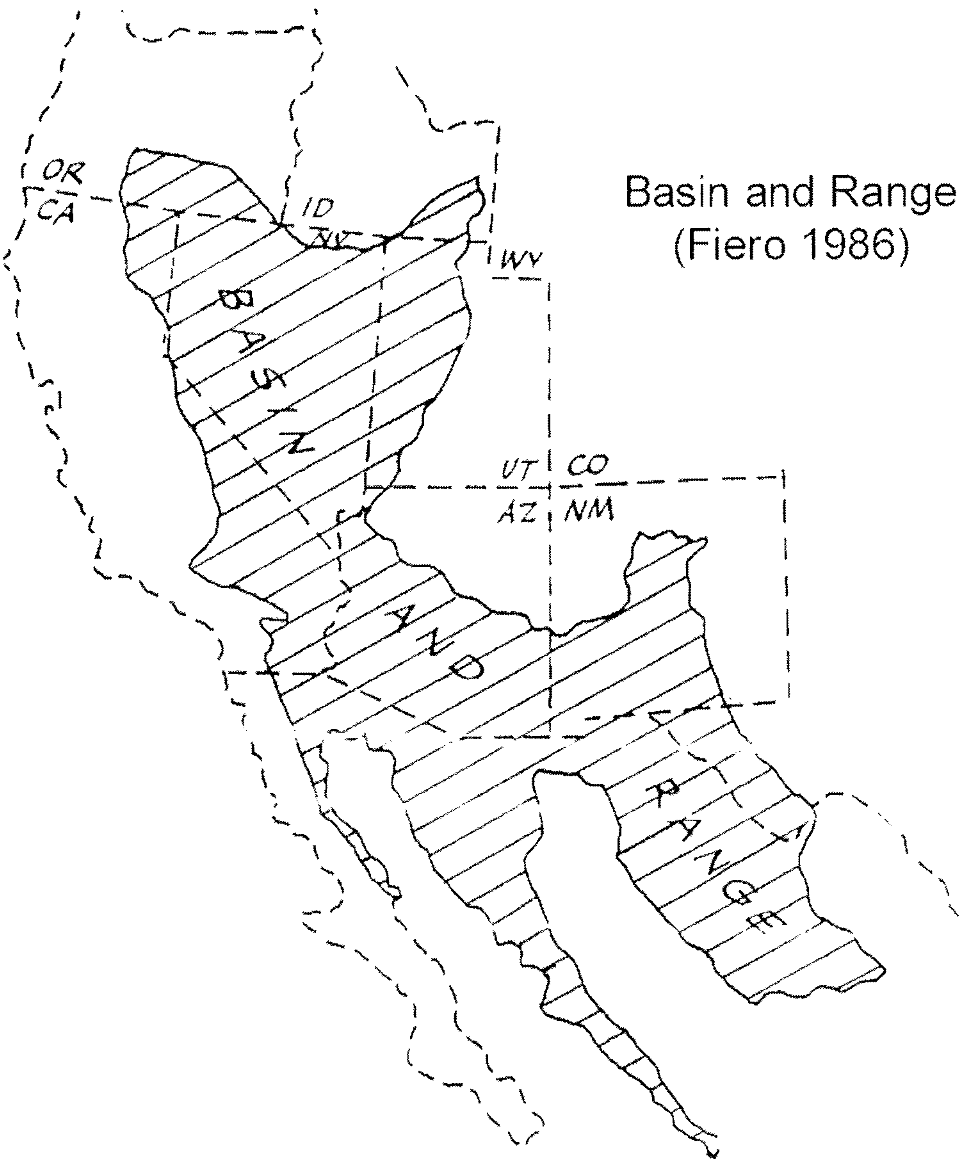
Figure 1: Extent of the Great Basin Physiological Province of Western North America.



WHAT IS THE BASIN AND RANGE?

The Great Basin includes portions of the so-called "Basin and Range" geological/structural province of western North America. The Basin and Range extends east through Arizona, New Mexico and Texas, and south into Mexico (Figure 2). The Basin and Range is characterized by numerous north-south trending fault blocks that form elongate, fault-bounded valleys ("grabens"), separated by elongate mountain ranges ("horsts"). The valleys of the Great Basin generally lie at elevations between 4,000 and 5,000 feet above sea level, with the highest mountains rising to >14,000 feet. As an extreme, Death Valley lies 282 feet below sea level, the lowest point on north American continent and is bounded to the west by the Panamint Range, which rises to >11,000 feet. Topographically, the Great Basin resembles an "upside-down, broken and cracked bowl" (Fiero 1986: 6) because the valley floors of the central region lie at higher elevations than along its flanks.

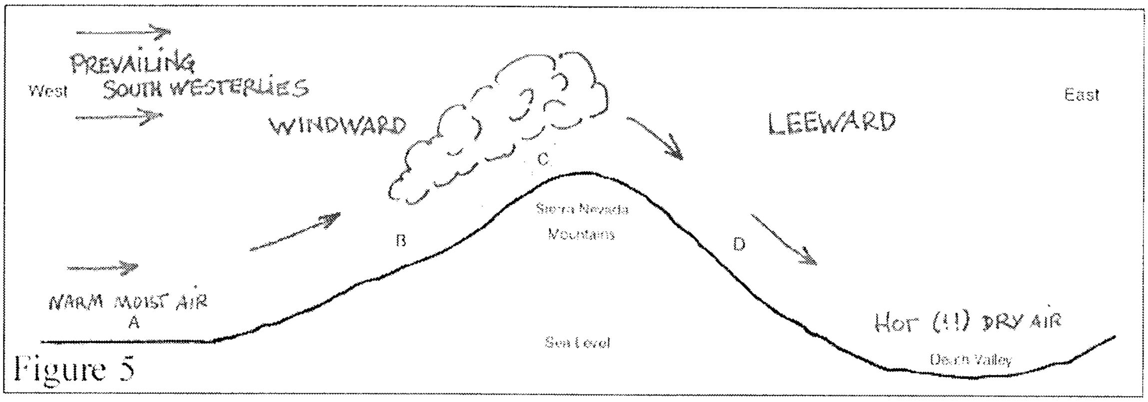
Figure 2: Extent of the Basin and Range Geological Province of Western North America.



GREAT BASIN CLIMATE AND HYDROLOGY

The Great Basin could also be called the "Great Desert". It is in fact the driest region in the United States. The Sierra Nevada Mountains and their northern extension, the Klamath Range, presently form a barrier to the moisture-laden air that moves onto the continent from the North Pacific (Figure 3). The Sierra Nevada was uplifted to its present height in the mid-Pleistocene, contributing to a post-glacial drying trend that has persisted into the present. The resulting orographic, or "rain shadow" effect (Figure 3), in combination with the geomorphic system of isolated, elongate valleys and mountain ranges, is responsible for unique hydrology of the Great Basin, characterized by numerous terminal lake basins (many with "playas" or dry lake beds), and ephemeral fluvial systems and their associated alluvial fans.

Figure 3: Orographic effects of large mountain ranges create rain shadow deserts.

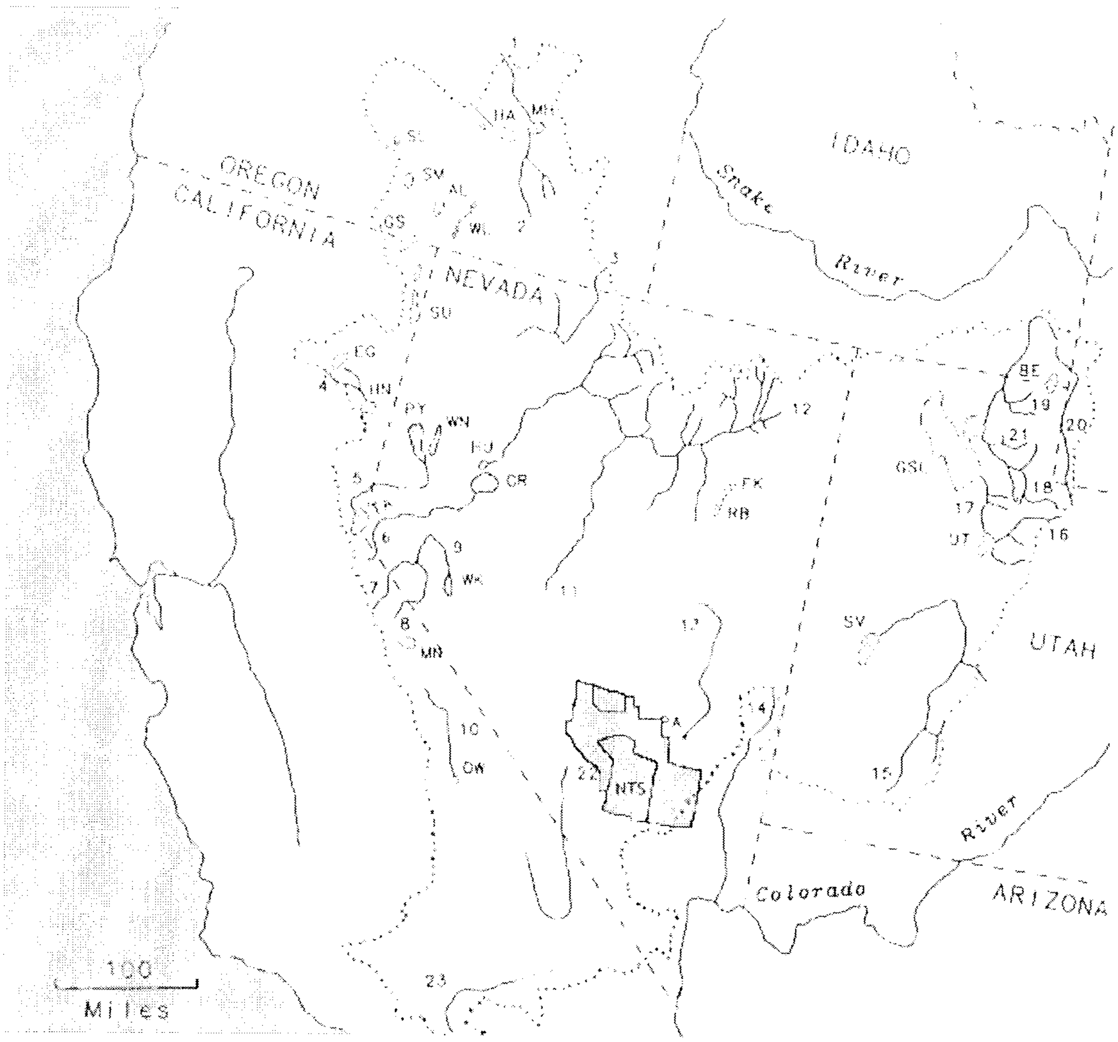


Fiero (1986: 169) noted that after a rain, the extremely flat valley floors of many playas become covered with a few inches of water, sometimes covering areas of tens to hundreds of square miles. On average, however, less than 3 inches of rainfall reaches the valley floors of the Great Basin in a single year. At higher elevations, up to ten times that amount can fall, and the tallest peaks can be blanketed with snow for 6 months out of the year. However, most of the precipitation at higher elevations is eventually lost through evaporation, or by percolation into the porous surfaces of alluvial fans.

Internal drainage of the Great Basin ensures that most of the surface water leaves by evaporation. The major rivers and lakes of this region are shown in (Figure 4); those along the field trip route are identified by number in Figure 6 as follows: Truckee River (5); Carson R. (6); Walker R. (7-9); and Humboldt R. (11, 12). The point here is that all of these rivers essentially begin and end within the Great Basin. But the picture at the surface is deceptive. As noted above, water that does not evaporate, infiltrates into the subsurface. Not surprisingly, many of the coarser sedimentary deposits along the margins of intermontane basins are also saturated subsurface aquifers and that water can come to the surface basinward, as springs.

Where heat flow is high (e.g. near shallow igneous intrusives), groundwater frequently resurfaces as thermal springs. Hot springs and seeps are common features of the intermontane basins of the western U.S., particularly in association with basin margin bounding faults. In fact, there are literally hundreds of active hot springs in Nevada alone. Active examples along our fieldtrip route include large travertine hot springs at Bridgeport; along the southeastern edge of Paoha Island in Mono Lake; silica springs at Steamboat Springs, just south of Reno (NV); and siliceous hot springs in Hot Creek, located in Long Valley caldera, south of the Mono Basin. Mineralizing hot springs are characterized by distinctive sedimentological features that make them easy to identify in ancient deposits. They also provide excellent targets for a microbial fossil record (Walter and Des Marais 1993; Farmer and Des Marais 1999).

Figure 4: Major rivers and lakes of the Great Basin.



COMMON SEDIMENTARY FACIES OF INTERMONTANE BASINS

Perhaps the most conspicuous features of intermontane basins in arid climatic settings are basinward-dipping alluvial fan deposits. (Most are familiar with the spectacular alluvial fans of Death Valley). Alluvial fans typically intergrade along basin margins with lacustrine fan-delta deposits (we will see good examples in the Mono Basin). Rises in lake level can result in fine-grained (impermeable) deeper water shale being laid down above porous and permeable, fan delta sands and conglomerates deposited nearer to shore. Fine-grained sedimentary deposits form aquicludes that retard the upward escape of water from subsurface aquifers, creating pressurized artesian systems. In Mono Lake, artesian spring flows are common on the floor of the lake today, where they are associated with the precipitation of carbonate minerals. The commonly proposed mechanism is disequilibrium mixing of calcium-rich groundwater and bicarbonate-rich alkaline-saline lake water (Ford 1989; Figure 6 from Council and Bennet 1993), although recent studies (Thomas and Farmer 2012) suggest important, if not dominant roles for microbial biofims.

In alkaline lake environments, freshwater springs and seeps are often focal points for the deposition of "tufa" (Italian; from Latin, tofus meaning porous rock), defined as a limestone deposited in continental settings at low temperatures (Bischoff et al. 1993b). In the case of lake floor springs, where flows are concentrated at a central vent, tufa towers form (Figures 5 and 6). Where spring flows are more diffuse, precipitation is manifested as pervasive cementation of lake floor sediments (by carbonate minerals). Examples of both types of deposits may be seen today at South Tufa Preserve and nearby Navy Beach (Figure 5). Additional information about tufa deposition is included with the road log entry for Stop 3 (South Tufa Preserve/Navy Beach).

Two things control the distribution of tufa deposits in Mono Lake: 1) The discharge rate of springs, which tends to be highest along the western margin of the basin, adjacent to the primary source of surface run-off (rainfall and snow melt from the Sierra Nevada Mts.); and 2) Places where pressurized aquifers at depth are intersected by faults and come to the surface (Council and Bennet 1993; Figure 6).

Figure 5: Tufa deposits exposed along the southern shoreline of Mono Lake. The left image shows tufa towers formed by deposition around lake floor springs (when the lake level was higher). The right image shows carbonate-cemented lake sediments on a wave eroded high stand lake terrace at Navy Beach.

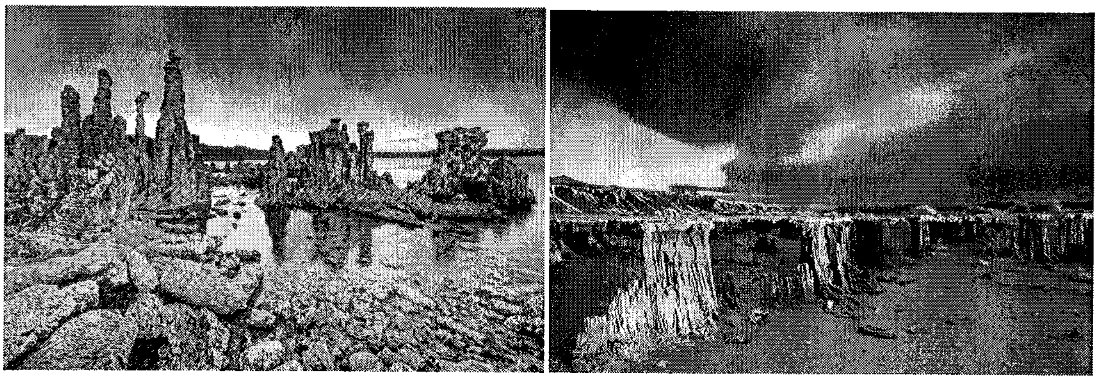


Figure 6: Model for tufa deposition by Council and Bennet (1993). Subsurface artesian aquifers develop in low stand fan delta sands, where they are overlain by an impermeable, high stand (deep basin) shale's. Where the lake floor sedimentary sequence is intersected by faults, pressurized groundwater can move upward along faults and emanate as springs on the lake floor. This process only occurs on the lake floor, and ceases when tufa deposits are exposed at the surface by a drop in lake level.

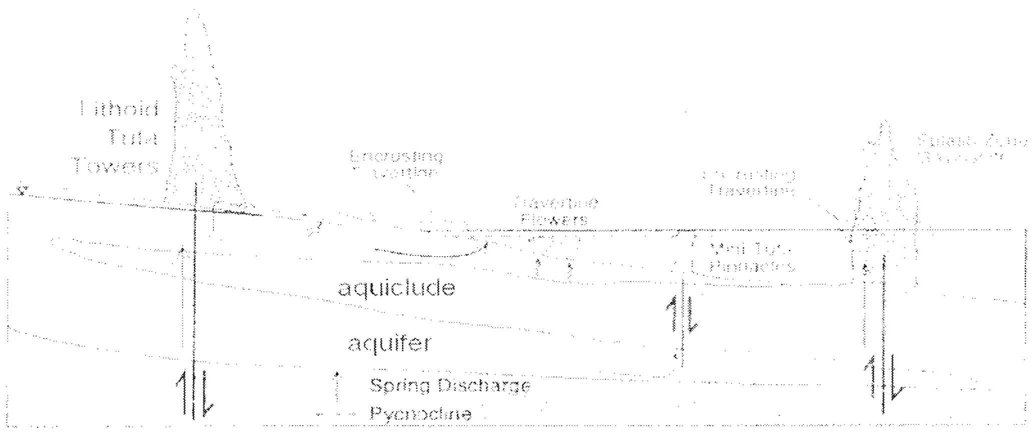


Figure 3. Hypothetical cross section through South Tufa study location showing probable relation between observed tufa forms and spring discharge.

From Council and Bennett, 1993.

Figure 6 (Caption on previous page)

PLAYA EVAPORITES

Evaporites form by precipitation from concentrated brines (see reviews by Kendall, 1978 and Warren, 1989). The concentration mechanism is usually evaporation at the air-water interface, but can also be achieved by freezing, or subsurface hydrological processes, such as ion filtration of connate water. Coarser sediments entering playa lake basins tend to be trapped at higher elevations on alluvial fans (Figure 7). Only the finest materials (mud and silt) make it to the deeper parts of the basin.

Most continental evaporites are deposited in association with playas (dry lake systems) in central basin areas, at lower elevations. In these settings, evaporite minerals usually precipitate in association with fine-grained, often organic-rich, detrital sediments. The extremely flat surfaces that characterize most dry lake basin floors are known by several names, including: playa (most common), continental sabkha or sebkha, salina, saltpan, and chott (Kendall 1978: 67). Smooth playa surfaces are comprised of fine-grained sediments that become broadly distributed over the basin floor, primarily by sheet flooding during rainstorms. The Great Basin as a whole, hosts many such playa lake basins. Along our Hwy 395 fieldtrip route, streams that flow out of the eastern Sierra Nevada Mountains (Figure 8), feed a series of perennial lakes that are essentially open hydrological systems. Although some of these lakes (e.g. Walker) may attain salinities of 15-20 ppt., they do not reach hypersalinity. Exceptions include the eastern shorelines of Mono and Walker Lakes, where playa-like depositional systems, manifested as extensive efflorescent salt crusts, can develop in the summer, due to persistent onshore winds and a gently sloping shoreline.

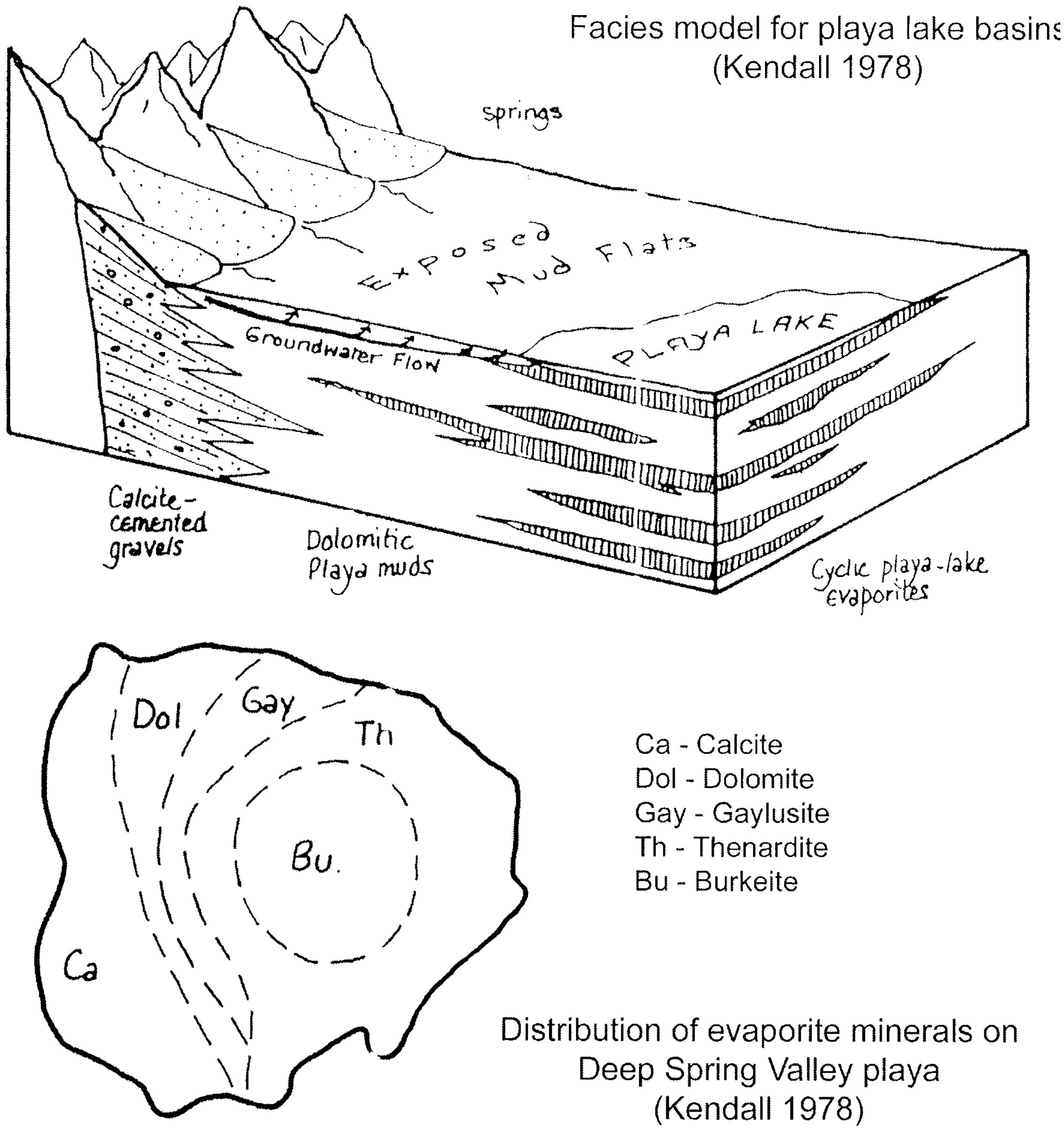


Figure 7: Depositional model (top) for an arid intermontane basin. Simple "bulls eye" pattern of evaporite deposition during a single evaporative cycle, based on solubility.

BASIN AND RANGE TECTONIC HISTORY

During the Mesozoic and Early Tertiary, the plate tectonic margin of western North America was dominated by convergence. Around 140 Ma, the Farallon Plate started to be consumed in a subduction zone located along the western edge of North America (Figure 8, right). Convergence was marked inland by compressional tectonics, uplift of the ancestral Sierra Nevada Mts. (Hill, 2006). and deformation of the western Cordillera (Miller 1970), as recorded in several late Cretaceous to early Tertiary orogenic events (Figure 9, left). Byproducts of this period of plate convergence included the development of island and continental volcanic arcs. The igneous processes are apparent today in the extensive granitic batholiths that make up the Sierra Nevada Mt. range (Hill, 2006).

The ancestral Sierra Nevada Mts., which formed during Mesozoic convergence, achieved an elevation similar to today. However, by 40 Myrs ago, the once towering range had been reduced a much smaller range with an average elevation of ~3,000 ft (Hill, 2006). By 40 Ma, the Basin and Range began to experience crustal extension. This may have been a result of back-arc extensional tectonics. However, several alternative models have been offered to date, as summarized in Figure 9 (Fiero 1986).

Extension was accelerated around 25 Ma, following the ridge-trench collision (Figure 8, left), and the modern Basin and Range Province began to take its present shape. As the ridge bounding the Farallon and Pacific plates collided with the trench system (Atwater, 1970), there was a change in plate motion from direct convergent, to oblique slip. This change in plate motion marked the birth of the present day transform fault boundary (i.e., San Andreas fault system). A remnant of the earlier convergent margin is visible today as the Cascadia Trench, located offshore of Oregon and Washington, which is presently consuming the Juan De Fuca Plate.

Between 8-10 Ma, there was renewed uplift of the Sierra Nevada Mts., with an acceleration of uplift during last 2-4 Ma. There is ongoing debate over what has produced the uplift, but it may involve isostatic adjustments, resulting from thickening of the continental crust, or the emplacement of granitic magmas at depth. This geologically recent increase in the height of the mountain range has had a marked effect on climate, with increased aridity across the Great Basin.

Figure 8: Evolving plate tectonic setting of western North America showing margin tectonics at 40 Myrs. (left) and at 25 Myrs. (right); Source: Atwater (2012; https://www.e-education.psu.edu/).

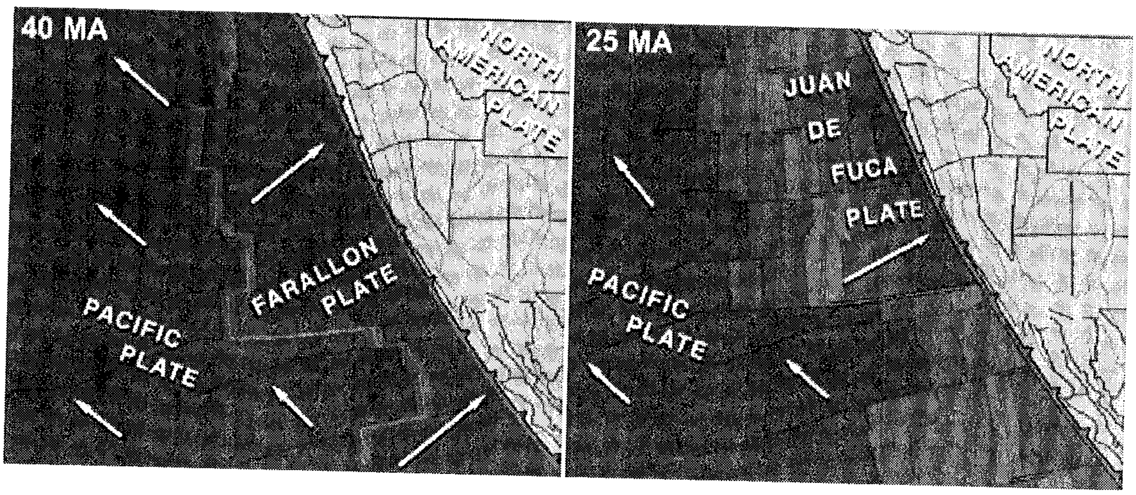
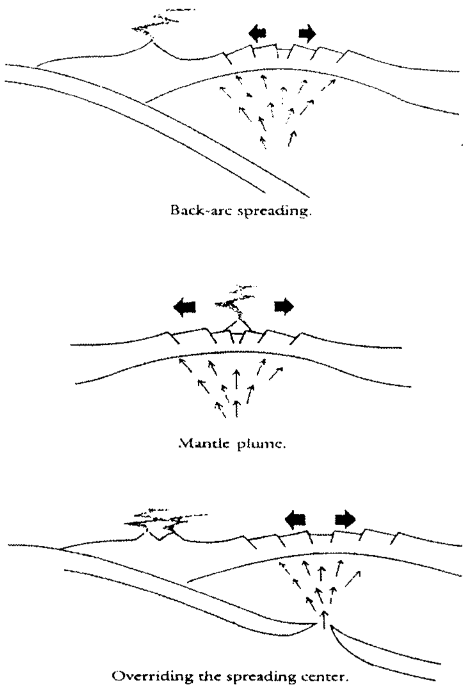


Figure 9: The left image shows three alterative models to explain Basin and Range extension (Fiero 1986). The illustration to the right shows the present plate tectonic setting western North America, dominated by transform faulting along the San Andreas Fault system.



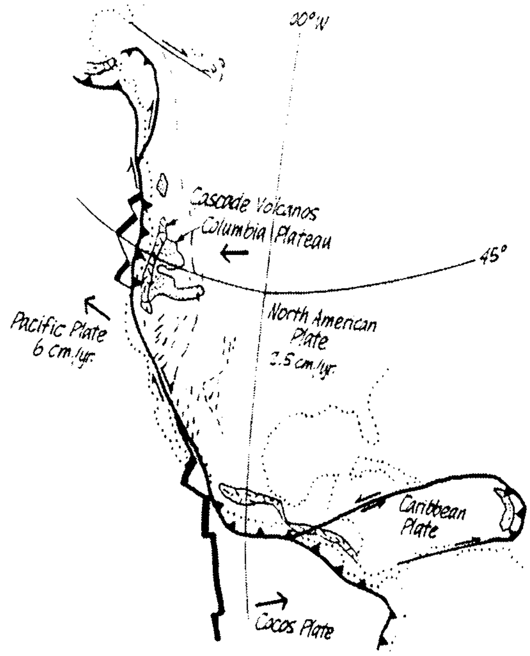
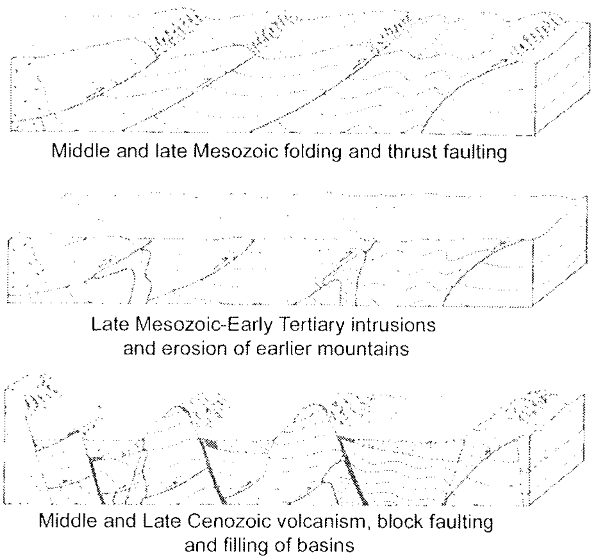


Figure 10: Three panels depict the temporal evolution of Basin and Range structure (and tectonics) from dominantly convergence during the middle Mesozoic to early Tertiary (Upper two panels), to dominantly extension Late Cenozoic (lowermost panel; Miller, 1970).



THE ARID CYCLE OF EROSION

The cycle of landscape evolution under arid climates can be described by the stages represented in Figure 12 (from Miller, 1970). Following uplift, steep canyons incise fault blocks depositing alluvial fans at the mouths of ephemeral streams (Figure 11, left). As mountain ranges are worn down, alluvial fans merge to form a broad rolling surface called a bajada. In the latest stage, a sloping surface called a pediment is formed (the arid equivalent of a peneplain) with isolated erosional remnants (inselbergs) of the former mountain range (Figure 11, right).

Figure 10. Left: Coalescing alluvial fans. Right: Bajada with inselbergs.

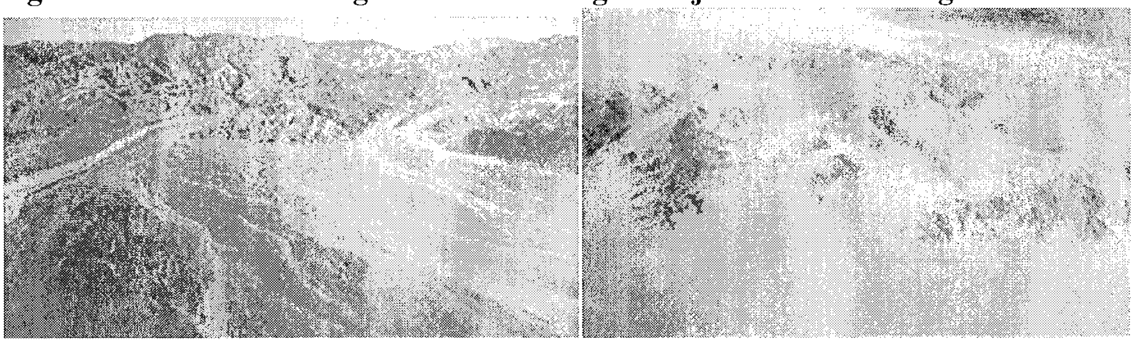
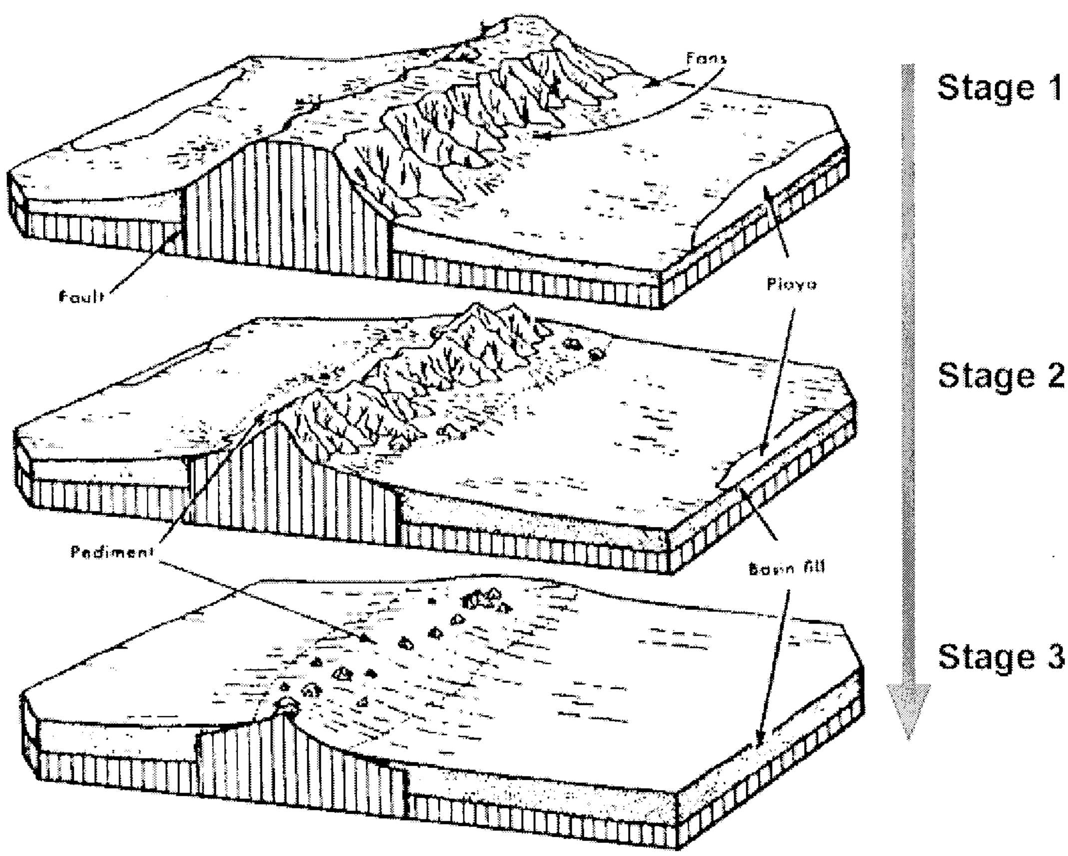


Figure 11: The cycle of erosion under an arid climate from Miller, 1970.



PLUVIAL LAKES OF THE GREAT BASIN

The presently arid, interglacial climate of the Great Basin contrasts markedly with Pleistocene glacial climates. The generally cooler temperature and higher rainfall that prevailed during glacial advances, resulted in much greater run-off to the terminal lakes of the Great Basin. (Mono Lake had up to six times its present water volume). Many basins filled and eventually overflowed, interconnecting to form larger lakes (Figure 12). In the western Great Basin, these "pluvial" lakes grew in size to form a single interconnected system called Lake Lahonton. In the eastern Great Basin, smaller lakes joined to form Lake Bonneville. Drainage from Lake Bonneville flowed north into the Snake River (as it does today), while drainage from the Lahonton system flowed to Death Valley (some speculate, Lahonton flows may have at times even made it to the Colorado River).

Beginning in the Holocene, ~10,000 yrs ago, the ice sheets withdrew and the climate of the Great Basin became drier. The enlarged lake systems began to shrink as the balance shifted toward decreased run-off and increased evaporation. This overall drying trend probably began earlier, in mid-Pleistocene time, starting with the most recent burst in uplift of the Sierra Nevada Mts. The mountain flanks surrounding many of the present valley floors are terraced at higher elevations, due to wave erosion during past lake "highstands". Remnants of former pluvial lakes still survive today in the wetter regions of the Great Basin. Along the eastern side, tucked up against the Wasatch Range, the Great Salt and Utah Lakes remain as survivors of former Lake Bonneville. To the west, tucked in against the Sierra Nevada, Pyrimid Lake, Walker Lake, Mono Lake, and (prior to 1913) Owens Lake, remained as remnants of Pleistocene Lake Lahonton.

Figure 12: Pluvial lakes of the Great Basin. The Lahonton Lake system in the west and the Bonneville system in the east. Light color indicates modern remnants of the former ice age lakes.

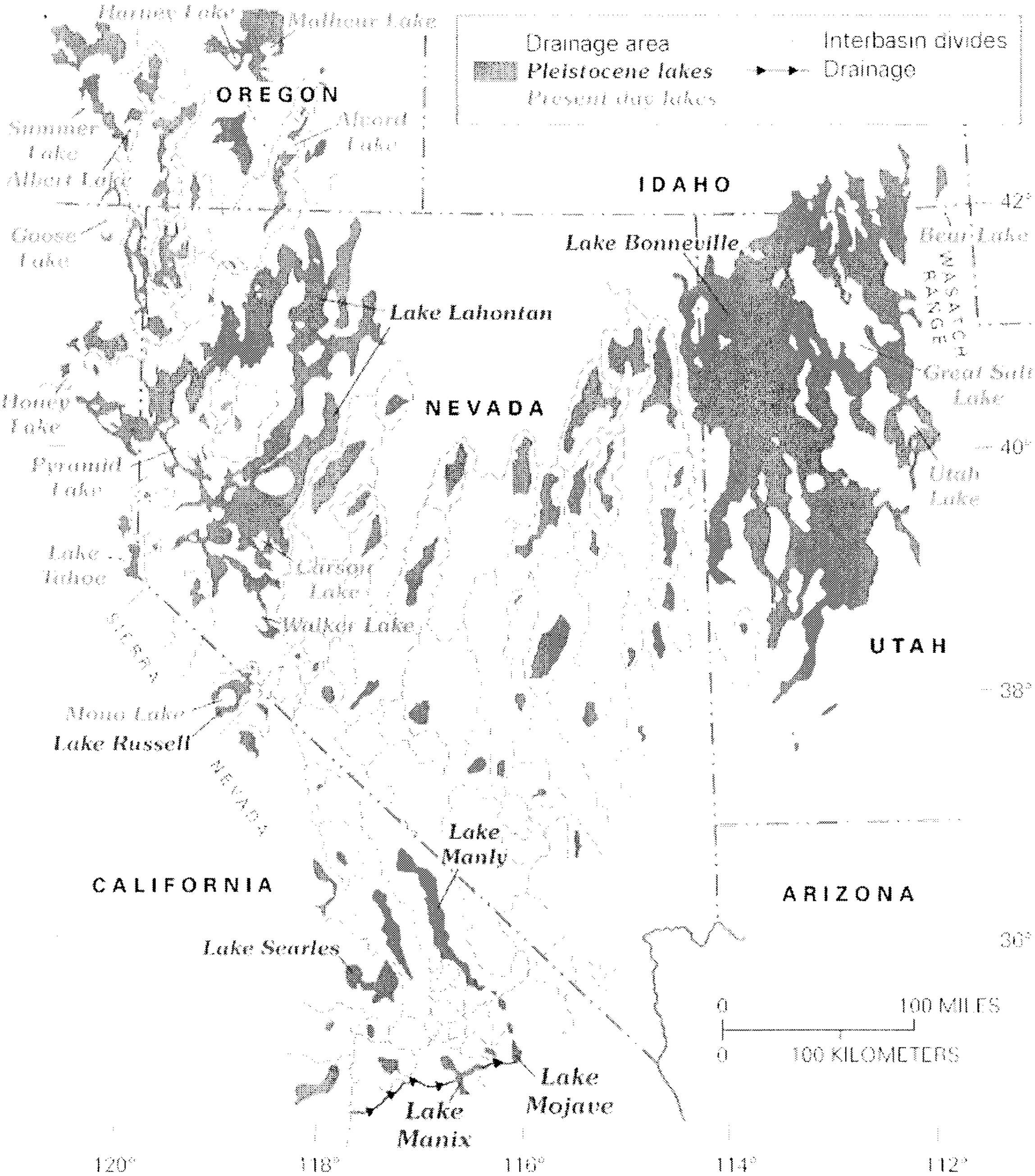


Figure 16: LandSat image of Mono Lake's South Tufa area. There are hints in this image of NE-SW linear trends in the distribution of tufa towers, suggesting fault control.

HALOPHILES AND ALKALIPHILES

Hypersaline environments are challenging for organisms (consider osmotic stress, low oxygen, high pH and the ephemeral nature of many highly saline environments (they tend to dry up!). We have no evidence that Mono Lake has ever dried up and this basis, can safely refer to Mono as a perennial hypersaline lake with salinity averaging ~2-3 times seawater (~70-100 ppt), with a pH between ~10.0-10.5. Despite these harsh conditions, Mono Lake is a veritable microbial factory that sustains a simple, but extremely productive ecosystem, dominated by microbial forms of life. Although fish cannot live at the salinities of Mono Lake, the high productivity of the lake is very important for larger animals as well. Mono is a major refueling station for migratory birds along the Pacific flyway, with large flocks of migratory birds often in view, in particular Wilson's Phalarope (*Phalaropus tricolor*), which feeds primarily on Mono Lake's brine shrimp (*Artemia monica*) and alakali files (*Ephydra hians*). In the months of June and July each year, Phalarope populations arriving from central Canada grow to between 80,000 and 125,000. During these months they double their weight and molt, then undertake an epic 3000 mile journey to Ecuador, where they winter over. Their trip is largely fueled by Mono Lake brine shrimp and alkali flies.

Representatives of all three Domains of life (Bacteria, Archaea and Eukarya) have made the ecological transition to hypersalinity. Halophilic diatoms are very important eukaryotic photosynthetic contributors to Mono Lake productivity and also its biosedimentology; their silica frustrules accumulate on the bottom of the lake in such high abundance, the bottom sediments are best classified as "diatomite" (Note: These deposits make up Paoha Island, which was uplifted from the lake floor during volcanic activity during the 17^{th} century; Hill, 2006).

The halophilic cyanobacteria are another important photosynthetic group, which occur both in the Mono Lake plankton and as biofilms on hard substrates within the photic zone, where they contribute to carbonate precipitation. The order Halobacteriales (more specifically, the halophilic Archaea) contains six genera, which all group with the methanogens in the Euryarchaeota. Halobacteria are salt-requiring and need a minimum salinity of 9% NaCl for growth (seawater is ~3.5%), with optima in the range 21-27% (Norton, 1992). Halobacteria are common in salt lakes of the Great Basin, and although individual species vary in their requirements for cations, other than sodium, the group as a whole can be found anywhere the basic salinity requirements above are met. The presence of halophiles is sometimes detectable by the pink to red color observed in many playa lake brines and associated salt deposits. This color is produced by rhodopsin family photopigments that are characteristic of these groups.

MONO LAKE TUFA TOWERS

Where sublacustrine artesian springs have persisted for some time (e.g. during wet climatic periods or in wetter locations), tufa mounds and larger composite forms (domes, pinnacles and ridges) develop in near shore areas of alkaline lakes. In most cases, clusters of tufa towers and their composite forms define linear trends that parallel regional fracture and fault patterns (Figure 16).

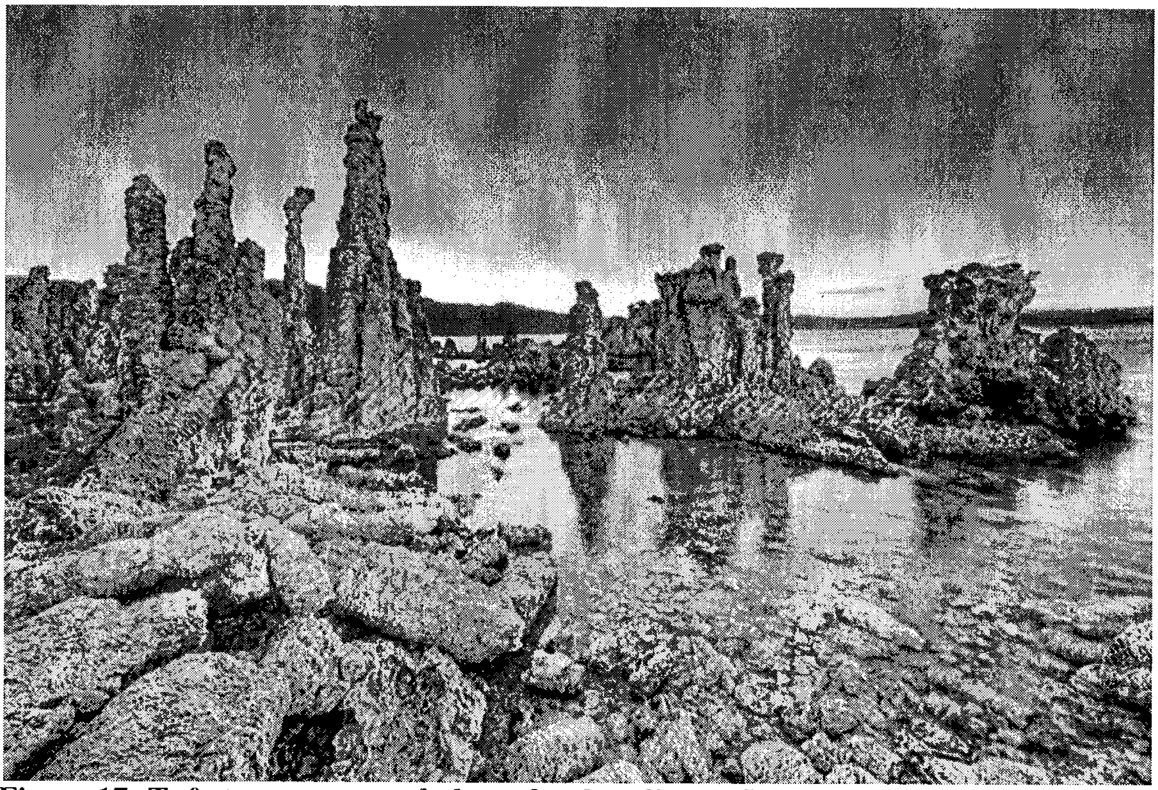


Figure 17: Tufa towers exposed along the shoreline at South Tufa, Mono Lake.

Towers are usually zoned internally, although their interiors are not often exposed. Where I have observed them, the interior regions of towers exhibit a complex network of "feeder tubes" that tend to be in vertically oriented clusters within the core region, becoming more recumbent and irregular as they approach surface of the tower. Tubes sometimes branch, but more typically do not. The walls of tubes are usually laminated and they may be completely infilled, with massive to concentrically laminated carbonate. Tubes tend to be larger and more open near the bases of towers, and narrower, with more infilling near their tops. Where the bases of towers are exposed, feeder tubes have been observed to grade into sand tufas developed within the underlying lake floor sediments.

Towers exhibit a broad range of forms with varying degrees of complexity. At Mono Lake's South Tufa Preserve, the towers tend to relatively simple in form (Figure 27A, B). The towers are usually found in clusters that share a composite base, but many smaller single towers are also observed. The tall, narrow towers at South Tufa are comparatively unstable and many have fallen over. Please do not climb on the towers or otherwise lean against them. In contrast, on the north shore of Mono Lake, at County Park Tufa Preserve, towers tend to occur as more massive "mushroom" shapes, with many projecting horizontal ledges with flat-bottomed overhangs. It is likely that such ledges and overhangs formed by precipitation along thermoclines, or haloclines, as the towers grew upward into shallow water, or "emerged" to shallower depths with drops in lake level.

NAVY BEACH SAND TUFA:

These deposits occur where unconfined aquifers percolated upward through coarse, sandy shoreline deposits, mixing with lake water in pore spaces precipitating calcite cement. At Navy Beach, exposed on terraces above the exposed sandy strata, tufas take several different forms, including crusts, mounds and small columns. These are interpreted to be the incipient tufa towers that were arrested in their development, probably because the water depth was too shallow to form large towers.

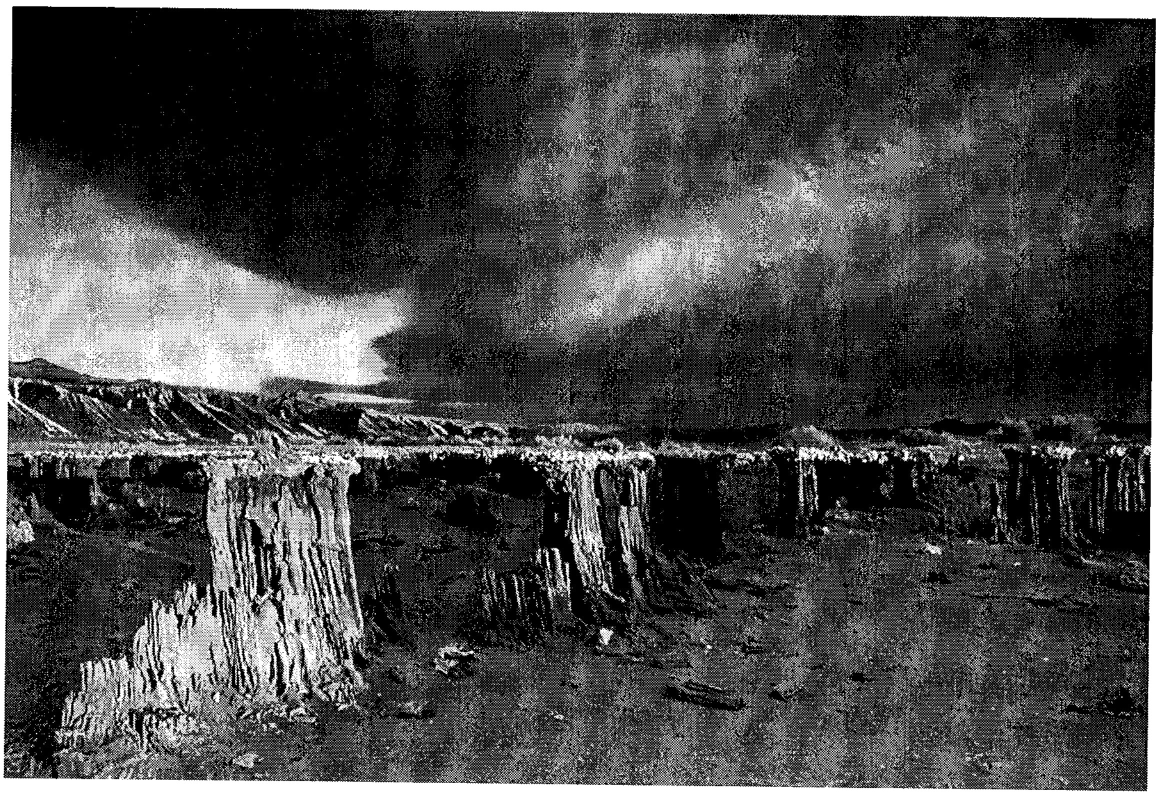


Figure 18: Sand tufas exposed by wave erosion along the southern shoreline of Mono Lake at Navy Beach.

Sand tufas formed where sediments were fluidized and selectively cemented, by upward-flowing groundwater that mixed with alkaline lake water in the sediments. The sandy "tubes" at Navy Beach are oriented ~perpendicular to bedding and penetrate through individual sedimentary beds. In places, the walls of the tubes are attached to layers of indurated, coarser sand that was a locus for carbonate cementation. The tubes are frequently hollow, but sometimes are filled with carbonate-cemented, homogenous, fine sand.

At other localities I have observed sand tufas lying immediately beneath tufa bowls and small towers and consider them to represent an early stage in spring related tufa deposition that precedes the formation of tufa bowls, heads and eventually, towers. Examples from other tufa localities in the Owens Valley (CA) and Wadsworth (NV) show concentrically laminated (lithoid) infills of calcium carbonate, with laminae oriented parallel to tube walls. Some older sand tufas have been observed to directly connect to the internal plumbing system of large tufa towers.

SHORELINE TUFAS AND CEMENTS

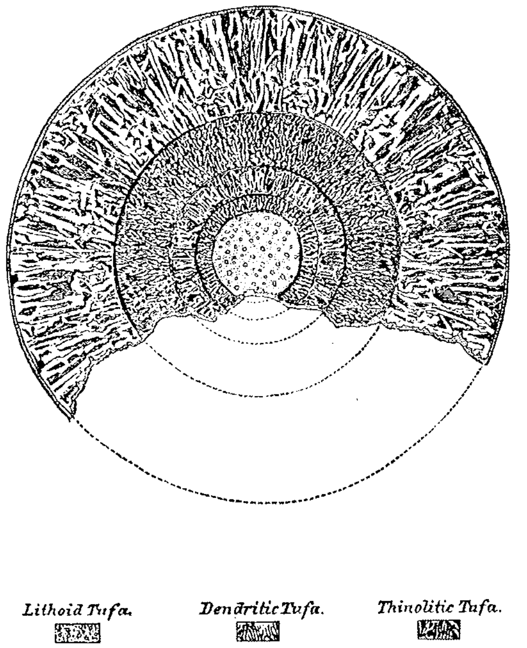
Where shallow aquifers are unconfined (i.e., lacking an aquiclude) or at the surface, groundwater outflows are more broadly distributed, being controlled primarily by the geometry and permeability of the aquifer (Los Angeles Dept. of Water and Power, 1987). In this situation, carbonate precipitation occurs within sediments, forming broadly-distributed sedimentary cements In shoreline areas where streams enter the lake, lenses of fresh water flows out onto the saline lake water forming a density stratified lens of freshwater. In near shore areas where waves cause mixing of the two water masses, tufa precipitates as cements and coatings on boulders and bedrock surfaces along the shoreline. This process is probably assisted by the increased degassing of CO_2 due to wave turbulence (Stine 1984: 35). Some beach tufas resemble stromatolites, excellent examples of which may be found in Walker Lake at Hawthorne (NV), about 50 miles NE of Mono Lake.

TUFA TEXTURES:

Exposed tufa towers have a hardened shell that forms with subaerial exposure due to meteoric weathering (dissolution/reprecipitation). In contrast, tufa tower interiors are generally porous, with a distinctly "cellular" fabric. In cross sections of tufa towers, Russell (1889) observed that outside of the porous interior zones of tufa towers is often seen a zone of "lithoid" tufa, composed of dense (sometimes laminated) carbonate (Figure 19, left). In cross sections of towers, lithoid tufas are usually succeeded by a zone of “dendritic” tufa, which consists of elongated clusters of outward expanding, dendritic growths of calcite. In other cases, interior surfaces are covered by "polyp-like" growths of tufa that are internally laminated, resembling columnar stromatolites.

Thinolite is a textural type that is a widespread in "ice age" tufas of the Mono Basin. It is common in the outer zone of older tufa towers exposed on highstand terraces of Pleistocene Lake Russell (Figure 19). It also forms less extensively in Mono Lake along the shoreline during winter, when temperatures reach ~4 °C. It forms preferentially where freshwater inflows from springs or streams mixes with denser, hypersaline lake water, along the shoreline of Mono Lake.

Another common tufa type forming in the lake today is a so-called "pustular" variety that exhibits a "popcorn-like" texture. Pustular tufas are common on upward facing, well-lighted surfaces in shallow water, where they form in association with cyanobacterial biofilms (Figure 20).



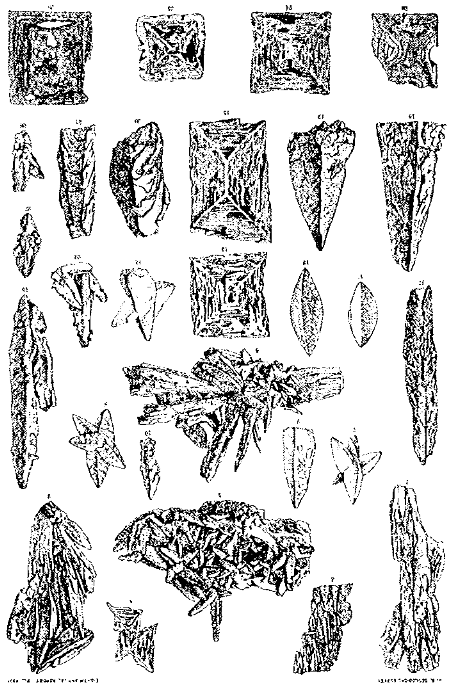


Figure 19: Tufa Textures. Left: Textural zonation of tufa towers documented by Russell (1889). Right: Ice age thinolite tufa varieties (also from Russell 1989).
Figure 18: Right (cont.): Thinolite is a calcite pseudomorph that forms by replacement of ikaite ($CaCO_3 \cdot 6H_2O$).