



The Mono Basin as it might have looked around the time of the penultimate glaciation, showing glaciers calving into Pleistocene Lake Russell, which episodically spilled to the southeast through Adobe Valley. Modern Mono Lake shown for perspective. Illustration by Eric Knight, www.ericknightmaps.com.

Friends of the Pleistocene Pacific Cell Annual Meeting

Mono Basin, California

October 2-4, 2025

Organizers: Zachary Smith & Kim Blisniuk

Contributions from: Genna Chiaro, William Hammond, Scott Stine, Lauren Harrison, Marith Reheis, Greg Stock, Elizabeth Haddon, Michael Oskin, Sidney Hemming, Jorge Vazquez, Bartshe Miller, Fred Phillips, Mark Woyshner, Chris Wills, Ken Hudnut, Natalie Homyk

Disclaimer: Contributions included in this field guide have not undergone peer review or extensive formatting. The views, interpretations, and conclusions presented herein are those of the individual contributors and do not represent those of their affiliated institutions or organizations. Although reasonable efforts have been made by individuals to ensure the accuracy of the information presented, the organizers and contributors make no warranties or representations regarding its completeness, reliability, or accuracy.

Table of Contents

Itinerary	6
Driving Directions	
Day 1	11
Day 2	16
Day 3	21
Day 1 – Mono Lake Shoreline and Volcanoes	24
Stop 1: Pumice Valley, Junction of Hwy 395 and Hwy 120	25
Introduction to the Mono Basin	26
Geologic Setting	27
Mono Basin Water Budget	30
<i>Scott Stine</i>	
Stop 2: Panum Crater	31
Regional Volcanism in the Mono Basin and Broader Long Valley Region	32
<i>Genna Chiaro</i>	
Age Constraints on the Formation of Paoha Island	38
<i>Lauren Harrison</i>	
Age and Mechanisms of Volcanism and Deformation in Mono Lake, California	44
<i>Zachary Smith, Kathleen Rodriguez</i>	
Volcanic History and Recent Observations of Climate Change	51
<i>Scott Stine</i>	
Stop 3: South Tufa	54
Conservation in the Mono Basin	55
<i>Bartshe Miller</i>	
Groundwater Source Characterization for Glass Mountain Meadows Restoration	58
<i>Erika Groh, Mark Woyshner, and Steve Hnat</i>	
Heat Flow as a Tracer of Tufa Formation in Mono Lake	61
<i>Zachary Smith</i>	
Tufa Formation and Implications for Age Dating	75
<i>Scott Stine</i>	
Stop 4: Panum Crater Block and Ash Flow	81
Panum Crater Block and Ash Flow and Mono Lake Volcanism	82
<i>Scott Stine</i>	
Stop 5: Mono Lake Boat Launch or Old Marina	87
Pumice Rafts and Lake Floor Mass Flow Deposits	89
<i>Natalie Homyk, Zachary Smith</i>	
Mono Lake Boat Launch Immediately West of Lee Vining Creek	97
<i>Scott Stine</i>	
Day 2 – Sierra Nevada Range Front and Tioga Pass	100
Stop 6: Scenic Overlook	101
Dextral Slip along the Sierra Nevada Range Front north of Owens Valley	102
<i>Michael Oskin</i>	

Table of Contents (continued)

A Reevaluation of Moraine Chronology at Pine Creek, Inyo County, California	107
<i>Fred Phillips, Jeffery Lee, and Kimberly Blisniuk</i>	
Active Tectonics and Fault Slip Rates from GNSS Measurements around the Mono Basin, Walker Lane, Western Great Basin	112
<i>William Hammond</i>	
Earthquake Scenarios for the Long Valley Caldera–Mono Lake Area	116
<i>Chris Wills</i>	
Normal Fault Displacement along the Eastern Sierra Nevada Frontal Fault System	121
<i>Ken Hudnut</i>	
Stop 7: Bloody Canyon	123
<i>Fred Phillips</i>	
Stop 8: Tioga Pass Stop 1	124
Last Glacial Maximum (Tioga) Ice Flow across Tioga Pass: Glacial Striations North of Tioga Pass	125
<i>Greg Stock</i>	
Stop 9: Tioga Pass Stop 2	127
Rates and Patterns of LGM (Tioga) Deglaciation at Tuolumne Meadows and Tioga Pass	128
<i>Greg Stock, Elizabeth Haddon, and Andrew Jones</i>	
Stop 10: Lee Vining Canyon	142
Contemporary Uplift of the Sierra Nevada from GNSS Networks	143
<i>William Hammond</i>	
Reconstruction of Late Cenozoic Fault Displacement on the Eastern Sierra Nevada Escarpment Based on Eastern Central Valley Sedimentary Structure	147
<i>Fred Phillips</i>	
Geomorphic Implications of Fracture Density Differences in Granitic Rocks across the Sierra Nevada Crest	153
<i>Greg Stock</i>	
Day 3 – Wilson Creek Formation and Lake Russell	157
Stop 11: Wilson Creek Canyon	158
Wilson Creek Canyon Stratigraphic and Geochemical Evidence for Lake Level Variations through the Last Glacial Cycle	159
<i>Sidney Hemming</i>	
Stop 12: Wilson Creek Formation at Grand Gash	165
Pleistocene History of Lake Russell	166
<i>Marith Reheis</i>	
Stop 13 (optional): High Shorelines – Beach Deposits of Early to Middle Pleistocene Lake Russell	169

FOP 2026 FIELD GUIDE



Friends of the Pleistocene
Pacific Cell Annual Meeting
Mono Basin, California
October 2-4, 2026

Itinerary

Day 1

Leave Campground: 8:00 AM

Stop 1: Pumice Valley, Junction of Hwy 395 and Hwy 120 (37.8866395, -119.0900869)(Talks Start ~9:00 AM)

1. Zach Smith Introductory Remarks
2. Scott Stine

Logistics

Drive Time 35 min + Parking Time 20 minutes

Introductory remarks + talks: 15 minutes

Stop 2: Panum Crater (37.913965, -119.046577)(Talks Start ~10:00 AM)

1. Genna Chiaro
2. Lauren Harrison
3. Scott Stine

Logistics

Drive Time 5 min + Parking Time 20 min

Talks: 30 minutes

Walk to Panum Crater: 90 minutes

Stop 3: South Tufa (37.93867, -119.02710) (Talks Start ~1:00 PM)

1. Bartshe Miller
2. Erika Groh
3. Zachary Smith
4. Scott Stine

Logistics

Drive Time 5 min + Parking Time 20 min

Talks and lunch: 60 Minutes

Walk to see tufa and possibly soda springs: 60 Minutes

Stop 4: Panum Crater Block and Ash Flow (37.94677, -119.05140)(Talks Start ~2:30 PM)

1. Scott Stine

Logistics

Drive Time 5 min + Parking Time 20 min

Talks: 10 Minutes

Walk to see block and ash flow: 15 Minutes

Stop 5: Mono Lake Boat Launch or Old Marina (37.93867, -119.02710)(Talks Start ~3:30 PM)

1. Scott Stine
2. Zachary Smith

Logistics

Drive Time 15 min + Parking Time 20 min

Talks: 15 Minutes

Walk to pumice raft and tufa: 20 Minutes

Return to Campground: 5:00 PM

Day 2

Stop 6: Scenic Overlook (Rates n Dates) (37.518887, -118.604094)(Talks Start ~9:00 AM)

1. Mike Oskin
2. Fred Phillips
3. Bill Hammond
4. Chris Willis
5. Ken Hudnut

Logistics

Drive Time 30 min + Parking Time 20 min

Talks: 60 Minutes

Stop 7: Bloody Canyon (37.894927, -119.127370)(Talks Start ~11:30 AM)

1. Fred Phillips

Logistics

Drive Time 70 min + Parking Time 20 min

Talks: 15 Minutes

See moraines: 15 minutes

Stop 8: Tioga Pass Stop 1 (37.912317, -119.257987) (Talks Start ~1:00 PM)

1. Greg Stock

Logistics

Drive Time 25 min + Parking Time 20 min

Talk: 15 Minutes

See glacial features: 15 minutes

Stop 9: Tioga Pass Stop 2 (37.917664, -119.255884)(Talks Start ~1:45 PM)

1. Greg Stock
2. Beth Haddon

Logistics

Drive Time 2 min + Parking Time 20 min

Talks: 30 Minutes

See glacial features: 30 minutes

Stop 10: Lee Vining Canyon (37.940434, -119.230002)(Talks Start ~3:30 PM)

1. Bill Hammond
2. Greg Stock
3. Fred Phillips

Logistics

Drive Time 10 min + Parking Time 20 min

Talks: 45 Minutes

Return to Campground: 5:00 PM

Day 3

Leave Campground: 9:00 AM

Stop 11a: Wilson Creek Formation - Wilson Creek Canyon (38.02869, -119.13120)(Talks Start ~10:00 AM)

1. Jorge Vazquez
2. Sidney Hemming
3. Scott Stine

Logistics

Drive Time 55 min + Parking Time 20 min

Talks: 30 minutes

Walk to Wilson Creek Canyon: 90 minutes

Stop 11b: Wilson Creek Formation - Seismites along road, Wilson Creek cut, (38.02869, -119.13120) (Talks Start ~12:00 AM)

1. Jorge Vazquez
2. Sidney Hemming
3. Scott Stine

Logistics

Talks: 5 minutes

Walk to Wilson Creek Canyon: 15 minutes

Stop 12: Wilson Creek Formation - Grand Gash (38.023631, -119.148737) (Talks Start ~1:00 PM)

1. Jorge Vazquez
2. Sidney Hemming
3. Scott Stine

Logistics

Drive Time 5 min + Parking Time 20 min

Talks: 30 minutes

Walk to Wilson Creek Canyon: 45 minutes

Stop 13: High Shorelines (38.09401, -119.08560)

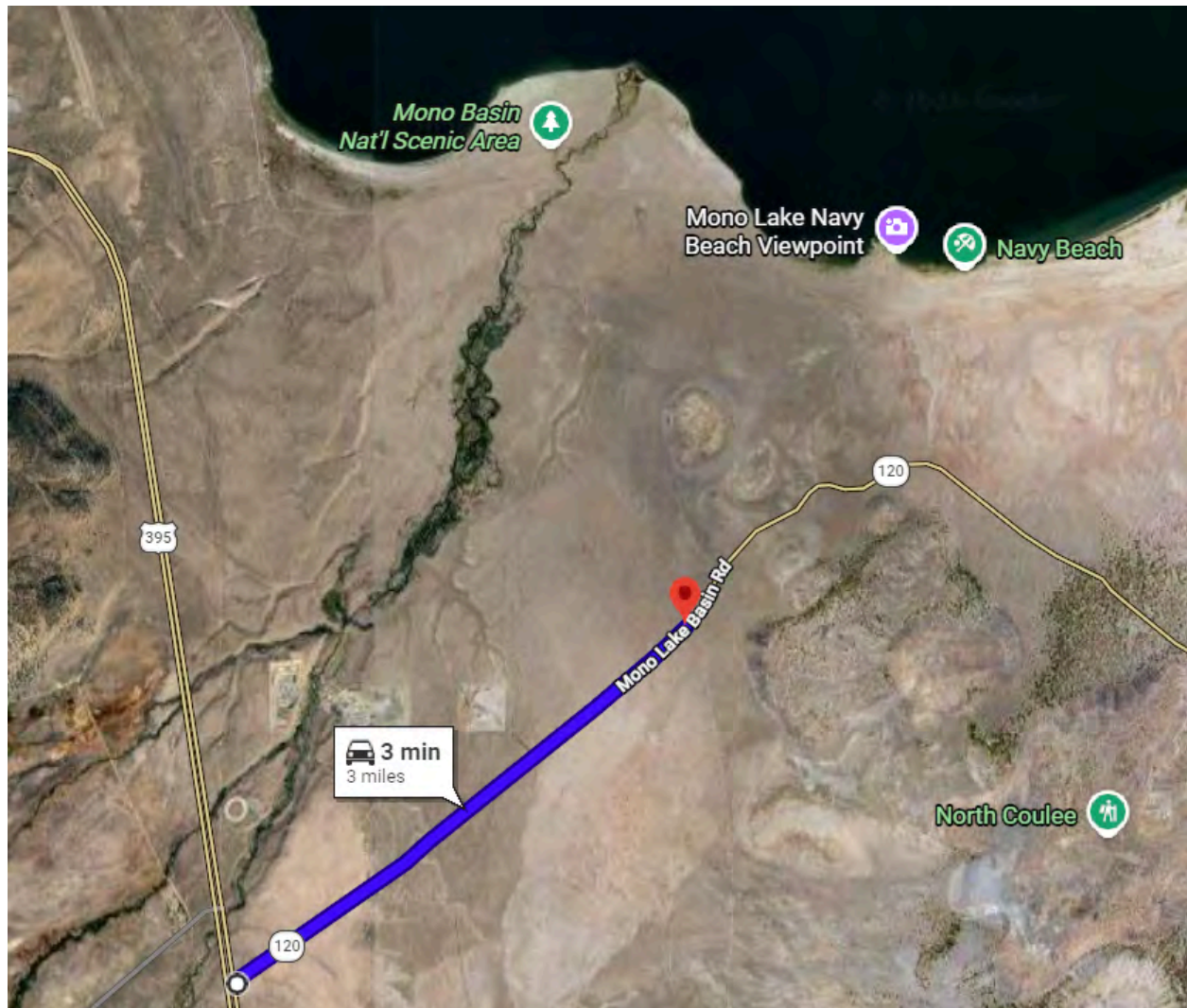
1. Marith Reheis

Day 1 Driving Directions

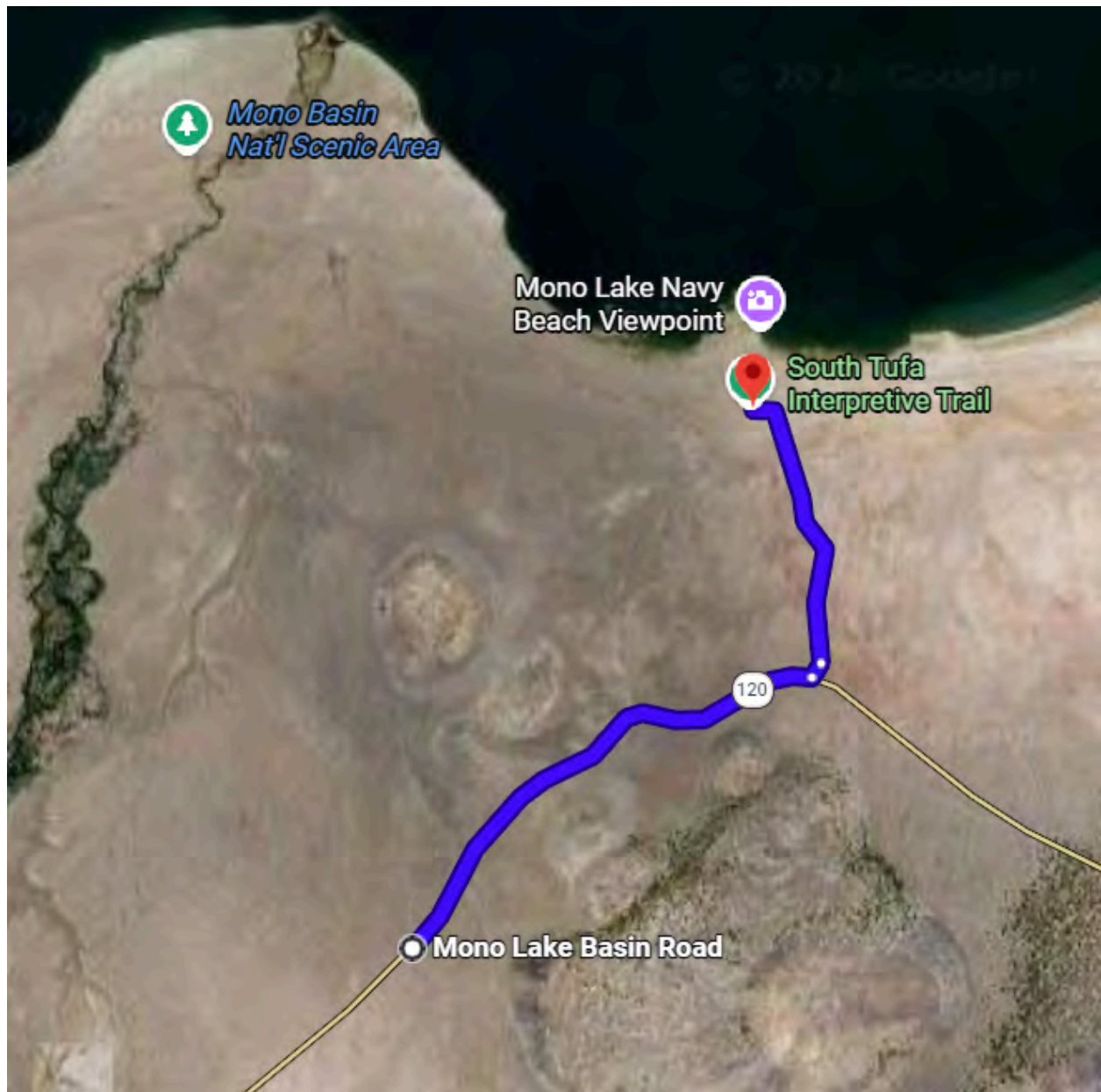
[Brown's Owen's River Campground to Hwy 395, Hwy 120 Junction](#)



[Hwy 395, Hwy 120 Junction to Panum Crater](#)



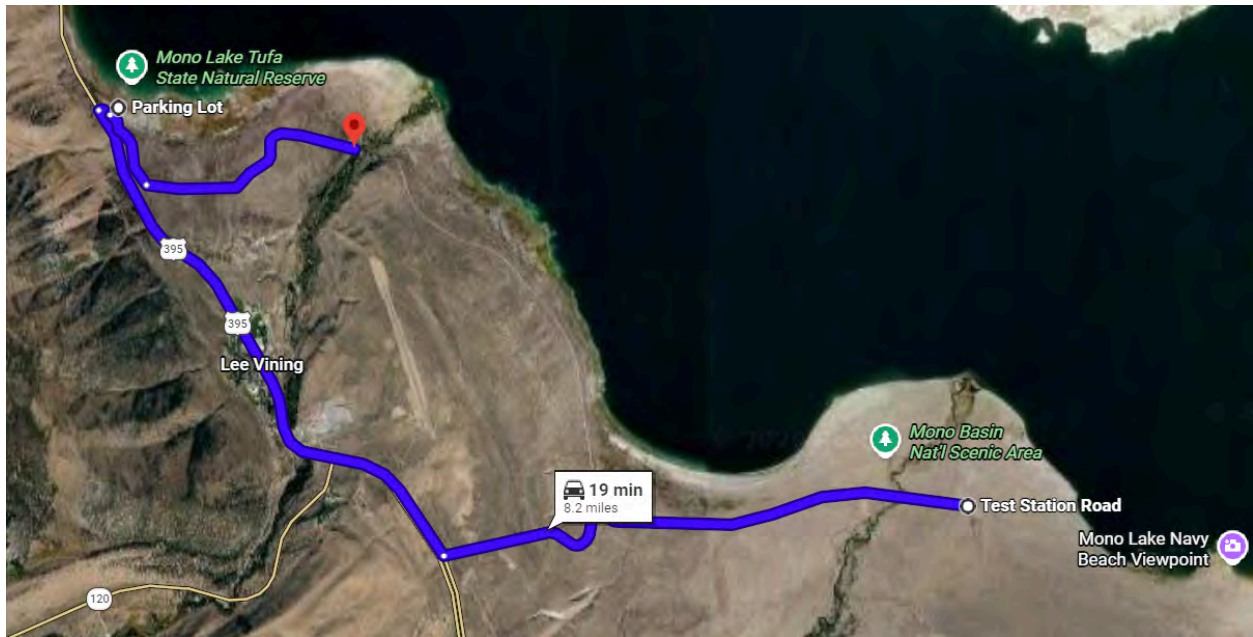
Panum Crater to South Tufa



South Tufa to Panum Crater Block



Panum Crater Block and Ash Flow to Mono Lake Boat Launch



If time is short we will hold presentations at the Mono Lake State Natural Reserve (Old Marina)



Day 2 Driving Directions

[Brown's Owens River Campground to 395 Vista Point](#)



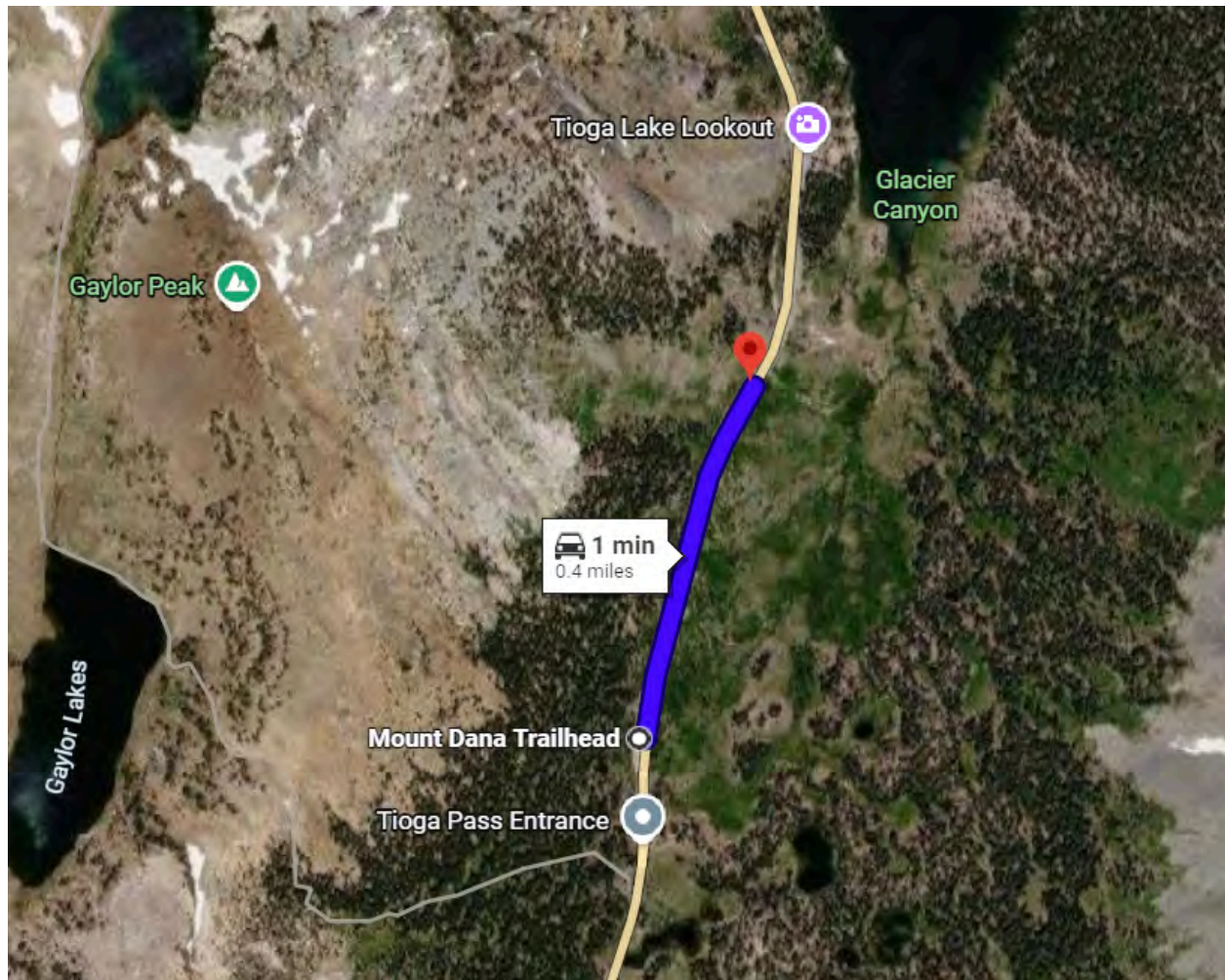
395 Vista Point to Bloody Canyon



Bloody Canyon to Tioga Pass Stop 1



Tioga Pass Stop 1 to Tioga Pass Stop 2



Tioga Pass Stop 2 to Lee Vining Canyon

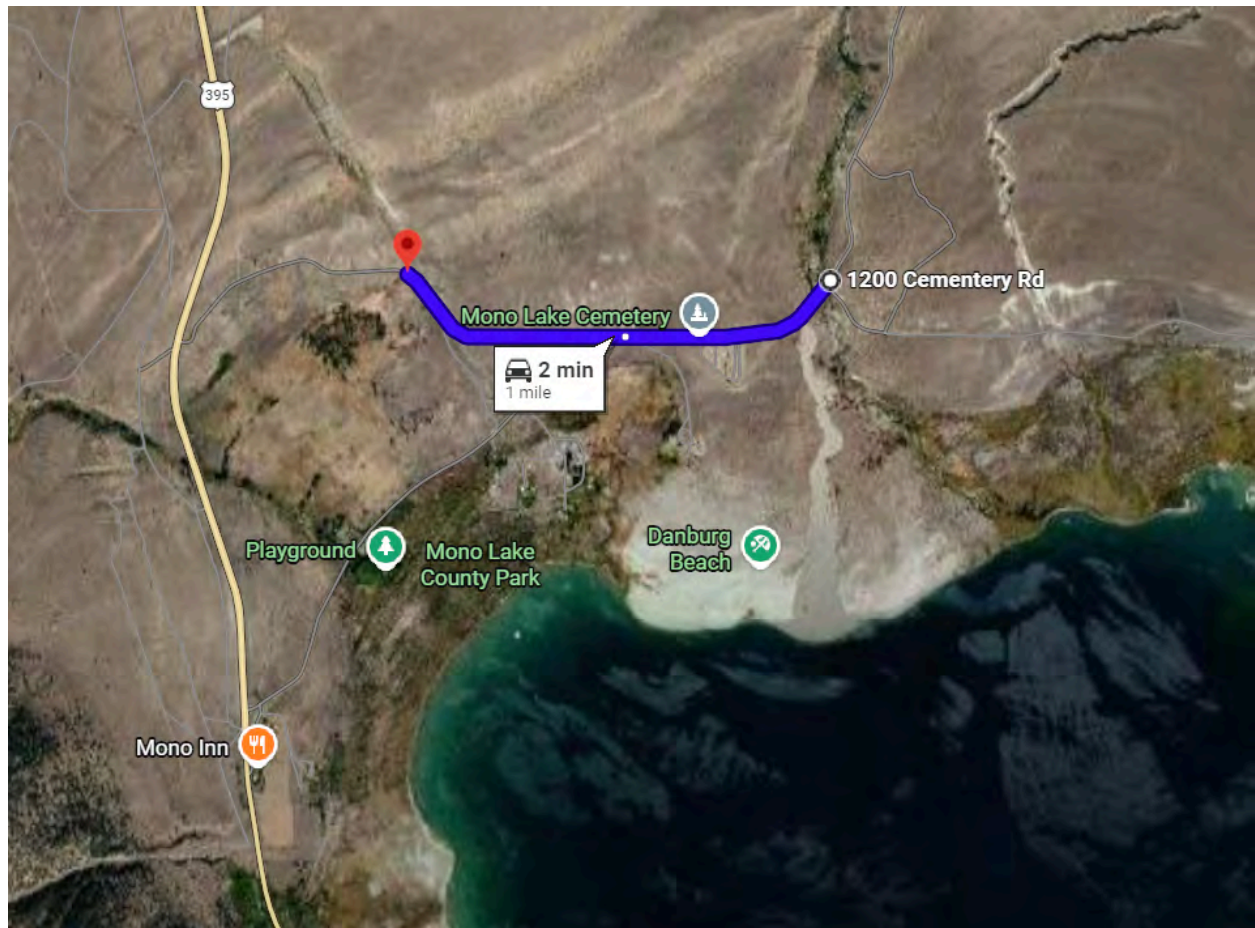


Day 3 Driving Directions

Brown's Owens River Campground to Wilson Creek Canyon



Wilson Creek Canyon to Grand Gash



FOP 2026 ITINERARY



FOP 2026 DAY 1



Day 1, S1: Pumice Valley, Junction of Hwy 395 and Hwy 120

FIELD TRIP STOP CONTRIBUTORS: Zachary Smith, University of California Berkeley; Kim Blisniuk, San Jose State University; Scott Stine, Emeritus California State University, East Bay

LOGISTICS: Park on the highway and join the group for introductory remarks and a talk about water.

WALKING DISTANCE: Minimal. Talks will be given along the road.

LOCATION: Pumice Valley (37.8866395, -119.0900869)

DIRECTIONS: [Brown's Owens River Campground to Hwy 395, Hwy 120 Junction](#)



Introduction to the Mono Basin

Welcome to the Mono Basin! Few places in North America pack so much Quaternary geology into such a small area. From a single vantage point on the basin rim you can see a volcanic island that emerged from the lake geologically yesterday, a chain of rhyolite domes less than a thousand years old, the stranded shorelines of a pluvial lake that once stood more than 200 m above today's water surface, moraines that spill out of Sierran canyons toward the lakeshore, and the fault scarps that have been building the whole landscape. Over the next three days we will work our way through that story: on Day 1 along the shores and volcanoes of Mono Lake, on Day 2 along the Sierra Nevada range front and up to the glaciated crest at Tioga Pass, and on Day 3 through the lake sediments and high shorelines that record the history of Pleistocene Lake Russell.

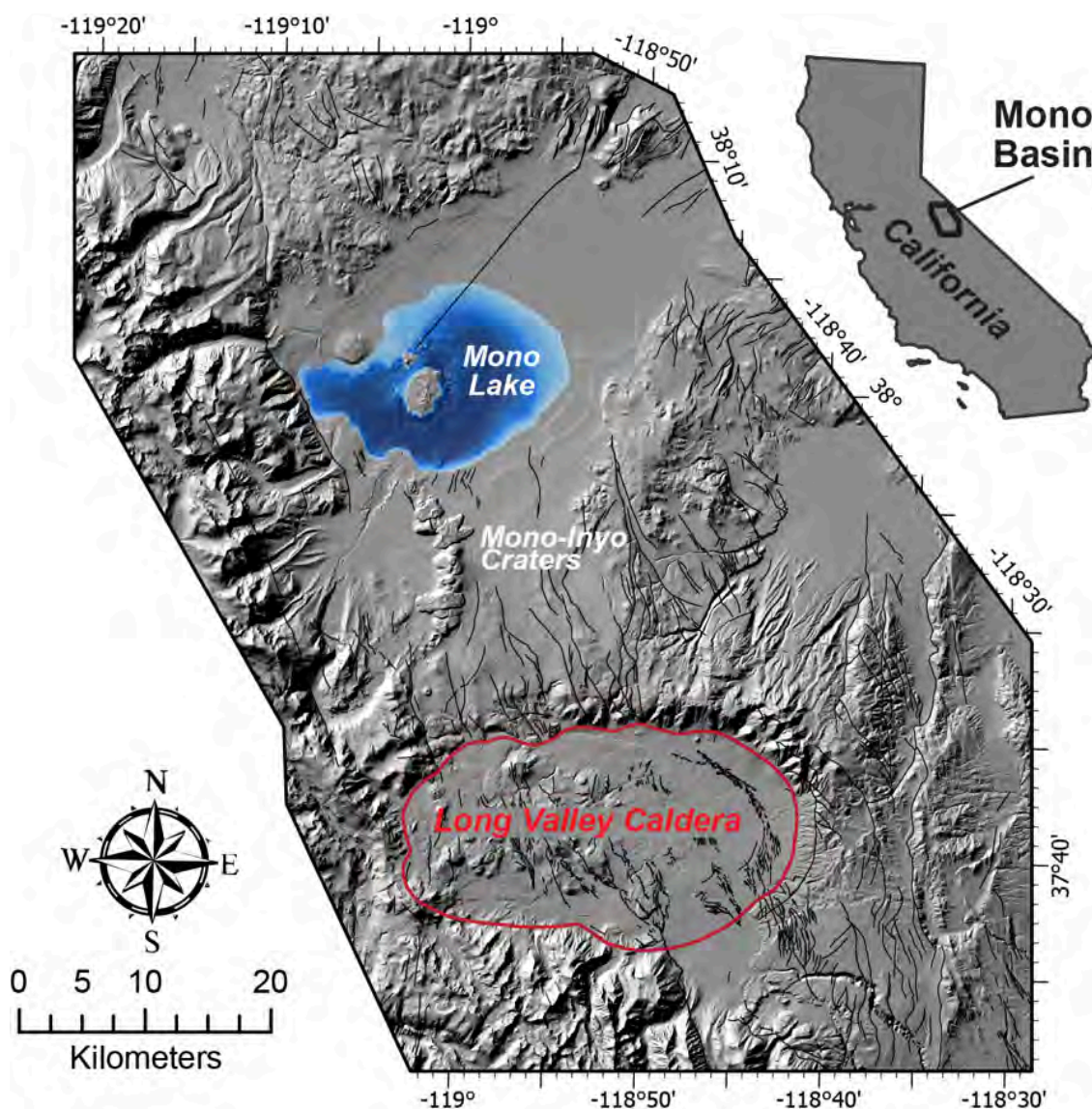


Figure S1.1 Volcanic features and Quaternary faults in and around the Mono Basin.

Geologic Setting

The Mono Basin lies where the eastern escarpment of the central Sierra Nevada meets the Basin and Range. It is a hydrographically closed basin, and at its floor elevation of roughly 1,950 m it is the highest of the large closed basins in the province (Reheis et al., 2002). The basin is bounded on the west by the Sierra Nevada, on the north by the Bodie Hills, and on the south and southeast by the volcanic highlands around Glass Mountain and the Long Valley Caldera and by the Adobe Hills. Streams draining the Sierra, chiefly Rush, Lee Vining, and Mill creeks, supply most of the water to Mono Lake, which has no outlet and loses water only by evaporation.

Tectonically, the basin sits in the central Walker Lane, a belt of transtensional deformation that accommodates a share of Pacific–North America plate motion east of the Sierra Nevada. Much of the dextral shear carried north through Owens Valley steps eastward here across the left-lateral faults of the Mina Deflection (Wesnousky, 2005; DeLano et al., 2019), leaving the Mono Basin as a zone dominated by extension. The west side of the basin is defined by the down-to-the-east Mono Lake fault along the Sierra Nevada range front. Bursik and Sieh (1989) showed that range-front faulting and the dike-fed volcanism of the Mono–Inyo chain are closely linked, and geodetic data continue to document active strain, magmatic inflation at Long Valley, and possible ongoing uplift of the Sierra Nevada relative to the Walker Lane (e.g., Hammond et al., 2024). Several of these themes, including how fast the faults are moving and how we date the landforms they offset, are the focus of our “Rates n Dates” discussions on Day 2.

Volcanism

The Mono Basin is part of the broader Long Valley volcanic field, which is best known for the ~767 ka eruption of the Bishop Tuff and the collapse of the Long Valley Caldera (Hildreth, 2004). North of the caldera, the Mono–Inyo volcanic chain extends to the south shore of Mono Lake. The Mono Craters alone comprise about thirty high-silica rhyolite domes and flows; dated ashes preserved in the lake sediments of the Wilson Creek Formation show that rhyolitic volcanism here began at least ~65 ka, and eruptions have continued into the late Holocene. The most recent episode, the North Mono eruption of about 700 years ago, produced Panum Crater and the Panum block-and-ash flow, both of which we visit on Day 1 (Sieh and Bursik, 1986). Volcanism has also occurred within the lake itself: the basaltic cone of Black Point erupted beneath the waters of Lake Russell, and Negit and Paoha islands are volcanic edifices built by young eruptions and intrusions into the lake-floor sediments. Paoha in particular appears to be very young, and its age and the mechanisms behind its ongoing deformation are topics of active research that we will discuss on Day 1 (see also Stine, 1990).

Pleistocene Lake Russell and Holocene Mono Lake

A lake has existed in the Mono Basin, at least episodically, since the late Pliocene (Gilbert et al., 1968). Its Pleistocene predecessor, Lake Russell, was named for Israel Russell, whose 1889 monograph was the first systematic study of the basin. During the early and middle Pleistocene, Lake Russell repeatedly rose to 2,205–2,280 m and overflowed northward through the Mount

Hicks spillway into the Walker Lake drainage; later faulting in the Adobe Hills lowered the southeastern rim, and the lake then spilled southward through Adobe Valley into the Owens River system (Reheis et al., 2002). The frontispiece of this guidebook illustrates the basin at about the time of the penultimate glaciation, when glaciers descending from the Sierra calved into an expanded Lake Russell. During the last glacial maximum (MIS 2) the lake reached about 2,155 m, roughly 200 m above the modern surface.

The late Pleistocene history of the lake is recorded in the silts, sands, and ashes of the Wilson Creek Formation, first described in detail by Lajoie (1968). Its nineteen ash layers provide a tephrochronologic framework that ties the stratigraphy together around the basin and anchors records of lake level, water chemistry, and paleomagnetism (e.g., Vazquez and Lidzbarski, 2012). In the Holocene, Mono Lake has continued to rise and fall with climate. Drowned tree stumps and stranded shorelines around the lake document deep lowstands during medieval droughts and a highstand during the Little Ice Age (Stine, 1987, 1990, 1994), making the lake one of the most important high-resolution records of late Holocene hydroclimate in western North America.

Glaciation

During Pleistocene glaciations, valley glaciers flowed east from the Sierra crest down Lee Vining, Bloody, and Rush Creek canyons and built the large lateral and terminal moraines that frame the western side of the basin. The moraines at Bloody Canyon are a classic locality in the development of the Sierra Nevada glacial sequence (Sharp and Birman, 1963) and were among the first moraines anywhere to be dated with cosmogenic nuclides (Phillips et al., 1990). On Day 2 we will follow these glaciers to their source at Tioga Pass, where new mapping of erratic boulders and cosmogenic exposure ages shed light on how the ice flowed across the Sierra crest and how quickly it disappeared at the end of the last glaciation.

Water, Conservation, and the Modern Lake

Mono Lake is also a landmark in the history of water policy in the western United States. In 1941 the Los Angeles Department of Water and Power began diverting water from the lake's tributary streams. Over the following four decades the lake dropped about 45 vertical feet, lost half its volume, and doubled in salinity, exposing the tufa towers that have become symbols of the basin. Following years of citizen advocacy and litigation, the California State Water Resources Control Board's Decision 1631 (1994) set a management level of 6,392 ft above sea level. As of September 1, 2026, the lake stood 9.6 ft below that level, and the State Water Board has signaled that a formal hearing on Los Angeles' water exports is forthcoming. These issues, and how geology informs restoration elsewhere in the region, are part of our discussion at South Tufa on Day 1.

References

Bursik, M., and Sieh, K., 1989, Range front faulting and volcanism in the Mono Basin, eastern California: *Journal of Geophysical Research*, v. 94, no. B11, p. 15587–15609.

- DeLano, K., Lee, J., Roper, R., and Calvert, A., 2019, Dextral, normal, and sinistral faulting across the eastern California shear zone–Mina deflection transition, California-Nevada, USA: *Geosphere*, v. 15, p. 1206–1239.
- Gilbert, C.M., Christensen, M.N., Al-Rawi, Y., and Lajoie, K.R., 1968, Structural and volcanic history of Mono Basin, California-Nevada, *in* Coats, R.R., et al., eds., *Studies in Volcanology*: Geological Society of America Memoir 116, p. 275–329.
- Hammond, W.C., Kreemer, C., and Blewitt, G., 2024, Robust imaging of fault slip rates in the Walker Lane and Western Great Basin from GPS data using a multi-block model approach: *Journal of Geophysical Research: Solid Earth*, v. 129, e2023JB028044.
- Hildreth, W., 2004, Volcanological perspectives on Long Valley, Mammoth Mountain, and Mono Craters: several contiguous but discrete systems: *Journal of Volcanology and Geothermal Research*, v. 136, p. 169–198.
- Lajoie, K.R., 1968, Late Quaternary stratigraphy and geologic history of Mono Basin, eastern California [Ph.D. dissertation]: Berkeley, University of California.
- Phillips, F.M., Zreda, M.G., Smith, S.S., Elmore, D., Kubik, P.W., and Sharma, P., 1990, Cosmogenic chlorine-36 chronology for glacial deposits at Bloody Canyon, eastern Sierra Nevada: *Science*, v. 248, p. 1529–1532.
- Reheis, M.C., Stine, S., and Sarna-Wojcicki, A.M., 2002, Drainage reversals in Mono Basin during the late Pliocene and Pleistocene: *Geological Society of America Bulletin*, v. 114, p. 991–1006.
- Russell, I.C., 1889, Quaternary history of Mono Valley, California: U.S. Geological Survey Eighth Annual Report, p. 261–394.
- Sharp, R.P., and Birman, J.H., 1963, Additions to classical sequence of Pleistocene glaciations, Sierra Nevada, California: *Geological Society of America Bulletin*, v. 74, p. 1079–1086.
- Sieh, K., and Bursik, M., 1986, Most recent eruption of the Mono Craters, eastern central California: *Journal of Geophysical Research*, v. 91, no. B12, p. 12539–12571.
- Stine, S., 1987, Mono Lake: The past 4,000 years [Ph.D. dissertation]: Berkeley, University of California.
- Stine, S., 1990, Late Holocene fluctuations of Mono Lake, eastern California: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 78, p. 333–381.
- Stine, S., 1994, Extreme and persistent drought in California and Patagonia during mediaeval time: *Nature*, v. 369, p. 546–549.
- Vazquez, J.A., and Lidzbarski, M.I., 2012, High-resolution tephrochronology of the Wilson Creek Formation (Mono Lake, California) and Laschamp event using ^{238}U - ^{230}Th SIMS dating of accessory mineral rims: *Earth and Planetary Science Letters*, v. 357–358, p. 54–67.
- Wesnousky, S.G., 2005, Active faulting in the Walker Lane: *Tectonics*, v. 24, TC3009.

Mono Basin Water Budget

Contributor: Scott Stine (Emeritus California State University, East Bay)

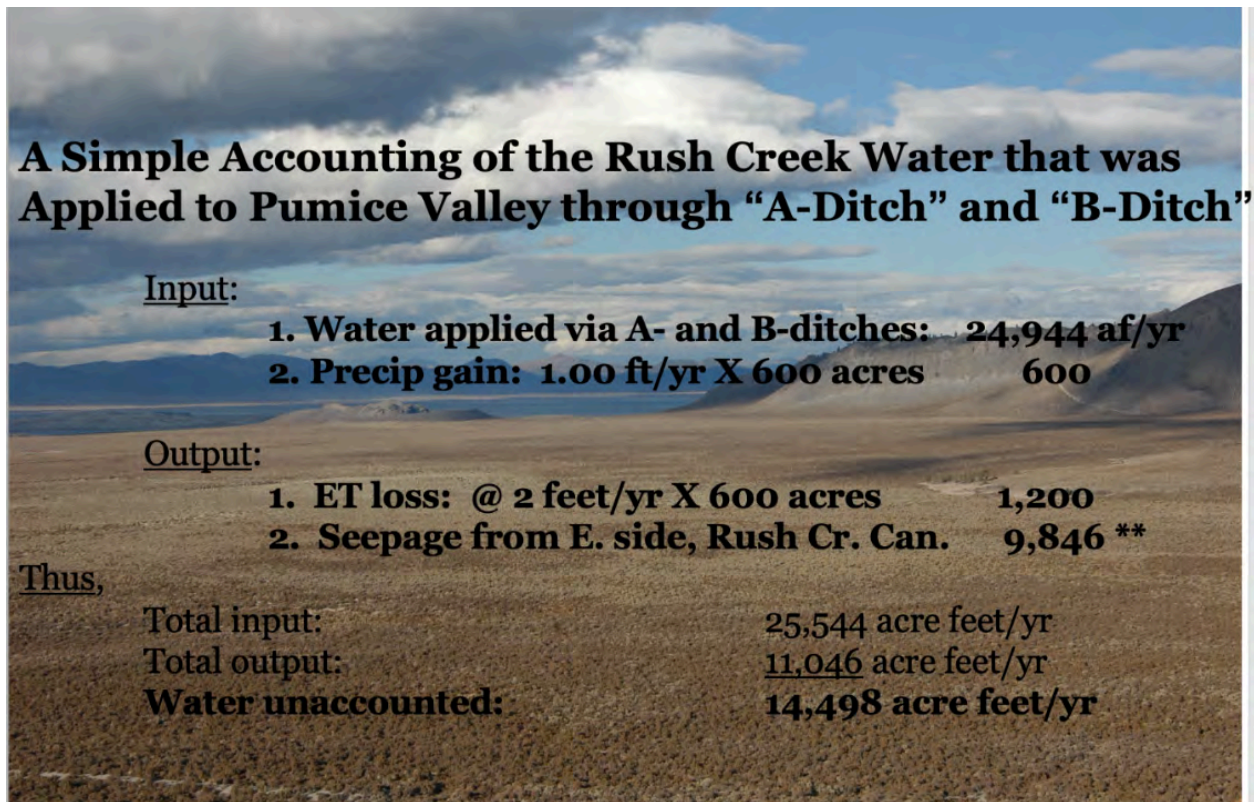


Figure S1.2 A Simple accounting of the Rush Creek water that was applied to Pumice Valley through the "A-Ditch" and "B-Ditch".

Day 1, S2: Panum Crater

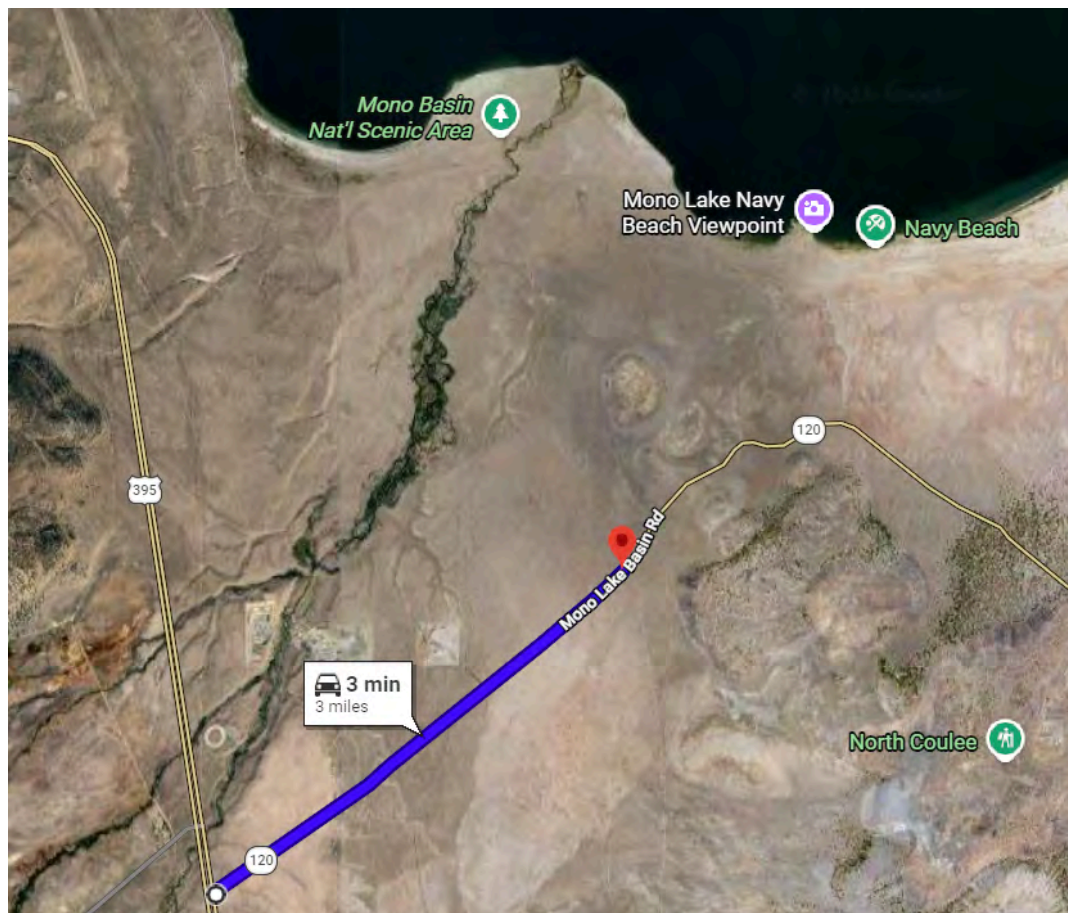
FIELD TRIP STOP CONTRIBUTORS: Genna Chiaro (U.S. Geological Survey), Lauren Harrison (Colorado State University), Scott Stine (Emeritus California State University, East Bay)

LOGISTICS: Park on the highway and hike to Panum Crater along the access roads. If accessibility by driving is needed please coordinate with organizers. The goals of this stop are to discuss the volcanic history of Mono Basin. We will also get a chance to hike around the rim of Panum dome, which erupted ~700 years ago during the most recent eruptive episode at Mono Craters. Volcanism in Mono lake will also be discussed.

WALKING DISTANCE: 2 miles across relatively level ground. There will be some areas with soft sandy ground and one relatively steep (~20%) decline and incline

LOCATION: Panum Crater (37.913965, -119.046577)

DIRECTIONS: [Hwy 395, Hwy 120 Junction to Panum Crater](#)



Regional Volcanism in the Mono Basin and Broader Long Valley Region

Contributor: Genna Chiaro (United States Geological Survey)

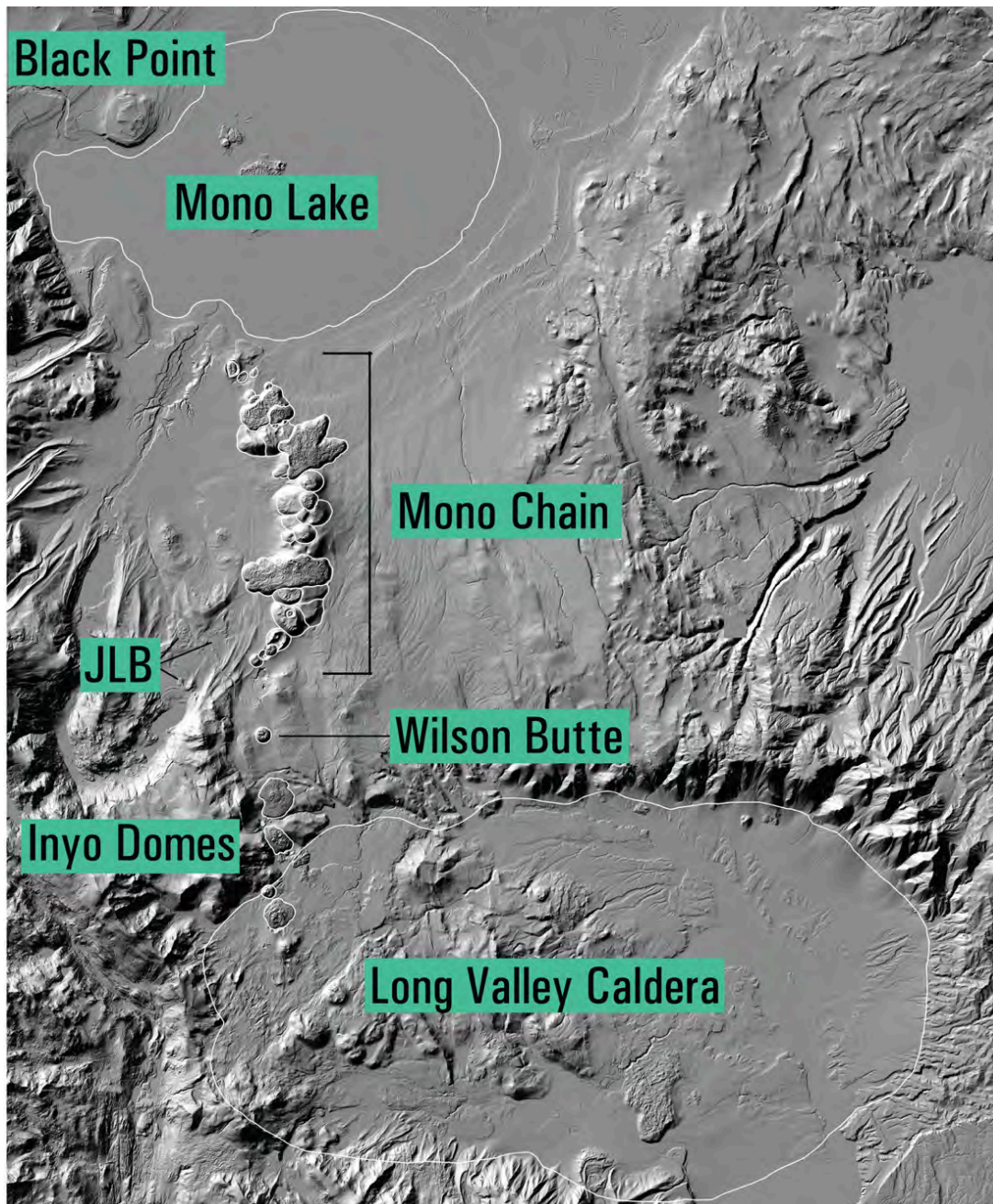


Figure S2.1: Regional map of the Long Valley volcanic field and Mono Basin. Mafic units near the Mono Craters are the Black Point scoria cone and June Lake basalt (JLB). The Bishop Tuff, which formed the Long Valley Caldera, erupted ~767 ka and is the largest eruption in California's history

The Mono Basin is part of the Long Valley region, an area with prodigious amounts of volcanic activity. Situated at the locus where Basin and Range extension meets the Sierra block, the Long Valley volcanic field and its eruptive products sit in an active transtensional zone (Bursik and Sieh, 1989; Hildreth et al., 2023). The first significant volcanic flareup occurred in the Pliocene and was associated with the onset of regional Basin and Range extension, erupting most of the basalt, andesite, and dacite lavas in the field between 3.9–2.6 Ma and ending shortly after initiation of the prominent Sierra Nevada range front fault (Bailey, 2004; Hildreth, 2004; Hildreth et al., 2023). After a ~400 kyr hiatus, the Long Valley volcanic field produced a total of ~850 km³ of rhyolite, ~700 km³ of which were high-silica rhyolite lavas and pyroclastic flows, from ~2.2–0.2 ka during eruption of the Glass Mountain rhyolites (~100 km³, 2.2–0.8 ka; Metz and Mahood 1985, 1991; Metz and Bailey, 1993), Bishop Tuff (~650 km³, 767 ka; Hildreth, 1979), and resurgent intracaldera rhyolites (~100 km³, 750–150 ka; Mankinen et al., 1986; Bailey, 1989; Hildreth et al., 2017). For a more detailed summary of the postcaldera rhyolites in the Long Valley Caldera, we refer the reader to Hildreth (2004), Hildreth et al., (2017), and Hildreth (2021).

Extending northward from the northwestern edge of the Long Valley Caldera to the southern shore of Mono Lake are the ~35 domes and flows of the Mono-Inyo volcanic chain (Fig. S2.1). The Inyo domes are at the southernmost end and include the north and south deadman creek domes, glass creek dome, and Obsidian Dome (~70–74 wt.% SiO₂; Sampson and Cameron, 1987). Wilson Butte (76.8 wt.% SiO₂) is just north of the Inyo domes and is somewhat of an outlier; it is not formally part of Mono Craters because it sits ~3 km to the south, but it is much more geochemically aligned with the Mono Craters than the Inyo domes (Hildreth, 2021). The Mono Craters include ~30 high-silica rhyolite (76.1–77 wt.% SiO₂) domes and one dacite (68 wt.% SiO₂) dome, dome 12 (Fig. S2.2). Dome 12 is thought to be older than all other exposed Mono domes due to the presence of tufa on its western slope. There is no tufa on the rhyolitic Mono domes, which suggests that dacitic dome 12 is the oldest lava dome at Mono Craters and may have been emplaced ~55–65 ka during the last high stand of Mono Lake (Zimmerman et al., 2006).

Direct and indirect dating of rhyolitic ash beds preserved in lake-silt beds of the informal Wilson Creek formation suggest that rhyolitic volcanism at the Mono Craters initiated at least as early as ~65 ka (Zimmerman et al., 2006; Vazquez and Lidzbarski, 2012; Ali et al., 2022). The oldest exposed rhyolitic Mono dome, however, is younger, dated at 42.2 ± 1.1 ka (2s, Marcaida et al., 2019). It is part of a sequence of biotite-bearing domes that range in age from 42–19 ka, all of which erupted along the dominant N-S trend of the Mono Domes (Marcaida et al., 2019). Biotite in these domes show increasing Fe content through time, likely reflecting fractionation from a single, melt-rich domain within the broader magmatic system (Kelleher and Cameron, 1990). Subsequent to the biotite-bearing domes and accompanying a change in mineralogy (i.e., biotite out, orthopyroxene in), orthopyroxene-bearing domes erupted 12–9 ka (Marcaida et al.,

2019). They contain mingled mafic enclaves, providing evidence for basalt at depth as the driver of Mono Craters volcanism (Kelleher and Cameron, 1990; Bray et al., 2017). No basaltic lavas, however, pierce the Mono chain; the only mafic eruptive units nearby are the June Lake basaltic andesites at its southern end and the basaltic scoria cone of Black Point north of Mono Lake (Fig. S2.1). The lack of mafic lavas between Black Point and the June Lake units suggests that there may be a “shadow zone” there created by a reservoir of silicic magma mush in the shallow crust (Hildreth, 2021).

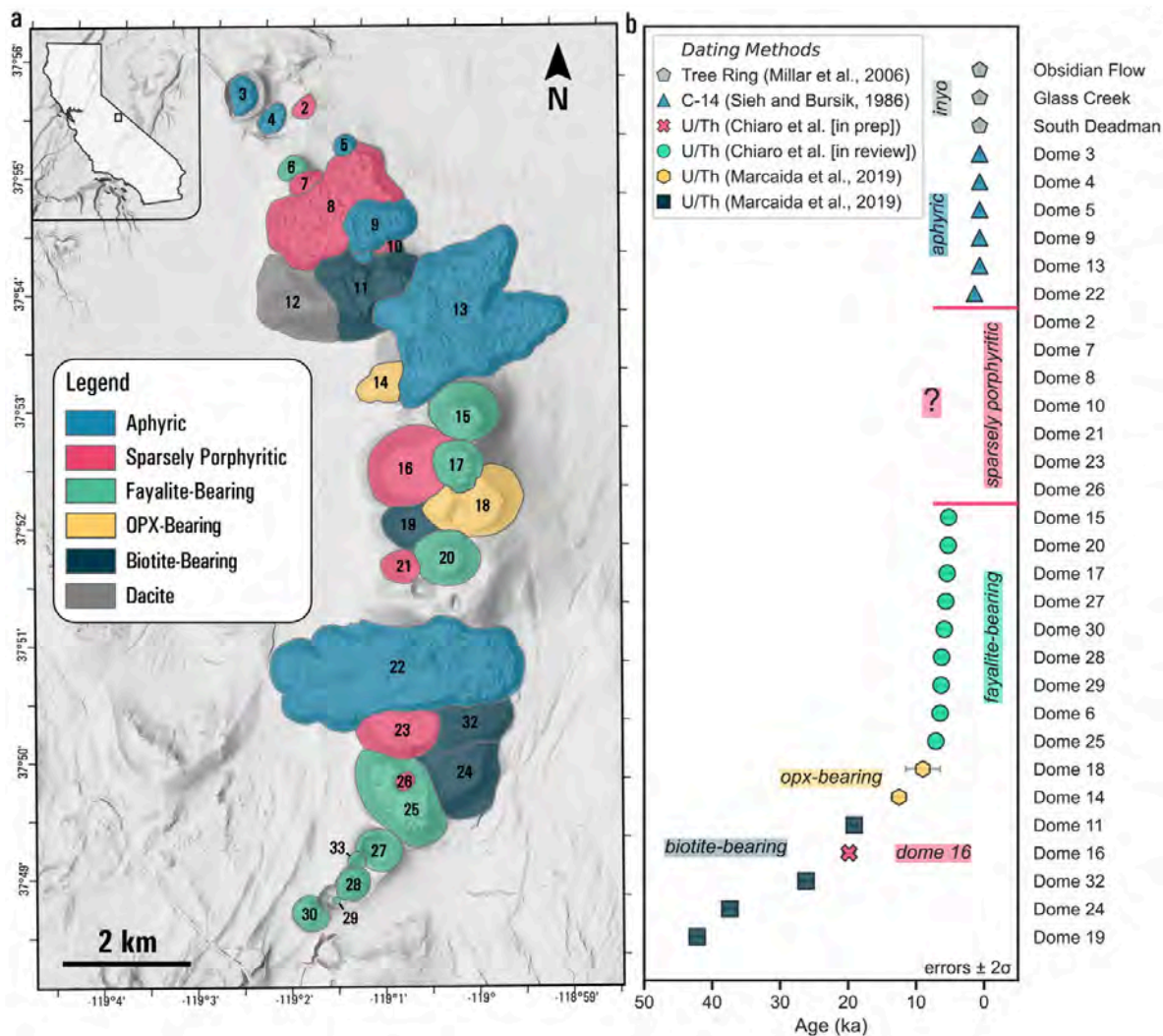


Figure S2.2. Geologic map of the Mono Craters and the Pleistocene-Holocene eruptive history of the Mono-Inyo domes. (a) A map showing the Mono Craters chain with numbering from Wood (1983) and colors based on groupings from Kelleher and Cameron (1990), subdividing domes by crystal content and diagnostic mineral phases. (b) Existing eruption ages for the Mono-Inyo craters dated using dendrochronology (Millar et al., 2006), radiocarbon dating (Sieh and Bursik, 1986), and ^{238}U - ^{230}Th geochronology (zircon and allanite crystal surfaces; Marcaida et al., 2019; Chiaro et al. (in review, in prep)).

It is estimated that there have been more than 10 eruptions along the Mono-Inyo chain in the past 4,000 yrs (Sieh and Bursik, 1986), but the only Holocene Mono lavas with well-constrained ages are the north Mono domes (3–5, 9, 13) and the south coulee (22; Fig. S2.2); these have been dated using radiocarbon and tephrostratigraphy (Bursik and Sieh, 2013). The south coulee (~1.4 ka) erupted ~0.32 km³ (DRE) of lava and marks the first eruption of aphyric rhyolite; since then, all subsequent eruptions at Mono Craters have been aphyric (Bursik and Sieh, 1989; Hildreth, 2004; Bursik et al., 2014). North Mono (~0.7 ka) produced ~0.07 km³ (DRE) of pyroclastic ash fall and flow deposits and five separate aphyric lavas (~0.44 km³; DRE) during the most recent eruptive episode at Mono Craters (Sieh and Bursik, 1986).

References

- Ali, G. A. H., Lin, K., Hemming, S. R., Cox, S. E., Ruprecht, P., Zimmerman, S. R. H., Stine, S., & Wang, X. (2022). Emergence of wet conditions in the Mono Basin of the Western USA coincident with inception of the Last Glaciation. *GSA Bulletin*, 134(9–10), 2267–2279. <https://doi.org/10.1130/B36084.1>
- Bailey, R. (2004). Eruptive History and Chemical Evolution of the Precaldera and Postcaldera Basalt-Dacite Sequences, Long Valley, California: Implications for Magma Sources, Current Seismic Unrest, and Future Volcanism (Professional Paper) [Professional Paper]. <https://doi.org/https://doi.org/10.3133/pp1692>
- Bailey, R. A. (1989). Geologic map of Long Valley caldera, Mono Inyo Craters volcanic chain, and vicinity, eastern California. U.S. Geol. Surv. Map I-1933, scale 1:62,500; pamphlet [Map]. U.S. Geological Survey Miscellaneous Investigations.
- Bursik, M., & Sieh, K. (1989). Range front faulting and volcanism in the Mono Basin, eastern California. *Journal of Geophysical Research: Solid Earth*, 94(B11), 15587–15609. <https://doi.org/10.1029/JB094iB11p15587>
- Bursik, M., & Sieh, K. (2013). Digital database of the Holocene tephtras of the Mono-Inyo craters [U.S. Geological Survey Digital Data Series]. <https://doi.org/https://doi.org/10.3133/ds758>
- Bursik, M., Sieh, K., & Meltzner, A. (2014). Deposits of the most recent eruption in the Southern Mono Craters, California: Description, interpretation and implications for regional marker tephtras. *Journal of Volcanology and Geothermal Research*, 275, 114–131. <https://doi.org/10.1016/j.jvolgeores.2014.02.015>
- Bray B, Stix J, Cousens B (2017) Mafic replenishment of multiple felsic reservoirs at the Mono domes and Mono Lake islands, California. *Bull Volcanol* 79:54. <https://doi.org/10.1007/s00445-017-1123-y>
- Chiaro, G., Vazquez, J., Fierstein, J., Calvert, A., Hurwitz, S., Robinson, J. An extensive and short-lived episode of mid-Holocene fayalite-bearing high-silica rhyolite eruptions at Mono Craters, California. *Bulletin of Volcanology* (in review)
- Hildreth, W. (1979). The Bishop Tuff: Evidence for the origin of compositional zonation in silicic magma chambers. In *Geological Society of America Special Papers* (Vol. 180, pp. 43–76). Geological Society of America. <https://doi.org/10.1130/SPE180-p43>

- Hildreth, W. (2004). Volcanological perspectives on Long Valley, Mammoth Mountain, and Mono Craters: Several contiguous but discrete systems. *Journal of Volcanology and Geothermal Research*, 136(3–4), 169–198. <https://doi.org/10.1016/j.jvolgeores.2004.05.019>
- Hildreth, W. (2021). Comparative Rhyolite Systems: Inferences From Vent Patterns and Eruptive Episodicities: Eastern California and Laguna del Maule. *Journal of Geophysical Research: Solid Earth*, 126(7), e2020JB020879. <https://doi.org/10.1029/2020JB020879>
- Hildreth, W., Fierstein, J., & Calvert, A. (2017). Early postcaldera rhyolite and structural resurgence at Long Valley Caldera, California. *Journal of Volcanology and Geothermal Research*, 335, 1–34. <https://doi.org/10.1016/j.jvolgeores.2017.01.005>
- Hildreth, W., Fierstein, J., & Calvert, A. T. (2023). Precaldera Mafic Magmatism at Long Valley, California: Magma-Tectonic Siting and Incubation of the Great Rhyolite System. *Journal of Volcanology and Geothermal Research*, 433, 107726. <https://doi.org/10.1016/j.jvolgeores.2022.107726>
- Kelleher PC, Cameron KL (1990) The Geochemistry of the Mono Craters-Mono Lake Islands Volcanic Complex, eastern California. *J Geophys Res Solid Earth* 95:17643–17659. <https://doi.org/10.1029/JB095iB11p17643>
- Marcaida, M., Vazquez, J. A., Stelten, M. E., & Miller, J. S. (2019). Constraining the Early Eruptive History of the Mono Craters Rhyolites, California, Based on 238 U- 230 Th Isochron Dating of Their Explosive and Effusive Products. *Geochemistry, Geophysics, Geosystems*, 20(3), 1539–1556. <https://doi.org/10.1029/2018GC008052>
- Metz, J. M., & Bailey, R. A. (1993). Geologic map of Glass Mountain, Mono County, California. <https://doi.org/https://doi.org/10.3133/i1995>
- Metz, J. M., & Mahood, G. A. (1985). Precursors to the Bishop Tuff Eruption: Glass Mountain, Long Valley, California. *Journal of Geophysical Research: Solid Earth*, 90(B13), 11121–11126. <https://doi.org/10.1029/JB090iB13p11121>
- Metz, J. M., & Mahood, G. A. (1991). Development of the Long Valley, California, magma chamber recorded in precaldern rhyolite lavas of Glass Mountain. *Contributions to Mineralogy and Petrology*, 106(3), 379–397. <https://doi.org/10.1007/BF00324565>
- Millar CI, King JC, Westfall RD, et al (2006) Late Holocene forest dynamics, volcanism, and climate change at Whitewing Mountain and San Joaquin Ridge, Mono County, Sierra Nevada, CA, USA. *Quat Res* 66:273–287. <https://doi.org/10.1016/j.yqres.2006.05.001>
- Sampson, D. E., & Cameron, K. L. (1987). The geochemistry of the Inyo Volcanic Chain: Multiple magma systems in the Long Valley Region, eastern California. *Journal of Geophysical Research: Solid Earth*, 92(B10), 10403–10421. <https://doi.org/10.1029/JB092iB10p10403>
- Sieh, K., & Bursik, M. (1986). Most recent eruption of the Mono Craters, eastern central California. *Journal of Geophysical Research: Solid Earth*, 91(B12), 12539–12571. <https://doi.org/10.1029/JB091iB12p12539>
- Vazquez, J. A., & Lidzbarski, M. I. (2012). High-resolution tephrochronology of the Wilson Creek Formation (Mono Lake, California) and Laschamp event using 238U-230Th SIMS dating of accessory mineral rims. *Earth and Planetary Science Letters*, 357–358, 54–67. <https://doi.org/10.1016/j.epsl.2012.09.013>
- Wood S (1983) Chronology of Late Pleistocene and Holocene Volcanics, Long Valley and Mono Basin Geothermal Areas, Eastern California. Boise State University

Zimmerman, S. H., Hemming, S. R., Kent, D. V., & Searle, S. Y. (2006). Revised chronology for late Pleistocene Mono Lake sediments based on paleointensity correlation to the global reference curve. *Earth and Planetary Science Letters*, 252(1–2), 94–106. <https://doi.org/10.1016/j.epsl.2006.09.030>

Age Constraints on the Formation of Paoha Island

Contributor: Lauren Harrison (Colorado State University)

This stop gives an excellent view of the islands within Mono Lake, Paoha and Negit Islands, which are both volcanically formed. The Mono Lake area has been volcanically active in the late Holocene, although the exact age of much of the recent activity is unconstrained by direct radiometric dates. At ~17 ka, basalt erupted explosively within Mono Lake to form Black Point; then at ~700 years ago high-silica rhyolite erupted to form Panum Crater on the south shore of Mono Lake; and possibly more recently, rhyolites and trachytes erupted within Mono Lake to form Paoha and Negit Islands and related islets (Bailey, 1989).

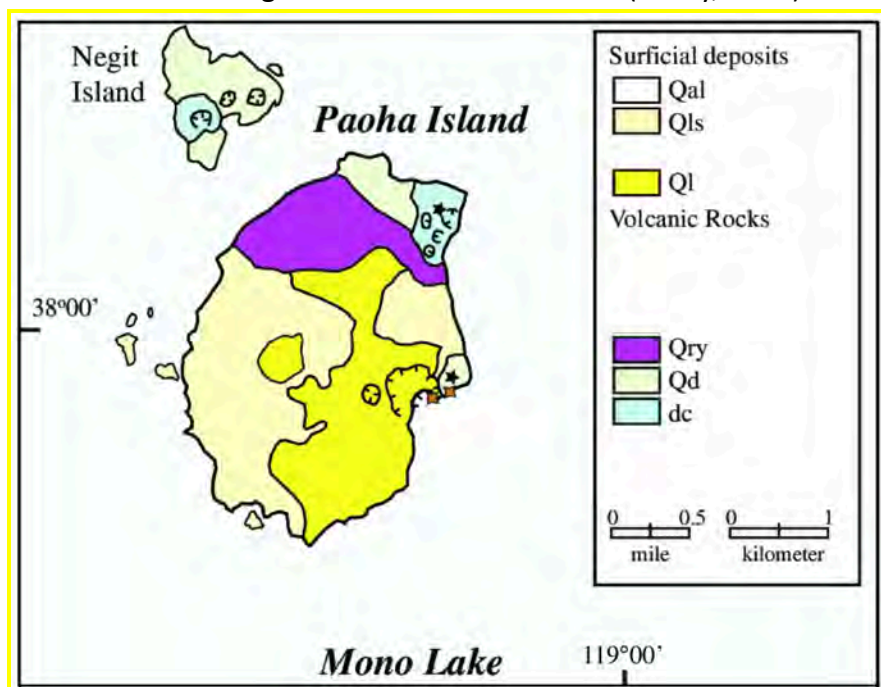


Figure S2.3. Generalized geologic map modified from Bergfeld et al., 2015 after Bailey, 1989. Volcanic rocks are as follows: Qry - rhyolite, Qd - trachydacite, dc - subaerial trachyte

Formation of Paoha Island

Biotite-bearing, sparsely porphyritic rhyolite erupted into Mono Lake sediments <2000 years ago, as constrained by the presence of tephra of known ages in sediments that mantle portions of the rhyolite. The tephra include the 0.7-ka North Mono-Inyo, the 1.3 ka South Mono, and either the 1.6 ka Northwest Coulée or the 1.7 ka Wilson Butte tephra. Previous fieldwork interpreted the sediment-and-tephra sequences as deposited within Mono Lake prior to intrusion of the rhyolite, which then subsequently slid off haphazardly with uplift of the Island (Coleman et al., 2014). It is possible that some tephra found on Paoha Island, especially the dacite tephra that have no extrusive equivalents south of Mono Lake along the Mono-Inyo

chains, post-date volcanic units that form Paoha Island. It is also unknown whether the eruption of the rhyolite built the island above lake level or if later magmatic intrusions uplifted the island. Following eruption of the rhyolite, two trachyte eruptions occurred: one that records contact with wet sediment, and another that was subaerial (Bray et al., 2017).

New Geochronology of Paoha Volcanic Units

Preliminary cosmogenic isotope exposure ages have been analyzed (L. Harrison, J. Licciardi) to investigate the timing of eruption and related deformation on Paoha Island. One sample was collected from the Paoha rhyolite and the subaerial trachyte flow for ^{36}Cl exposure dating. Application of this dating method to these features is challenging due to the aphyric and glassy nature of the lavas, moderate native Cl concentrations (~220-250 ppm), and the young age of the volcanic rocks. However, recent advances in ^{36}Cl measurement methodologies now offer the potential for reliable dating of young volcanic rocks (Lesnek et al., 2024; Schimmelpfennig et al., 2011).

Paoha Rhyolite



Figure S2.4. A knob of rhyolite on a ridge that was sampled for cosmogenic exposure dating.



Figure S2.5. The knob of rhyolite that was sampled for cosmogenic exposure dating was in situ and composed of columnar jointed rhyolite. The top 3 cm of rhyolite that was at least 4 cm from an edge was collected and cut for analysis. Hammer for scale.

Paoha Subaerially Erupted Trachyte



Figure S2.6. Panoramic view of the northern subaerial trachyte flow. Evidence that the flow was erupted subaerially is the spiny a'a texture of the lava flow, apparent in this photo (the cobbled foreground is a pre-existing eruption that likely experienced wave erosion when either the Island was at a lower elevation or the lakeshore of Mono Lake was higher).

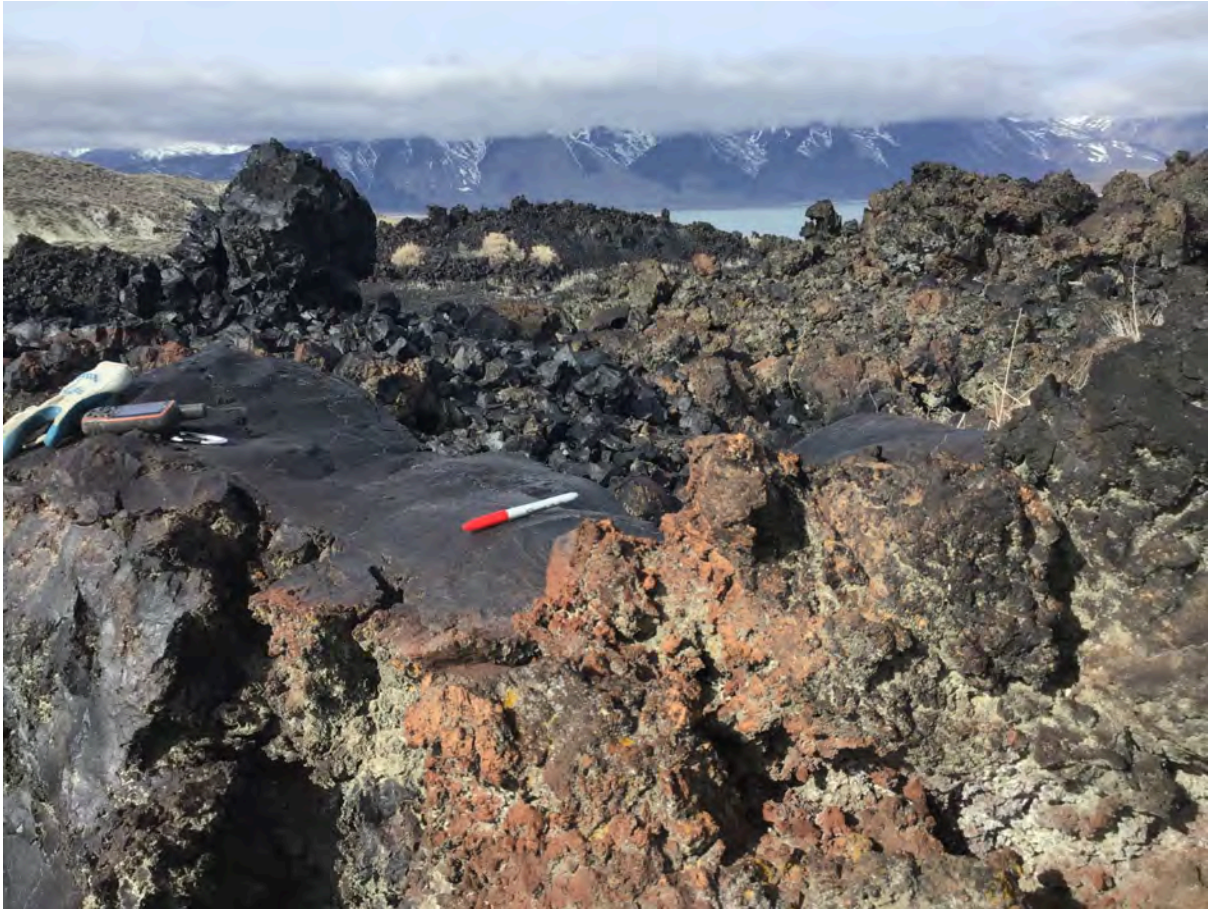


Figure S2.7. The smooth-topped a'a spine in the subaerial trachyte flow that was sampled for cosmogenic exposure dating. Sharpie for scale.

Preliminary Cosmogenic Exposure Ages of Paoha lavas

Unpublished preliminary cosmogenic exposure ages of Paoha island rhyolite is 760 ± 40 years and of the Paoha Island trachyte is 620 ± 30 years (Lauren Harrison, Joseph Licciardi), establishing the presence of Paoha Island above lake level by ~ 760 years ago.

Cosmogenic exposure dates have the advantage of being direct dates of the time that a lava was exposed on the surface of the Earth, usually interpreted as the eruption age. Here, the subaerially erupted trachyte cosmogenic exposure age is unambiguously interpreted as the eruption age of the trachyte. The sample collected for dating was *in situ*, not affected by major topographic shielding, and collected from a smooth, flat surface necessary for a reliable cosmogenic exposure age. On the other hand, the rhyolite is interpreted to erupt into lake bottom sediments, uplifting the Island above lake level. Cosmogenic nuclides measured for calculating an exposure age will only start accumulating once the rock is covered by less than

one meter of water, and more when it is exposed above lake level. It is unknown if the rhyolite erupted synchronously with uplift of Paoha Island above lake level; field exposures of the rhyolite are more eroded than the subaerial trachyte and characterized by volcanic features such as quenched glass and formation of columnar joints that may be more characteristic of a large intrusion into cool sediments that quenched the carapace while insulating the interior of the flow enough to allow for columns to form (however, there was no peperite observed in exposures of the rhyolite). Therefore our exposure age is best interpreted as the age of uplift of Paoha Island above lake level.



Figure S2.8. Quench glass on some exposures of rhyolite. Hammer for scale.

References

Bailey, R. A. (1989). Geologic map of Long Valley caldera, Mono Inyo Craters volcanic chain, and vicinity, eastern California. U.S. Geol. Surv. Map I-1933, scale 1:62,500; pamphlet [Map]. U.S. Geological Survey Miscellaneous Investigations.

- Bergfeld, D., Evans, W.C., Howle, J.F., Hunt, A.G. (2015). Magmatic gas emissions at Holocene volcanic features near Mono Lake, California, and their relation to regional magmatism. *Journal of Volcanology and Geothermal Research*, 292, 70-83.
- Bray B, Stix J, Cousens B (2017) Mafic replenishment of multiple felsic reservoirs at the Mono domes and Mono Lake islands, California. *Bull Volcanol* 79:54. <https://doi.org/10.1007/s00445-017-1123-y>
- Colman, S.M., Hemming, S.R., Stine, S., Zimmerman, S.R.H. (2014). The effects of recent uplift and volcanism on deposition in Mono Lake, California, from seismic-reflection (CHIRP) profiles. *Journal of Geophysical Research:Solid Earth*, 119, 3955-3970.
- Lesnek, A., Licciardi, J., Hidy, A., Anderson, T. (2024) Technical note: Optimizing the in situ cosmogenic ^{36}Cl extraction and measurement workflow for geologic applications. *Geochronology*, 6(3), 475-489.
- Schimmelpfennig, I., Benedetti, L., Garreta, V., Pik, R., Blard, P.-H., Burnard, P., Bourles, D., Finkel, R., Ammon, K., Dundai, T. (2011) Calibration of cosmogenic ^{36}Cl production rates from Ca and K spallation in lava flows from Mt. Etna and Payun Matru. *Geochemical et Cosmochimica Acta*, 75(10), 2611-2632.

Age and Mechanisms of Volcanism and Deformation in Mono Lake, California

Contributor: Zachary Smith (University of California, Berkeley), Kathleen Rodrigues (Desert Research Institute, University of Nevada, Reno), Michael Manga (University of California Berkeley), Natalie Homyk (University of California, Berkeley), Matthew Hornbach (Southern Methodist University)

The islands in the center of Mono Lake, are among the youngest volcanic landforms in California and lie at the northern end of the Mono-Inyo Craters volcanic chain (Stine, 1987). Negit Island is built of dacite domes and lava flows, and similar dacite cones and flows form the northeast corner of Paoha Island. Most of Paoha Island, by contrast, consists of pale lake sediments that were lifted off the lake floor by a shallow intrusion. From radiocarbon-dated shorelines around the lake, Stine (1987) proposed that the oldest of the island dacites is slightly younger than 2,000 yr B.P., that the youngest postdates 220 yr B.P., and that uplift of Paoha Island and extrusion of the associated rhyolite may have occurred between A.D. 1720 and 1850 (Stine, 1987).

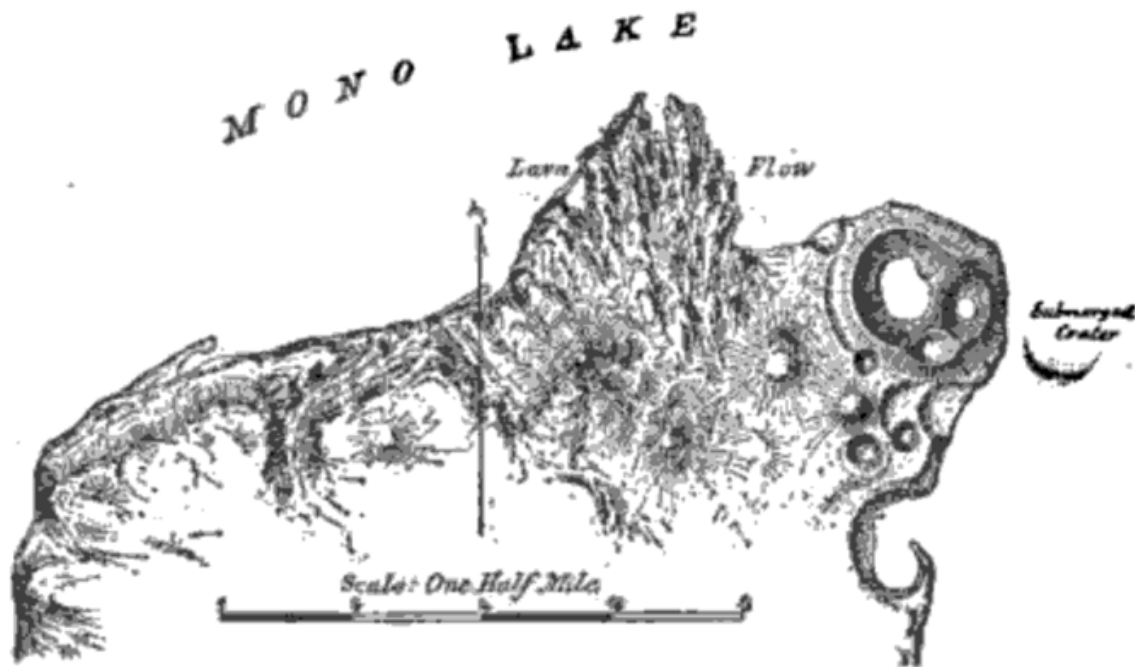


FIG. 10.—North end of Paoha Island.

Figure S2.9. Dacite flows and cones and rhyolite cryptodome on the northern end of Paoha Island as documented by Russell (1889).

The islands have drawn attention since the first systematic survey of the basin. Russell (1889) sketched the north end of Paoha Island, including its lava flow, a cluster of craters, and a

submerged crater just offshore (Fig. S2.9). Hot springs on Paoha were noted early as well; Waring (1915) reported spring temperatures of about 80 °C (176 °F) when the lake stood high. Today, fumaroles and hot springs remain active at Hot Springs Cove.

Geophysical surveys show that the islands sit above an active magmatic and hydrothermal system. Three-dimensional magnetotelluric imaging of Mono Basin reveals conductive zones interpreted as crystal–melt mush beneath the Mono Craters and conductive pathways that link magma sources to Holocene vents (Peacock et al., 2015). A high-resolution aeromagnetic survey of Mono Basin and Long Valley (Ponce, 2020) shows a strong positive anomaly centered on Negit Island and the northern part of Paoha Island, which we associate with the dacitic volcanism in the lake (Fig. S2.10). A 2021 lake-floor heat flow survey by Southern Methodist University, UC Berkeley, and the USGS California Volcano Observatory recorded some of the highest sediment temperatures just east and south of Paoha Island.

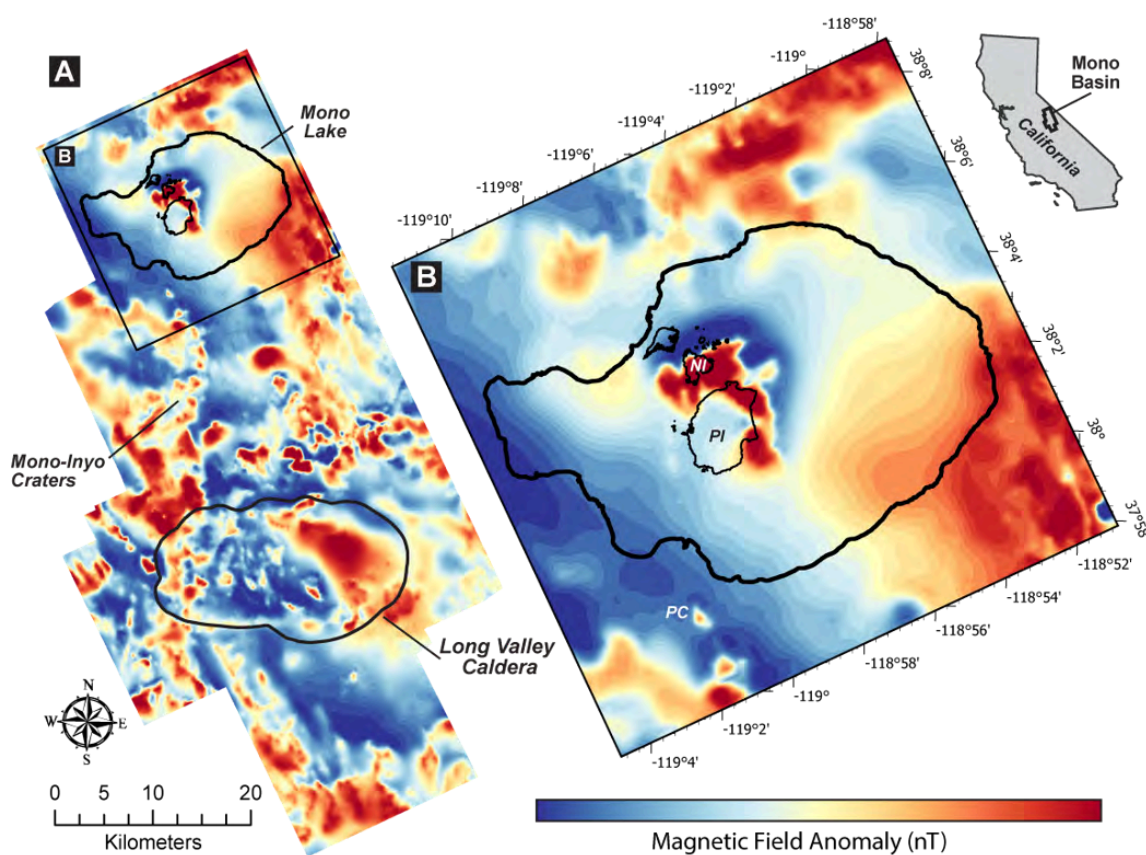


Figure S2.10. Magnetic survey covering Long Valley and Mono Basin. Magnetic data reprocessed from Ponce et al. (2019). Magnetic anomalies are associated with the dacitic volcanism in Mono Lake.

Active Volcanic Deformation

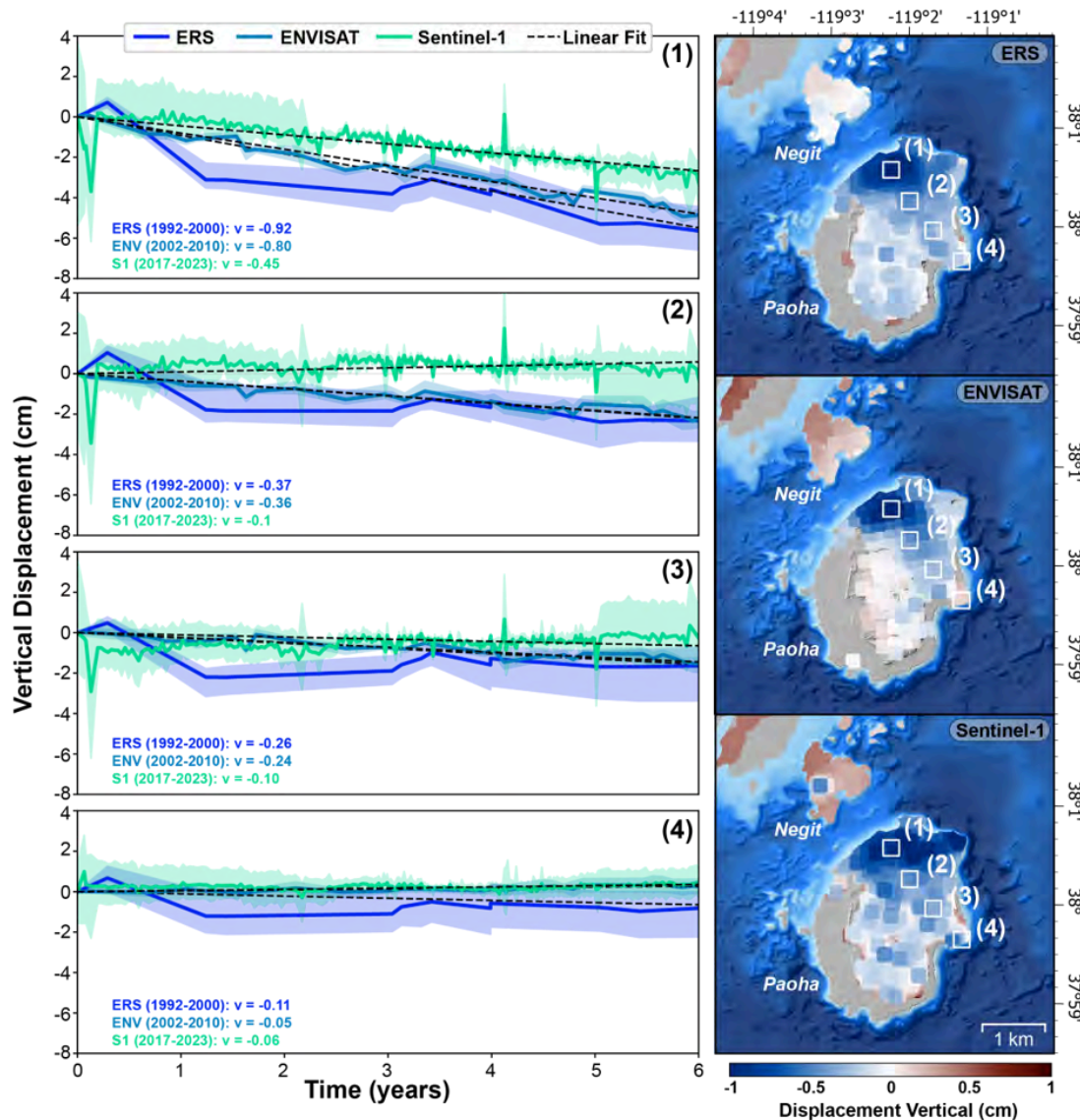


Figure S2.11. InSAR displacement observations from Paoha Island showing ongoing subsidence on the northern third of Paoha Island over the last 34 years. Subsidence appears to be slowing.

Volcanic islands experience periods of uplift and subsidence driven by magma migration, flank instability, volcanic loading, tectonic extension, and cooling of a magma intrusion or hydrothermal system, but constraining specific mechanisms is challenging. We couple new luminescence dating, InSAR, LiDAR, and heat flow observations with thermo-hydro-mechanical modeling (Figs. S2.11) to investigate the age of Paoha Island and mechanisms driving subsidence

on the island. InSAR analysis (1992-present) shows that the northern third of the island is subsiding at 1 cm/yr (Fig. S2.11), but the mechanisms driving subsidence and whether it extends offshore are unknown. LiDAR analysis reveals that the highest shoreline, formed 106 ybp during the 1919 historic highstand, has maintained its elevation indicating that active deformation is confined to the interior of the island or started in the last few decades. Given the localized extent of subsidence, only magma migration or cooling of a magma intrusion or hydrothermal system can produce the observed spatial distribution of subsidence. Inversion of InSAR and shoreline surface displacements suggests that the subsidence source is at ~ 1 km depth; shallower than possible magma bodies inferred from magnetotelluric studies at 2-5 km (Peacock et al., 2015). Cooling of a magma intrusion would require high cooling rates and a young intrusion age of 100-150 ybp (Figs. S2.13). In contrast, cooling and water vapor condensation in a hydrothermal system at 100-1100 m depth could produce highly localized subsidence. However, our new heat flow survey shows no anomaly around the subsiding zone. Heat flow and magnetic surveys (Fig. S2.10) in conjunction with modeling suggest that the subsiding zone could be hydraulically connected to the Hot Springs Cove fumarole field, 1.2 km to the southeast (Fig. S2.13), where the heat emerging may be advected laterally along a fault from a cooling hydrothermal system. This study demonstrates the importance of geodetic, geomorphic, and geophysical observations for constraining the mechanisms of active deformation on volcanic islands.

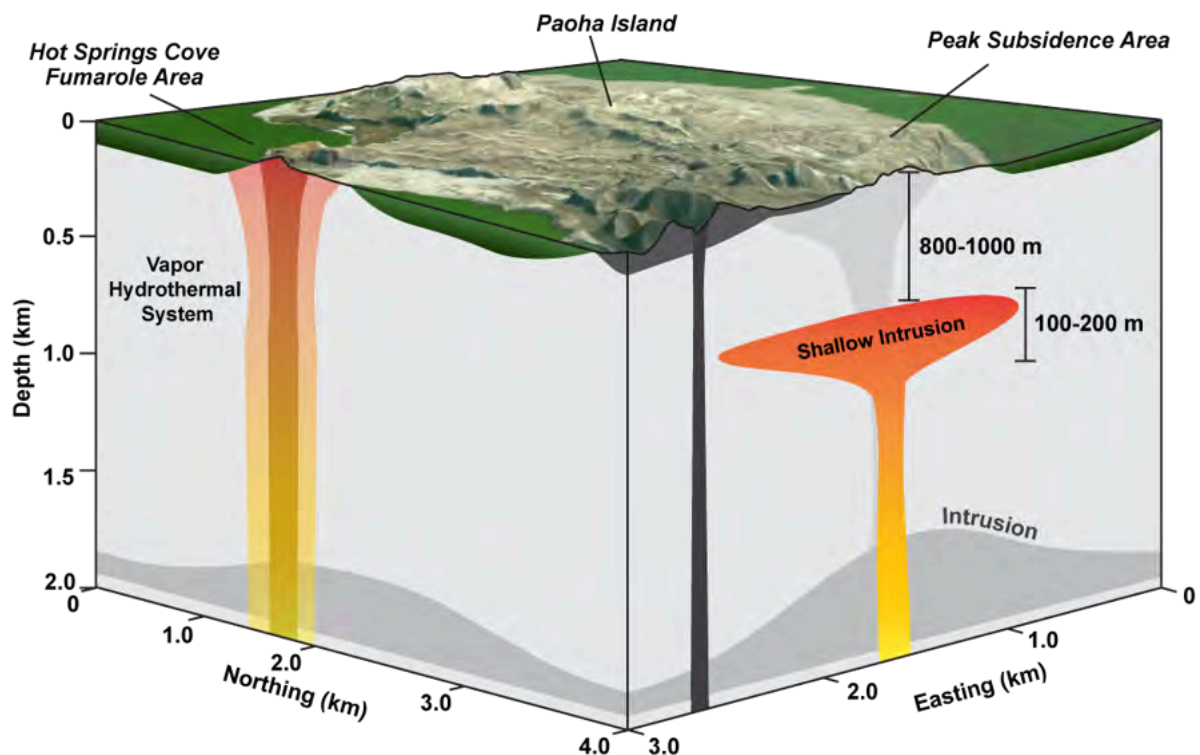


Figure S2.12. Conceptual model for double intrusion. Older intrusion occurred around 500-700 ybp and a younger intrusion later form at around 100-150 ybp. Cooling of the younger intrusion may be the source of the ongoing subsidence on the northern side of Paoha Island.

Luminescence Ages of the Island Volcanics

Luminescence dating was used to estimate the timing of volcanic activity on Paoha and Negit Islands. Samples were collected from lava flows and pyroclastic bombs using the same general approach previously applied at the Soda Lake Volcanic Center in northwestern Nevada (Rodrigues et al., 2026). Intact blocks were removed from the outcrop and later cored through their centers under controlled laboratory lighting at the DRI Luminescence Research Laboratory. Using material from the protected interior minimizes the possibility of exposing the dated material to sunlight during sampling. Tephra was collected using a more conventional sediment-sampling approach, by hammering an opaque tube horizontally into a freshly cleaned exposure.

Luminescence dating takes advantage of energy that becomes trapped within mineral grains through exposure to naturally occurring radiation. During a volcanic eruption, the pre-existing luminescence signal can be reset by intense heating and, particularly for fragmented or transported material such as tephra, potentially by a combination of heat, sunlight exposure, and mechanical stress. Once the volcanic material cools and is emplaced or buried, the luminescence signal begins to accumulate again. The amount of stored energy (the equivalent dose) is measured in the laboratory and compared with the rate at which the sample receives radiation from its surroundings (the dose rate), providing an estimate of the time elapsed since eruption and emplacement. Incomplete resetting can result in residual luminescence and an apparent age that is older than the actual eruption.

Here we provide preliminary results from tephra on Paoha Island that yielded an apparent age of 8.4 ± 0.9 ka. This tephra may be a mix of material from multiple eruptions and therefore represents an old average age of multiple eruptions. A pyroclastic bomb from within the same deposit yielded 0.49 ± 0.07 ka. The pyroclastic bomb has the same chemistry as the northern dacite flow on Paoha Island (probably the lava flow sketched by Russell, 1889; Fig. S2.9). The dacite flow gives an age of 0.62 ± 0.09 ka which is in excellent agreement with cosmogenic isotope exposure ages (Harrison and Licciardi, this field guide). At Negit the central dome was dated at 1.2 ± 0.2 ka which is in agreement with age estimates from Stine (1987).

Models for the Subsidence

Active subsidence has been observed on Paoha Island since interferometric synthetic aperture radar was developed in the early 1990's (Tizzani et al. 2007). However, the mechanism(s) driving subsidence are still unknown. Possible mechanisms include a landslide, compaction under a volcanic load, magma migration, cooling hydrothermal system, or cooling magma intrusion. New LiDAR surveys show that the shorelines on the northern end of Paoha Island are not deformed, indicating the most of the active subsidence is confined to the interior of the island.

This suggests that a landslide and or compaction are not the mechanism because both would have deformed the shorelines.

The new luminescence ages also bear directly on the competing explanations for the subsidence. The 0.49-0.62 ka ages for the northern dacite and its pyroclastic bomb fall within the 500–750 ybp window that we now confirm is a minimum age for the emergence of Paoha Island. However, an intrusion from around 500-750 ybp would not likely still be causing subsidence today and the geometry of the subsiding zone does not correspond to the size of the island. This suggests that there may have been a younger intrusion or a cooling localized vapor phase hydrothermal system responsible for the ongoing subsidence.

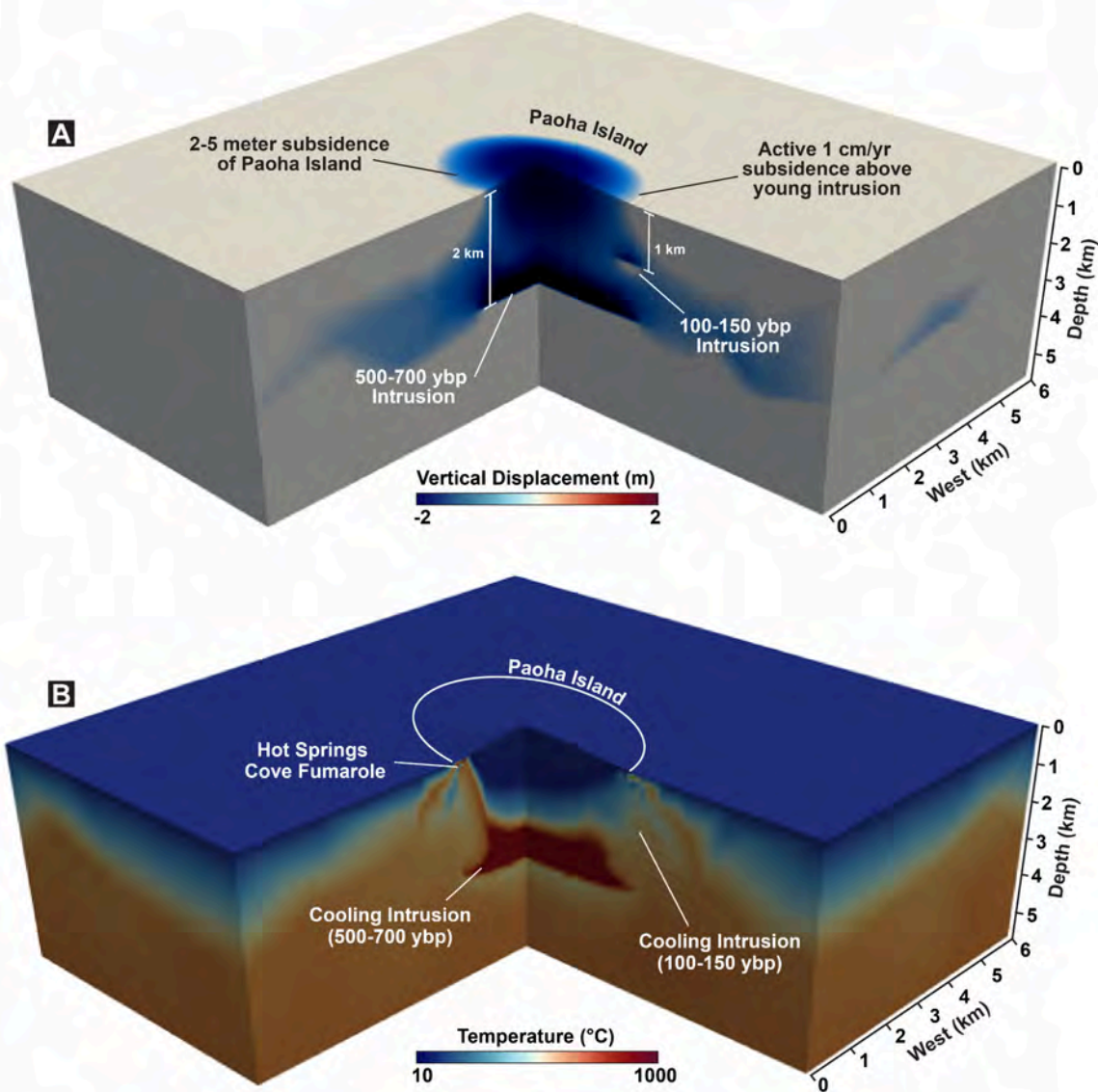


Figure S2.13. 3D thermo-hydro-mechanical (THM) finite element model for double intrusion. (A) Vertical displacement showing subsidence at Paoha Island due to cooling of magma intrusions. (B) Temperature

Questions for Discussion

(1) How do we explain the lack of old shorelines on Paoha Island? (2) What is the mechanism driving deformation today? (3) What structure could link the subsiding zone to the Hot Springs Cove fumaroles, 1.2 km away, while leaving no heat flow anomaly above the subsiding zone itself?

References

- Bailey, R.A., 1989, Geologic map of Long Valley caldera, Mono-Inyo Craters volcanic chain, and vicinity, eastern California: U.S. Geological Survey Miscellaneous Investigations Series Map I-1933, scale 1:62,500, 11 p., <https://doi.org/10.3133/i1933>.
- Harrison, L, and Licciardi, [initials], 2026, [title of chapter], in [editors], Friends of the Pleistocene 2026 Mono Basin Field Trip Guidebook, p. [pages].
- Peacock, J.R., Mangan, M.T., McPhee, D., and Ponce, D.A., 2015, Imaging the magmatic system of Mono Basin, California, with magnetotellurics in three dimensions: *Journal of Geophysical Research: Solid Earth*, v. 120, no. 11, p. 7273–7289, <https://doi.org/10.1002/2015JB012071>.
- Ponce, D.A., 2020, High-resolution aeromagnetic survey of Mono Basin and Long Valley, California: U.S. Geological Survey data release, <https://doi.org/10.5066/P9671UZU>.
- Rodrigues, K., Ruprecht, P. P., and de Polo, C. M.: Luminescence Dating Reveals Multiple Late Quaternary Eruptions at the Soda Lake Volcanic Center, Nevada, *EGU sphere* [preprint], <https://doi.org/10.5194/egusphere-2026-4195>, 2026.
- Russell, I.C., 1889, Quaternary history of Mono Valley, California: U.S. Geological Survey Eighth Annual Report, pt. 1, p. 261–394.
- Smith, Z., Manga, M., Hornbach, M., Lindsay, D., Homyk, N., DeMarines, J., and Bürgmann, R., 2024, Constraints on the distribution and mechanism of active subsidence on Paoha Island in Mono Lake, California: *Geological Society of America Abstracts with Programs*, v. 56, no. 5, paper 157-3, <https://doi.org/10.1130/abs/2024AM-403208>.
- Stine, S., 1987, Mono Lake: The past 4,000 years [Ph.D. dissertation]: Berkeley, University of California, 615 p.
- U.S. Geological Survey, 2021, New survey plumbs Mono Lake's depths for hydrothermal heat: U.S. Geological Survey News, <https://www.usgs.gov/news/new-survey-plumbs-mono-lakes-depths-hydrothermal-heat>
- Waring, G.A., 1915, Springs of California: U.S. Geological Survey Water-Supply Paper 338.

Volcanic History and Recent Observations of Climate Change

Contributor: Scott Stine (Emeritus California State University, East Bay)

The Mono Basin is part of the Long Valley region, an area with prodigious amounts of volcanic activity. Situated at the locus where Basin and Range extension meets the Sierra block, the Long Valley volcanic field and its eruptive products sit in an active transtensional zone (Bursik and Sieh, 1989; Hildreth et al., 2023). The first significant volcanic flareup occurred in the Pliocene and was associated with the onset of regional Basin and Range extension, erupting most of the basalt, andesite, and dacite lavas in the field between 3.9–2.6 Ma and ending shortly after initiation of the prominent Sierra Nevada range front fault (Bailey, 2004; Hildreth, 2004; Hildreth et al., 2023). After a ~400 kyr hiatus, the Long Valley volcanic field produced a total of ~850 km³ of rhyolite, ~700 km³ of which were high-silica rhyolite lavas and pyroclastic flows, from ~2.2–0.2 ka during eruption of the Glass Mountain rhyolites (~100 km³, 2.2–0.8 ka; Metz and Mahood 1985, 1991; Metz and Bailey, 1993), Bishop Tuff (~650 km³, 767 ka; Hildreth, 1979), and resurgent intracaldera rhyolites (~100 km³, 750–150 ka; Mankinen et al., 1986; Bailey, 1989; Hildreth et al., 2017). For a more detailed summary of the postcaldera rhyolites in the Long Valley Caldera, we refer the reader to Hildreth (2004), Hildreth et al., (2017), and Hildreth (2021).



Figure S.14. View of the Mono Basin's Mount Warren showing the "dirty" remnants of the once-persistent snow patch that formerly filled the nivation hollow.

Before beginning our examination of Panum Crater, look to the skyline at roughly N60°W from our parking area to see prominent Mt. Warren (ele. 12,373 ft = 3771 m). Barring snowfall between the time of this writing and our visit on Oct. 2, you will notice a conspicuous NE-facing snow patch a short distance below the summit. Having kept a close eye on Sierran snow patches for decades, I can say that the Warren patch is one of the most persistent patches in the range.

Between 2012 and 2015, much of California experienced 4 years of record-breaking drought, with annual runoff from the Mono Sierra dropping as low as 40% of normal. In the autumn of 2013, with the patch having shrunk dramatically, I hiked to it in order to examine the steep, newly exposed ground surface. Reduction in size, and a blackening of the ice surface, continued through 2015 (Fig. S.14), and so I made additional visits. Emerging from the receding edge of the shrinking mass were large pieces of fresh-looking wood which, upon ignition in the field, smelled distinctly like pine. Close by were two newly exposed bighorn sheep horn cores with small amounts of flesh still attached. Radiocarbon analyses by Tom Guilderson of the Center for Mass Spectrometry at Lawrence-Livermore National Laboratory yielded dates of 5065 ±30 BP on the wood (CAMS 171880) and 4540 ±30 BP on the sheep (CAMS 176809). These results, especially those on the meaty horns, suggest that the specimens must have been "under refrigeration" for the past thousands of years. Based on the snow patch, it would seem that the drought of the 20-teens may have been the most severe of the latter half of the Holocene.

References

- Bailey, R.A., 1989, Geologic map of Long Valley caldera, Mono-Inyo Craters volcanic chain, and vicinity, eastern California: U.S. Geological Survey Miscellaneous Investigations Series Map I-1933, scale 1:62,500.
- Bailey, R.A., 2004, Eruptive history and chemical evolution of the precaldern and postcaldern basalt-dacite sequences, Long Valley, California—Implications for magma sources, current seismic unrest, and future volcanism: U.S. Geological Survey Professional Paper 1692, 75 p.
- Bursik, M., and Sieh, K., 1989, Range front faulting and volcanism in the Mono Basin, eastern California: *Journal of Geophysical Research*, v. 94, no. B11, p. 15,587–15,609.
- Hildreth, W., 1979, The Bishop Tuff: Evidence for the origin of compositional zonation in silicic magma chambers, in Chapin, C.E., and Elston, W.E., eds., *Ash-Flow Tuffs: Geological Society of America Special Paper 180*, p. 43–75.
- Hildreth, W., 2004, Volcanological perspectives on Long Valley, Mammoth Mountain, and Mono Craters: Several contiguous but discrete systems: *Journal of Volcanology and Geothermal Research*, v. 136, p. 169–198.

- Hildreth, W., 2021, Comparative rhyolite systems: Inferences from vent patterns and eruptive episodicities: Eastern California and Laguna del Maule: *Journal of Geophysical Research: Solid Earth*, v. 126, e2020JB020879, <https://doi.org/10.1029/2020JB020879>.
- Hildreth, W., Fierstein, J., and Calvert, A.T., 2017, Early postcaldera rhyolite and structural resurgence at Long Valley Caldera, California: *Journal of Volcanology and Geothermal Research*, v. 335, p. 1–34, <https://doi.org/10.1016/j.jvolgeores.2017.01.005>.
- Hildreth, W., Fierstein, J., and Calvert, A.T., 2023, Precaldera mafic magmatism at Long Valley, California: Magma-tectonic siting and incubation of the Great Rhyolite System: *Journal of Volcanology and Geothermal Research*, v. 433, 107726, <https://doi.org/10.1016/j.jvolgeores.2022.107726>.
- Mankinen, E.A., Grommé, C.S., Dalrymple, G.B., Lanphere, M.A., and Bailey, R.A., 1986, Paleomagnetism and K-Ar ages of volcanic rocks from Long Valley caldera, California: *Journal of Geophysical Research*, v. 91, no. B1, p. 633–652.
- Metz, J.M., and Bailey, R.A., 1993, Geologic map of Glass Mountain, Mono County, California: U.S. Geological Survey Miscellaneous Investigations Series Map I-1995, scale 1:24,000.
- Metz, J.M., and Mahood, G.A., 1985, Precursors to the Bishop Tuff eruption: Glass Mountain, Long Valley, California: *Journal of Geophysical Research*, v. 90, no. B13, p. 11,121–11,126.
- Metz, J.M., and Mahood, G.A., 1991, Development of the Long Valley, California, magma chamber recorded in precaldern rhyolite lavas of Glass Mountain: *Contributions to Mineralogy and Petrology*, v. 106, p. 379–397.

Day 1, S3: South Tufa

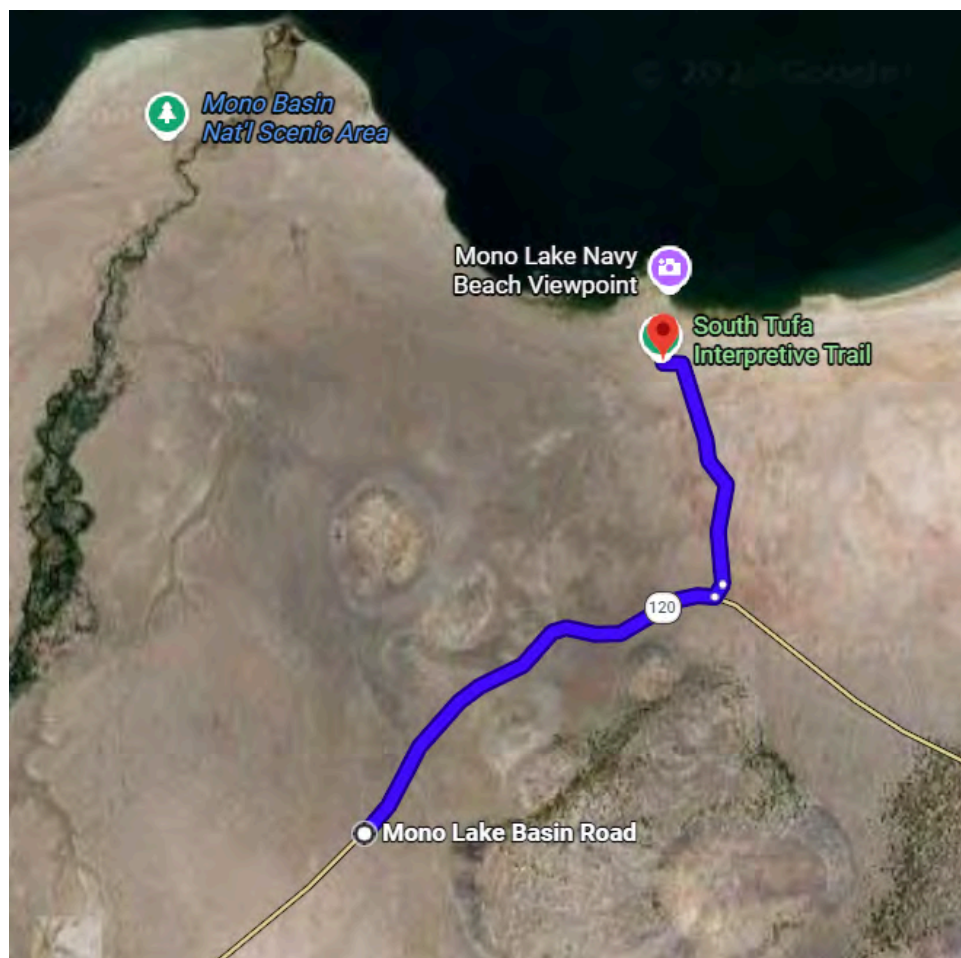
FIELD TRIP STOP CONTRIBUTORS: Bartshe Miller, Mono Lake Committee; Mark Woysner, Balance Hydrologics; Zachary Smith, University of California Berkeley; Scott Stine, Emeritus California State University, East Bay

LOGISTICS: Park at South Tufa parking lot and take the scenic hike to the lake shore.

WALKING DISTANCE: 0.7 miles across relatively level ground. Optional hike to soda springs between South Tufa and Navy Beach.

LOCATION: South Tufa (37.93867, -119.02710)

DIRECTIONS: [Panum Crater to South Tufa](#)



Conservation in the Mono Basin

Contributor: Bartshe Miller (Mono Lake Committee)

Nine decades of water diversions at Mono Lake—chronic problems

In 1941, the Los Angeles Department of Water & Power (DWP) began diverting water from Mono Lake's tributary streams, sending it 350 miles south to meet the growing water demands of Los Angeles.

As a result, over the next 40 years Mono Lake dropped by 45 vertical feet, lost half its volume, and doubled in salinity—threatening the survival of the nesting California Gull population, air quality with toxic dust storms, and this unique and critical ecosystem. Thanks to successful citizen advocacy, and a long, landmark legal effort in California, a management lake level mandate and a set of rules governing DWP's water exports were finally achieved in 1994, and Mono Lake was saved from continued decline and destruction.

Raising the lake to the mandated management level of 6,392 feet above sea level is critical to the lake's ecological health, and the single most important metric for restoring Mono Lake. However, after 32 years the lake is only halfway to the mandated level.

Current export Rx not working to raise Mono Lake

When Mono Lake is between 6,380 and 6,391 feet above sea level, DWP can export up to 16,000 acre-feet of water per year. If the lake drops to between 6,377 and 6,380 feet, DWP's exports are reduced to a maximum of 4,500 acre-feet of water a year. And if Mono Lake is below or forecasted to be below 6,377 feet, DWP cannot export any water.

According to the current export rules, once Mono Lake rises to the target level of 6,391 feet above sea level, a new set of rules will apply to DWP's water exports. However the current rules are not getting the job done. One reason is that after the lake rises in wet years DWP continues to export water, chipping away at the progress made.

DWP exported 16,000 acre-feet—the maximum allowed—last runoff year ending March 31, 2026. Immediately on April 1, 2026, DWP began exporting water from the Mono Basin for the current runoff year, maximizing its diversions to the detriment of Mono Lake.

DWP strives to maximize diversions whenever conditions in the Los Angeles Aqueduct allow, and Mono Lake is not progressing toward its mandated lake level. This, despite Los Angeles' commitment, made in 1994, to raise Mono Lake to its 6,392' foot level. The current diversion prescription is not working for Mono Lake, and it's up to the State Water Board to change the export rules.

Public Trust values under threat, Mono Lake has yet to fully recover

Mono Lake must rise to 6,392 feet above sea level, which is the management level set by the California State Water Resources Control Board's [Decision 1631](#) in 1994.

The management level was chosen for several reasons:

- reduce the lake's salinity so that alkalis flies and endemic brine shrimp can thrive, in turn providing ample food for millions of migratory and nesting birds including phalaropes and eared grebes.
- improve air quality by covering much of the emissive, exposed dry lakebed, thereby reducing toxic dust storms
- protect [California Gull nesting islets](#) from on-shore predators with a substantial moat of water
- leave scenic tufa towers exposed around the shoreline
- to provide a buffer in order to protect these resources during lake level declines caused by dry years and droughts

As of September 1, 2026 Mono Lake is 9.6 feet below the management level.

A Decision with a health plan and check-up

In 1994 the State Water Board acknowledged that its transition plan to raise the lake in about 20 years could be thrown off schedule by actual climate and hydrology in the years following the decision. They wisely put a date two decades later, in 2014, onto the calendar and said that if the lake had not risen as expected, they would hold a hearing to see if the water exports allocated to DWP might need adjusting to solve the problem.

In other words, the State Water Board's lake protection requirement would not change over the years, but the rules for achieving it might.

The slower-than-expected lake rise scenario, unfortunately, has materialized. And the Board's once-distant calendar date passed a decade ago. DWP has more than enough available water to replace Mono Basin diversions. Los Angeles uses less water now than it has at any other time in its last 50 years despite adding a million more people to its population. Water demand is declining and new groundwater recharge, groundwater treatment, and water recycling facilities are coming online and adding significantly to DWP's water portfolio, meeting the city's commitment to produce 70% of its water locally.

A State Water Board hearing, next steps

On March 17, 2026, the California State Water Resources Control Board focused on Mono Lake's low level in a five-hour discussion at their regular meeting in Sacramento. A new [Mono Lake report and hydrology model from the UCLA Center for Climate Science](#) was the impetus for the agenda item, and there was robust discussion, followed by extensive public comments. At the conclusion, Board members agreed: a formal, decision-making hearing about the volume of Los Angeles Department of Water & Power (DWP) water diversions is necessary and should initiate this year.

As of this October, the date of that hearing has not yet been set, but Committee preparation is well underway, and an announcement for a hearing process and timeline is expected soon.

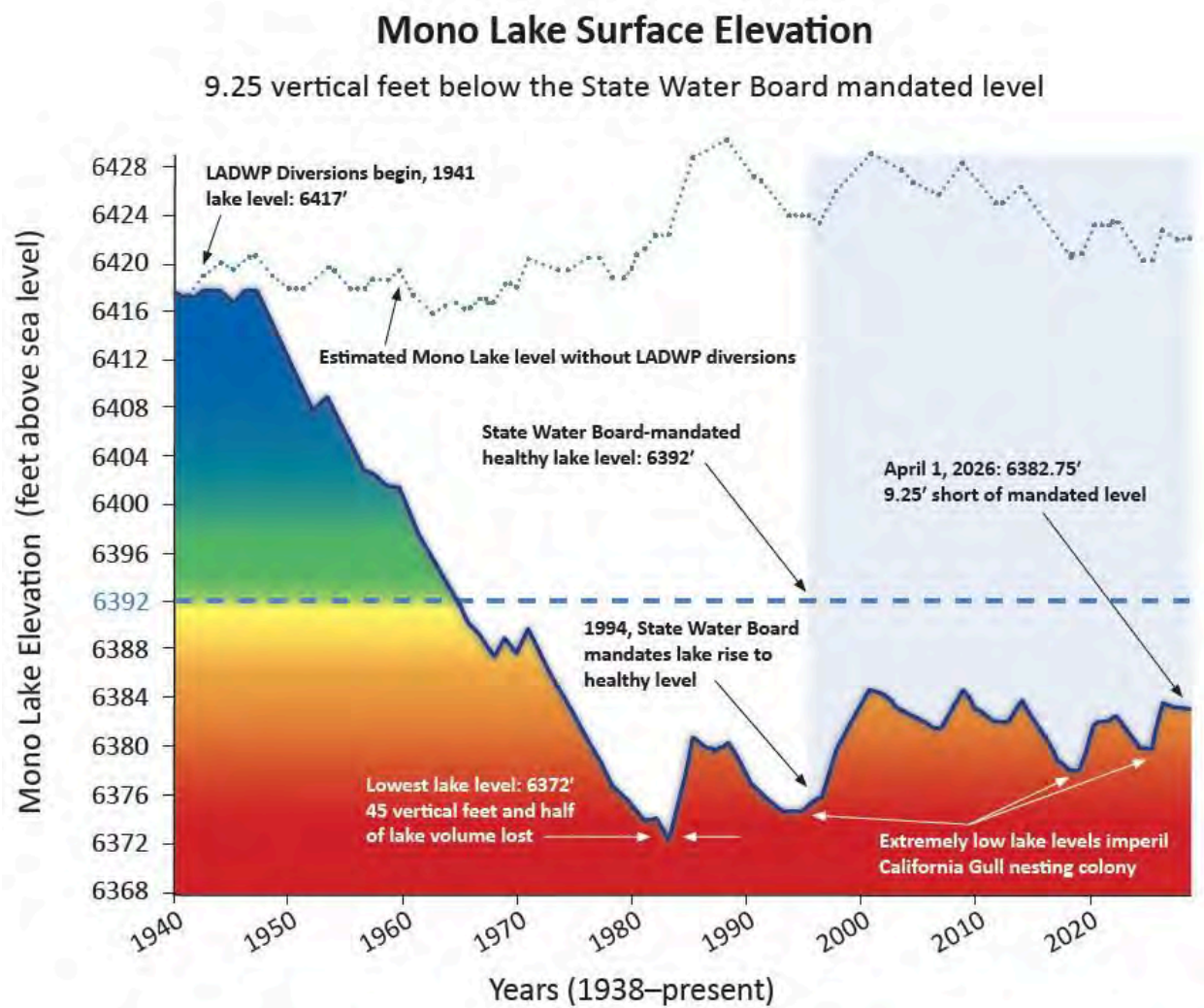


Figure S3.1. Mono Lake level changes after the diversion of water from tributaries.

Groundwater source characterization for Glass Mountain meadows restoration, Inyo National Forest, Mono County, California

Contributors: Erika Groh, Mark Woyshner, and Steve Hnat, PG (Balanced Hydrologics)

Situated at the north end of the Owens Valley Rift and at the base of the Sierra Nevada frontal fault escarpment, the Long Valley Caldera is one of major volcanic regions of California, has a complex tectonic and volcanic history, and remains tectonically active with a swarm of earthquakes occurring in the 1980s and with ongoing gas emissions, steam vents, and hot springs. California Trout (CalTrout) is working with the Inyo National Forest to identify meadow restoration areas and inform future restoration actions in the Glass Mountains as part of the Eastern Sierra Climate and Communities Resilience Project. Given its morphology and spring sources, Crooked Meadow is potentially more persistently wet than other meadows within the region but was classified in very poor condition based on a reconnaissance level rapid assessment (VNLC, 2024).

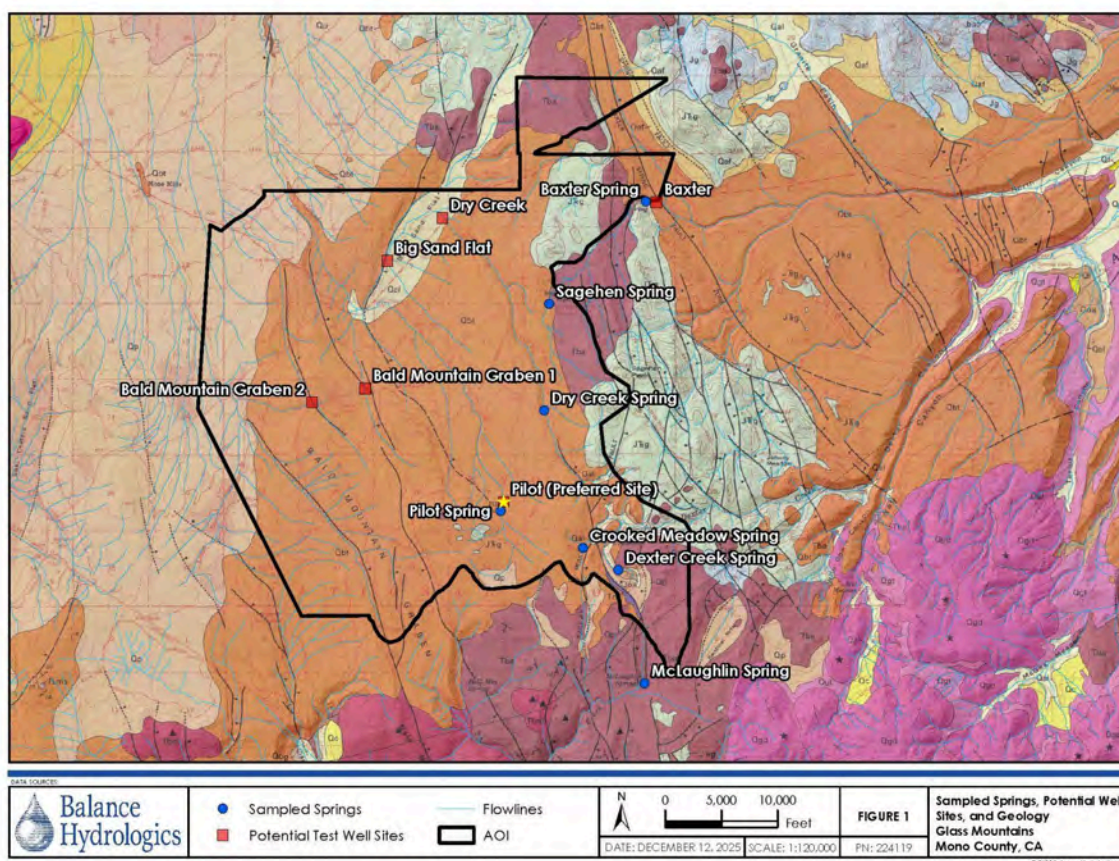
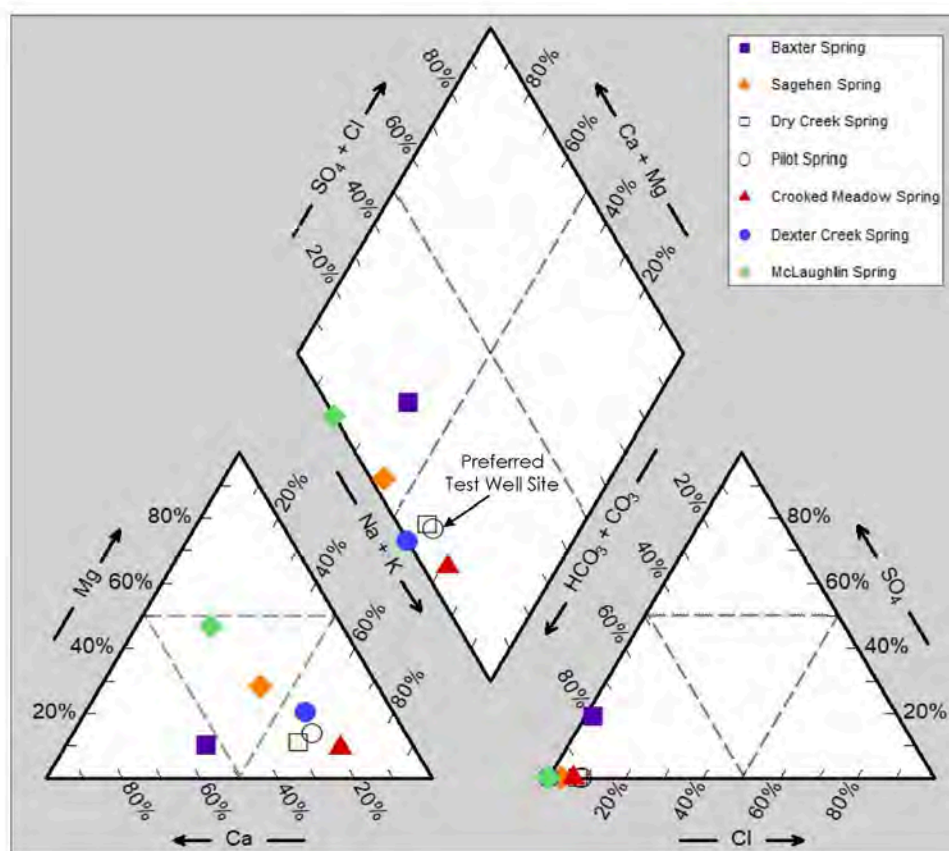


Figure S3.2. Sampled Springs, Potential Well Sites, and Geology Glass Mountains, Mono County, CA.

Crooked Meadow has a high potential for restoration. The 115-acre meadow is situated near the watershed divide of Dexter Creek which drains northeast from volcanic domes and flows on the north rim of the Long Valley Caldera. West of the divide, Dry Creek and Pilot Creek flow north down to Mono Lake, and south of the divide, McLaughlin Creek flows to Long Valley. Restoration of the degraded portions of the meadow has many benefits to meadow and downstream habitats, and as such, a proposed water well was sited to supply water for livestock to lessen diversions impacts on the meadow.



This diagram shows cations in the ternary graph on the left and anions on the right graph. The diamond graph in the center illustrates both cations and anions. Hardness dominated water plots to the left and top of the diamond graph, soft monovalent-salt dominated water to

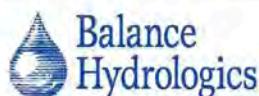


Figure 2. Piper diagram illustrating ionic signatures of spring samples collected in the Glass Mountains, Mono County, California. Crooked Meadow, Dexter Cr, Dry Cr, and Pilot spring samples are classified as sodium bicarbonate groundwater, while Baxter, Sagehen, and McLaughlin spring samples are a mixed cation bicarbonate groundwater with calcium primary at Baxter Spring and magnesium at McLaughlin Spring.

Figure S3.3. Piper diagram of springs sampled.

We sampled springs for general mineral, oxygen-18 and deuterium analyses within a reconnaissance area covered on the USGS 15-minute Cowtrack Mountain quadrangle and geologic map by Bailey (1989). We used the isotope and water-quality results as a line of reasoning independent of the interpreted geologic framework to assist with understanding the sources of groundwater supply for the water-well selected for drilling. Groundwater samples classified as sodium bicarbonate groundwater were interpreted as a Bishop Tuff bedrock source, while other samples of mixed cation bicarbonate groundwater indicated sources in other bedrock types – in basalt, in basalt and Bishop Tuff, and in granodiorite and Bishop Tuff. Interpretation of the stable isotope results from the spring samples along with prior USGS sample results of nearby wells suggested springs with higher elevation recharge and larger groundwater sheds, preferred for well siting. We identified six potential well sites, ranked three of them, and proposed a preferred site for review by the project team.

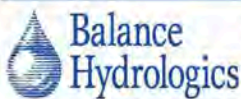
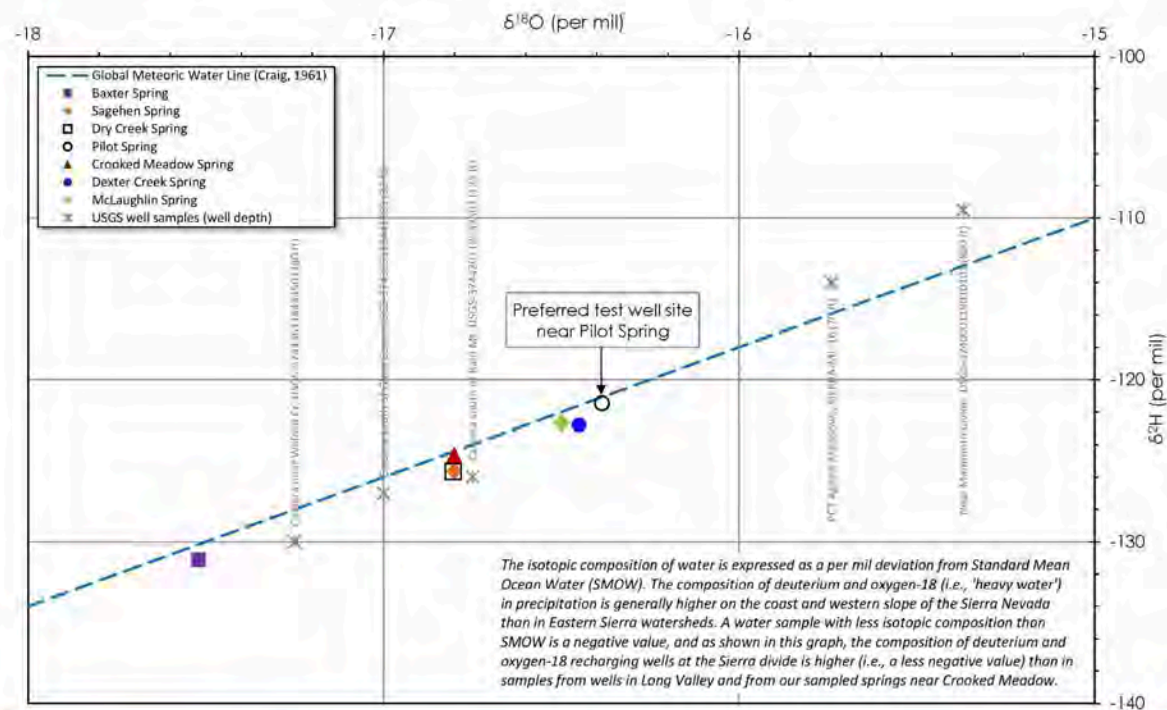


Figure 3. Stable isotopes oxygen-18 ($\delta^{18}\text{O}$) and deuterium ($\delta^2\text{H}$) sampled in springs in the Glass Mountains, Mono County, California. Results for Pilot Spring, McLaughlin Spring, and Dexter Creek Spring samples suggest higher elevation recharge than the other springs we sampled. Our samples were collected on July 8 and 9, 2025. USGS samples of nearby wells were added for reference.

224119_11_O_isotope_Data_2025-12-18.xlsx; Glass Mountain Springs (2)

©2025 Balance Hydrologics, Inc.

Figure S3.4. Stable isotope results.

Heat Flow as a Tracer of Tufa Formation in Mono Lake

Contributor: Zachary Smith (University of California Berkeley), Matthew Hornbach (Southern Methodist University), Michael Manga (University of California Berkeley)

Along the southern shore of Mono Lake, groundwater emerges from lake-floor sediments and along faults. Tufa grows where calcium-bearing spring water seeping up through the lake bed mixes with the carbonate-rich, alkaline lake water. The iconic towers at South Tufa formed beneath the lake and are now stranded on shore, but the same process continues today in the sediments offshore. That ongoing precipitation, and in places dissolution, of carbonate changes the properties of the shallow aquifer and of the faults that feed it.

At this stop we show how temperature can be used to see these processes below the lake floor. We paired vertical temperature profiles measured in lake-floor sediments with more than 30 years of monthly water-column temperature monitoring. From those data we estimated three quantities by solving the advection–diffusion equation for heat, using the long-term bottom-water record as the upper boundary condition: heat flow, vertical groundwater velocity, and the bulk thermal diffusivity of the sediment. Three results are worth discussing at the stop:

- Deep-sourced groundwater rising along faults carries heat into the lake (positive heat flow), whereas emerging shallow groundwater produces negative heat flow.
- Warm, CO₂-charged springs and cool, CO₂-free springs emerge within tens of meters of each other on the same fault system. The warm springs dissolve carbonate and the cool springs build it.
- Sediment thermal diffusivity is up to three times higher where carbonate cement is forming above the faults. Temperature profiles can therefore map where tufa is growing in the subsurface.

Mono Lake and its tufa

Mono Lake is a closed-basin, hypersaline and highly alkaline lake. Its water is rich in dissolved carbonate but poor in calcium. Wherever calcium-bearing groundwater discharges into the lake, calcium carbonate precipitates rapidly around the spring orifice, building the towers and mounds for which the lake is known (Russell, 1889; Dunn, 1953). Much of this carbonate precipitates first as ikaite, a hydrated calcium carbonate that is stable in cold water. Ikaite then converts to calcite, which gives many tufa deposits their porous, recrystallized texture (Bischoff et al., 1993; Council & Bennett, 1993).

Tufa towers grow beneath the water and are exposed only when the lake falls. Mono Lake rose and fell by tens of meters during the late Holocene, and tufa at different elevations around the basin records those stands (Stine, 1990). The towers along the South Tufa shoreline were largely exposed after stream diversions beginning in 1941 lowered the lake by more than 13 m by the early 1980s. Sand tufa (Figure S3.5D–E) forms where groundwater rises through sandy lake-floor sediment and cements the pipes and channels that carry it. The resulting upward-branching columns and sheets are later exhumed by wind and waves. Microbial communities are closely

involved in sand-tufa cementation (Souza-Egipsy et al., 2005). Stacked sand tufa, like the layered exposure in Figure S3.5E, records precipitation at successive lake levels.

Walking to the shore, look for all stages of the process (Figure S3.5). Small gas seeps pock the shallow sandy lake floor (Figure S3.5A–B), and a cold-water spring discharges at lake level (Figure S3.5C). Landward are the sand tufa (Figure S3.5D–E) and tufa towers (Figure S3.5F), which formed when the lake was higher and are no longer active. They are the exhumed equivalents of the cementation that we infer is happening offshore today.

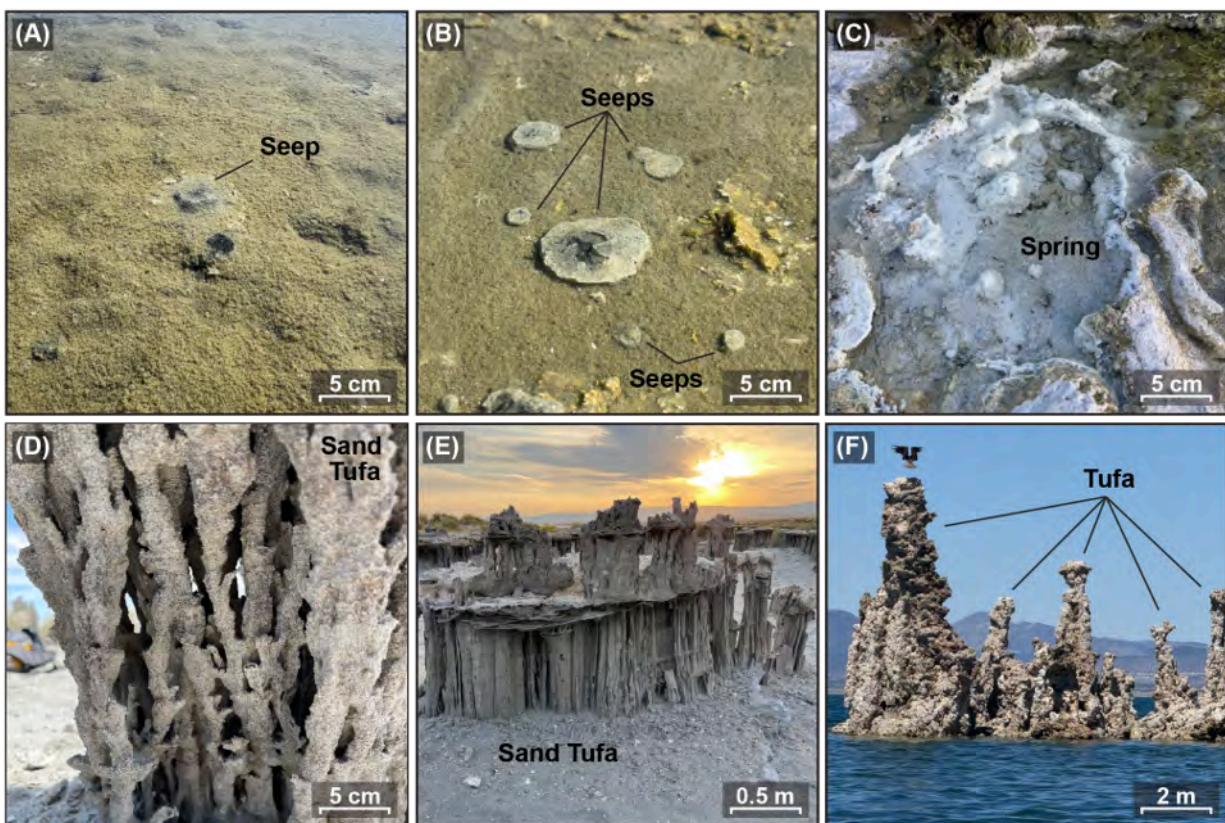


Figure S3.5. Images showing active springs and tufa and sand tufa formations at South Tufa. (A–B) CO₂-bearing seeps in shallow sandy sediment on the lake floor. (C) Cold-water spring at lake level. (D) Upward-branching sand tufa. (E) Sand tufa showing successive layers associated with different lake levels. (F) Tufa towers. Structures shown in D–F formed when the lake was higher and are no longer active, but they provide insight into active carbonate precipitation offshore.

Volcanic and structural setting

South Tufa sits at the north end of the Mono–Inyo volcanic chain, just north of Panum Crater, the northernmost of the Mono Craters (Bursik & Sieh, 1989; Kelleher & Cameron, 1990). Faults that cross the southern shoreline at South Tufa and Navy Beach connect the lake to a deeper magmatic system. They act as conduits for groundwater and CO₂, although the exact location of the magmatic source is uncertain. Between South Tufa and Navy Beach, CO₂ emerges along the shoreline, and gas from the Navy Beach soda spring carries a magmatic signature (Bergfeld et al., 2015).

Offshore, high-resolution seismic reflection data (Jayko et al., 2013) image a fault-bounded, deformed mound that extends north to northwest from South Tufa into the lake (Figure S3.6). Between two faults, a block of Unit 3 is uplifted and internally faulted beneath the lake-floor mound. Tufa mounds rise above the lake floor, and disturbed sediment and a possible gas vent sit above the eastern fault. On shore, springs emerging from the center of the mound are about 30–32 °C and bubble magmatic CO₂. Springs along the eastern and western edges of the mound are cooler (roughly 14–22 °C) and carry no CO₂ bubbles. The CO₂-charged water has a near-neutral pH of about 6–7. Where it emerges, the sediment is soft because carbonate is dissolving. Where Ca- or Na-bearing groundwater without excess CO₂ enters the alkaline lake, it reacts with dissolved carbonate and precipitates tufa.

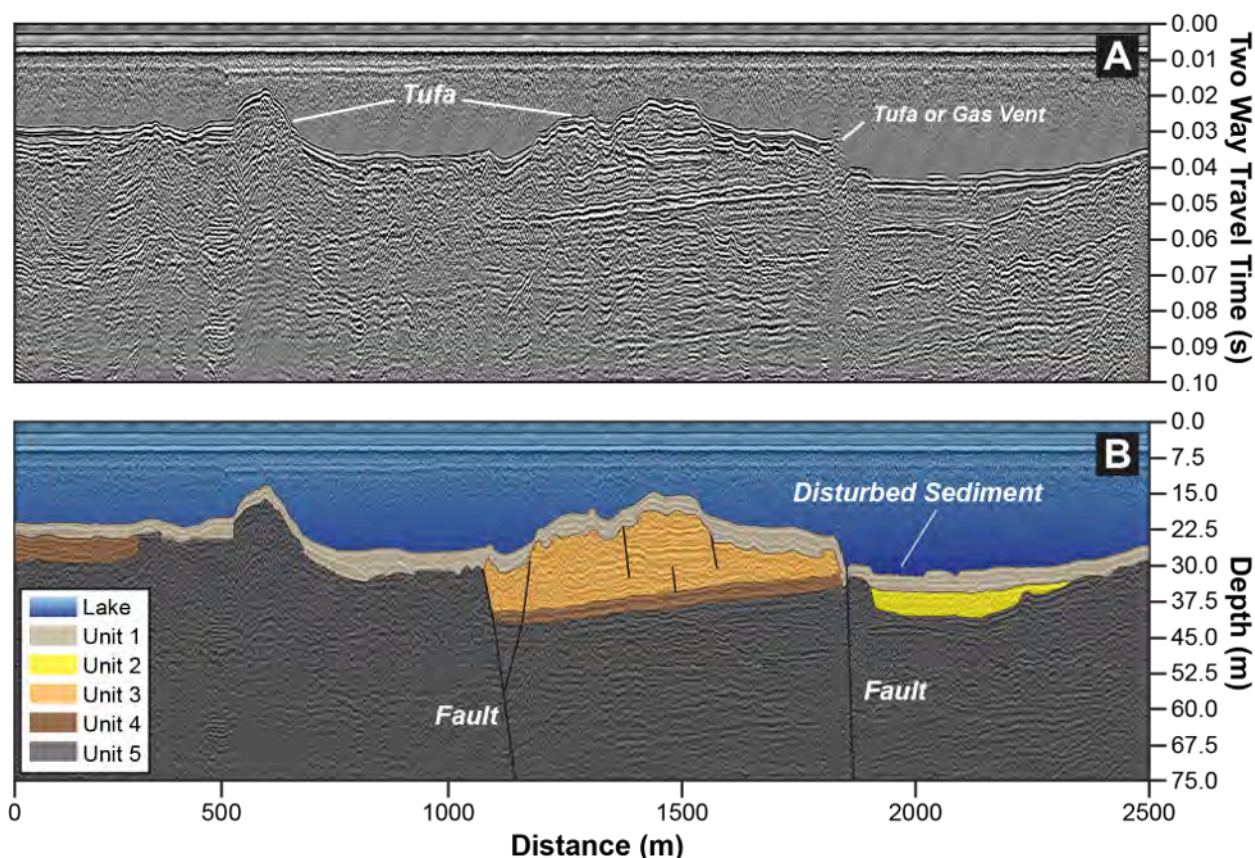


Figure S3.6. Seismic reflection survey across the South Tufa area. (A) Seismic profile and (B) interpretation of the seismic reflection data showing seismic Units 1–5, tufa mounds, two faults bounding a deformed block of Unit 3 beneath the lake-floor mound, and disturbed sediment above the eastern fault. Data collected by Jayko et al. (2013).

Measuring heat flow beneath a shallow lake

Heat flow is usually measured in deep boreholes or in deep ocean basins, where surface temperature changes do not reach the measurement depth. In a shallow lake, bottom-water temperature changes with the seasons and from year to year, and those changes diffuse into the sediment. They can easily swamp the small signal from geothermal heat or groundwater

flow. Satellite thermal images show the problem at the lake surface (Figure S3.7): temperature patterns are dominated by eddies, plumes, and mixing rather than by any geothermal input.

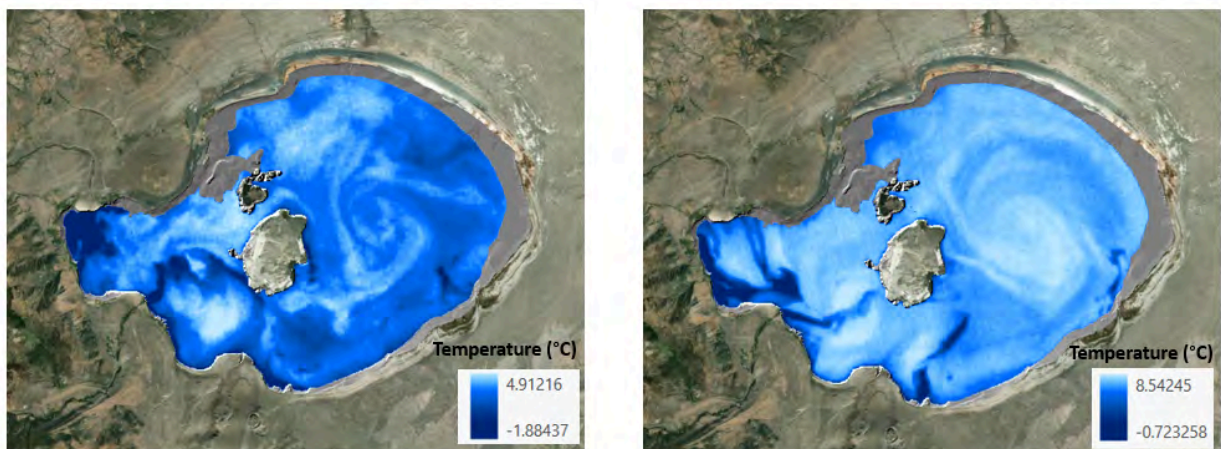


Figure S3.7. Surface temperature of Mono Lake as observed from Landsat 8 thermal infrared imaging. Surface temperature patterns reflect lake circulation and mixing, so geothermal inputs cannot be identified from the lake surface alone.

Mono Lake's stratification history adds another complication. The lake alternates between years of normal seasonal overturn and multi-year periods of meromixis, when a freshened surface layer prevents overturn (Jellison & Melack, 1993). During meromixis, water below the thermocline (15–20 m) stays near a steady 3–4 °C. The exceptionally wet 2017 runoff year produced meromictic conditions that persisted until 2020. Mono Lake has, however, more than 30 years of monthly conductivity–temperature–depth (CTD) profiles. These give a bottom-water temperature history that we can use as the upper boundary condition for a thermal model of the sediments.

Lake-floor temperature profiles were measured in March 2021, March 2022, and August 2022 with a Hybrid Lister-Outrigger (HyLO) probe, originally developed for rapid marine geothermal gradient measurements (Hornbach et al., 2021; Figure S3.8). The probe carries eight thermistors spaced along a 1.5-m stainless steel cylinder. It also carries pressure and humidity sensors and an accelerometer that confirms vertical penetration. Each insertion lasted at least 5 minutes (about 7 minutes on average). Insertion produces a frictional heat pulse that decays toward the equilibrium sediment temperature (Figure S3.8.C), and the records were reduced following Villinger and Davis (1987) and Hornbach et al. (2021). We collected 26 lake-floor profiles. Sixteen are near South Tufa. The other ten are on the southern margin of Paoha Island in Johnson Basin, in water up to 35 m deep, and serve as a reference for undisturbed lake sediment. On shore, 197 shallow profiles (0–50 cm) were measured with soil temperature probes and corrected for surface temperature fluctuations using ERA5 skin temperature.

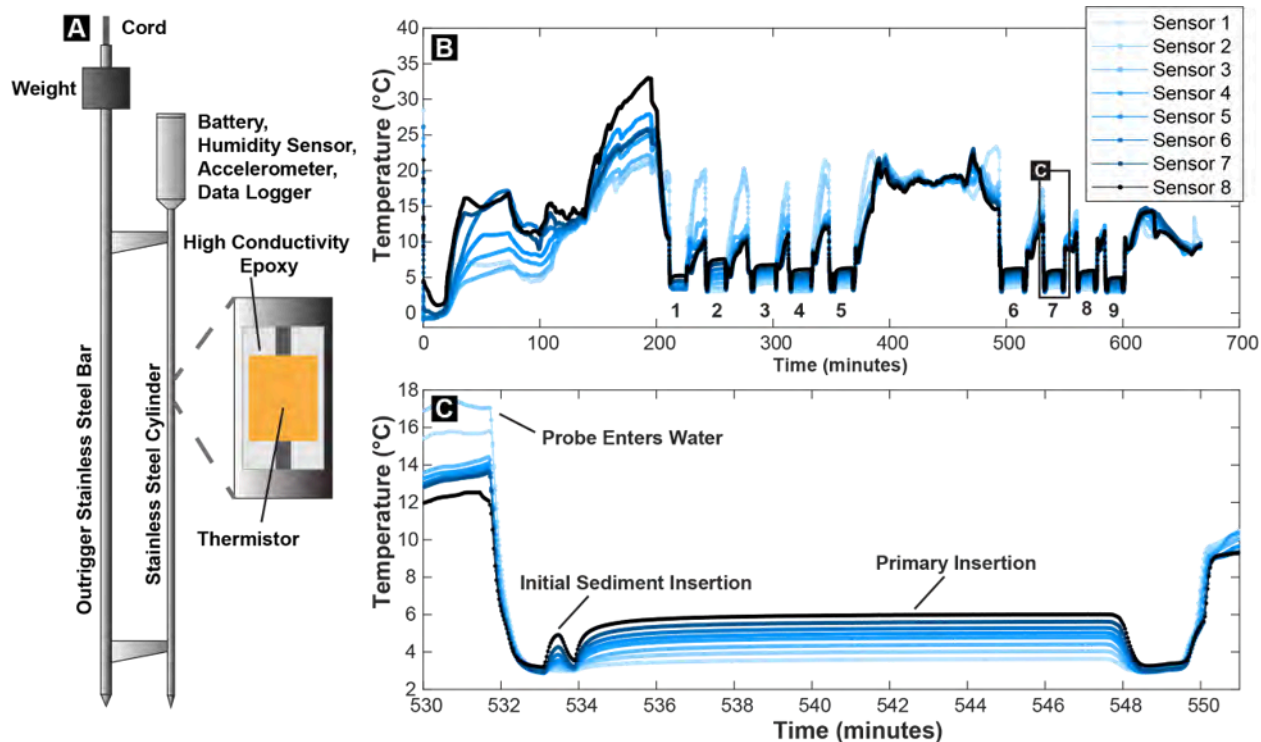


Figure S3.8. Heat flow probe design and example data. (A) Hybrid Lister-Outrigger (HyLO) probe with outrigger support bar and weights; thermistors are set in high-conductivity epoxy within a stainless steel cylinder. (B) Temperature time series for a full day of deployment (March 18, 2022) with nine insertions near South Tufa (numbered). (C) Detail of insertion 7, showing entry into the water, the initial sediment insertion, and the primary insertion used to estimate the equilibrium temperature gradient.

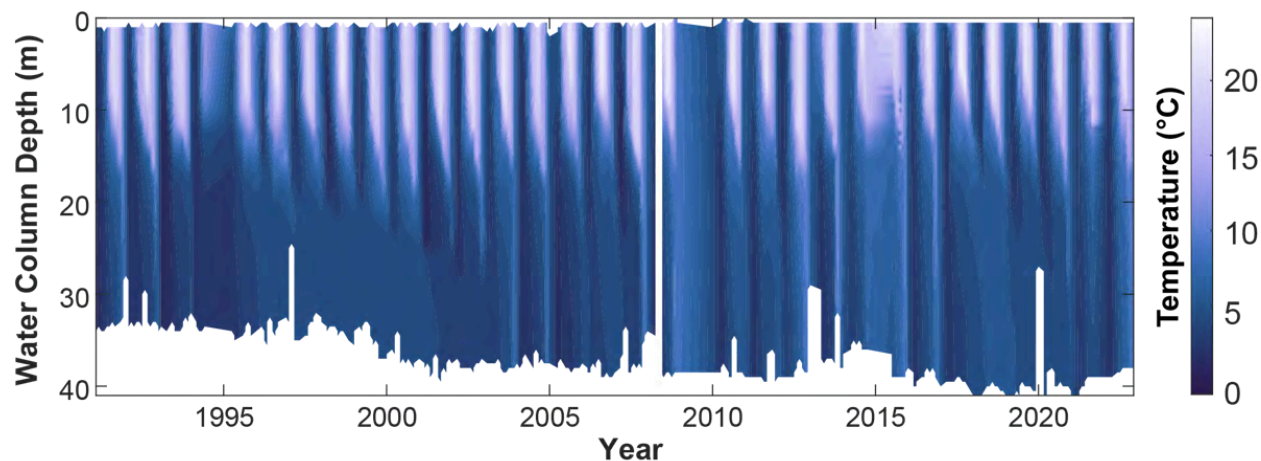


Figure S3.9. Monthly water column temperature measurements from Mono Lake, California as measured by the University of California Santa Barbara and the Los Angeles Department of Water and Power.

Each measured profile is compared with a one-dimensional model of transient heat conduction and advection in saturated sediment (Bredehoeft & Papadopoulos, 1965; Stallman, 1965; Manga

et al., 2012). The model is driven at the top by the bottom-water temperature record as constrained from monthly CTD measurements (Figure S3.9). Heat flow, fluid velocity, and thermal diffusivity are adjusted to minimize misfit with the observed temperatures, and the equation is solved with an explicit fourth-order compact scheme (Nawaz et al., 2021). The approach extends the long tradition of using heat as a groundwater tracer (Suzuki, 1960; Anderson, 2005; Hatch et al., 2006; Constantz, 2008) and recent spectral methods applied to Yellowstone Lake sediments (Sohn & Harris, 2021). The difference here is that an existing limnological record replaces a long-term sensor deployment on the lake floor.

What heat flow reveals

Across the lake, heat flow anomalies cluster around Paoha Island and along the southern shore (Figure S3.10). A strong positive anomaly lies offshore of Hot Springs Cove on the east side of Paoha, and negative anomalies surround the rest of the island. Johnson Basin, by contrast, is thermally quiet. Its heat flow is only $0.01\text{--}0.06\text{ W m}^{-2}$, and its profiles show no measurable vertical fluid flow; the best-fitting velocities sit at the lowest value tested, 10^{-10} m s^{-1} . Paired positive and negative anomalies appear offshore of Navy Beach and South Tufa.

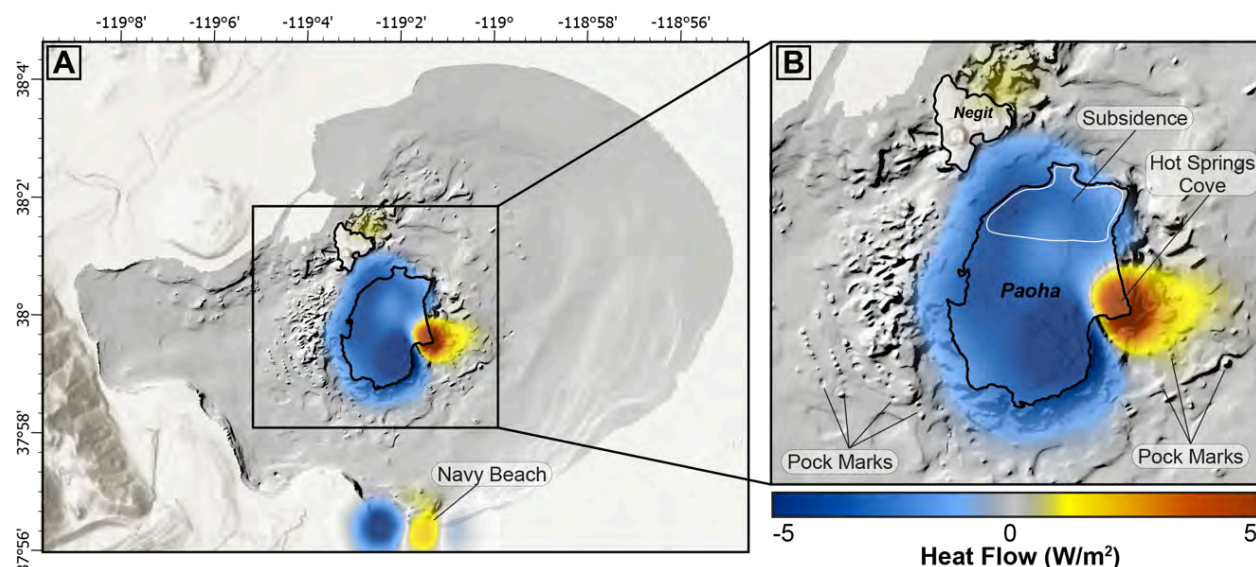


Figure S3.10. Map of heat flow anomalies in Mono Lake. (A) Lake-wide map showing anomalies around Paoha Island and offshore of Navy Beach. (B) Detail of Paoha Island showing a positive anomaly at Hot Springs Cove, negative anomalies around the island, an area of subsidence, and fields of pock marks on the lake floor.

Near South Tufa, heat flow is two to three orders of magnitude more variable than in Johnson Basin, ranging from roughly -10 to $+10\text{ W m}^{-2}$ (Figure S3.11). Vertical fluid velocities range from effectively zero (10^{-10} m s^{-1}) to 10^{-6} m s^{-1} , so groundwater discharge is distributed along the shoreline and nearshore lake floor. The onshore and lake-floor measurements were processed independently but give comparable results. The highest heat flow lies along the shoreline between South Tufa and Navy Beach, where mapped faults reach the shore. Positive heat flow

marks deep-sourced groundwater discharging diffusely along the faults. Negative heat flow marks shallow groundwater, which is cooler than the seasonal surface forcing alone would predict.

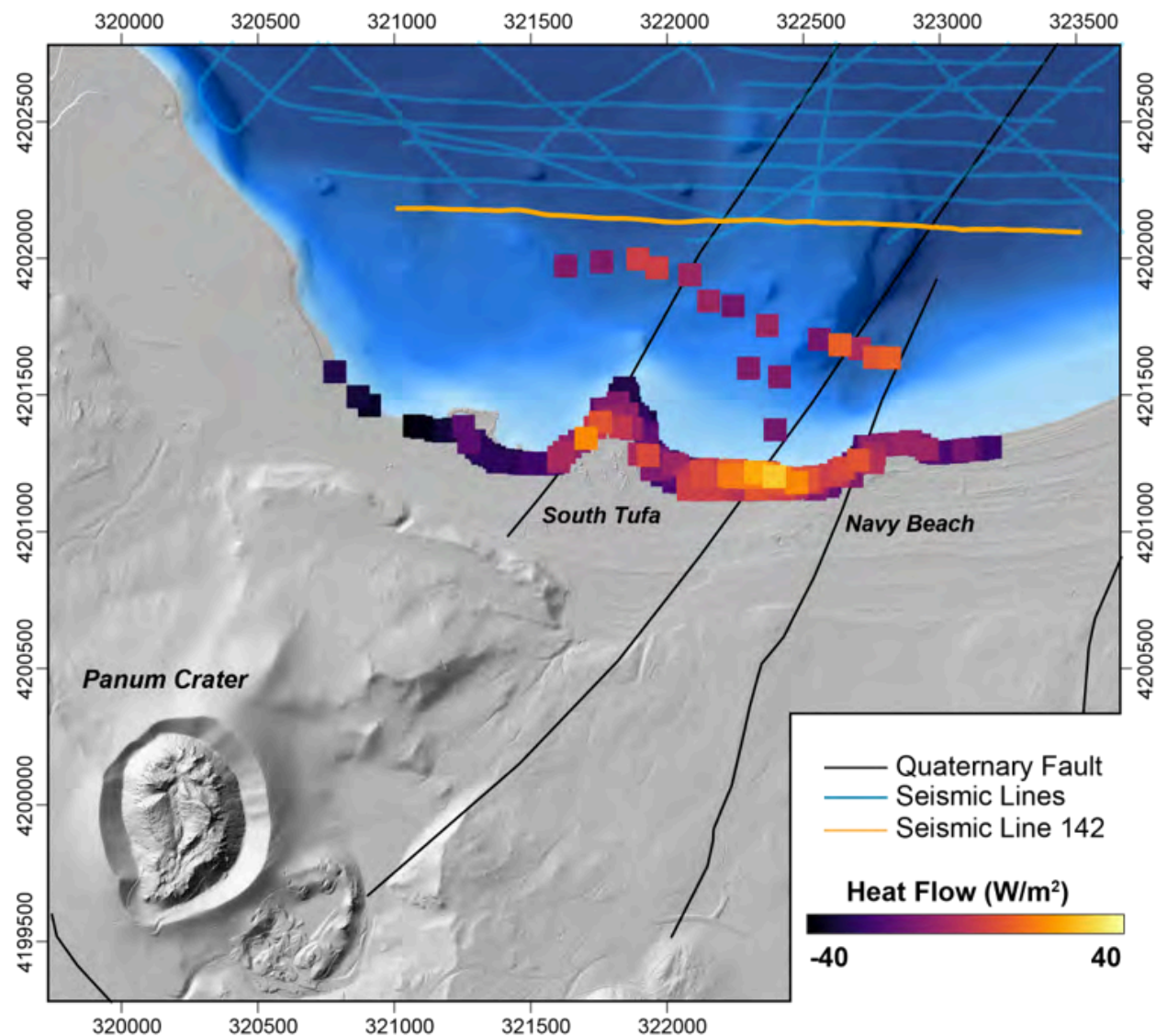


Figure S3.11. Heat flow near South Tufa from lake-floor (offshore squares) and onshore vertical temperature profiles. Black lines are faults. High heat flow areas are associated with faults and deeply sourced groundwater, while low heat flow anomalies are associated with shallow groundwater flow. The orange line marks the seismic profile shown in Figure S3.6.

Thermal diffusivity as a tracer of subsurface tufa formation

The most surprising result is how much thermal diffusivity varies at South Tufa. Every Johnson Basin site is best fit by the same value, $2.5 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$. At South Tufa, best-fitting values span the full range tested, $1.25\text{--}7.5 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$ (Figure S3.12). Typical lake sediments fall between

about 1.5 and $3.5 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$. The highest values, up to about $7.5 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$, lie on the eastern and western flanks of the mound above the faults, where the cooler, CO_2 -free springs emerge and tufa forms.

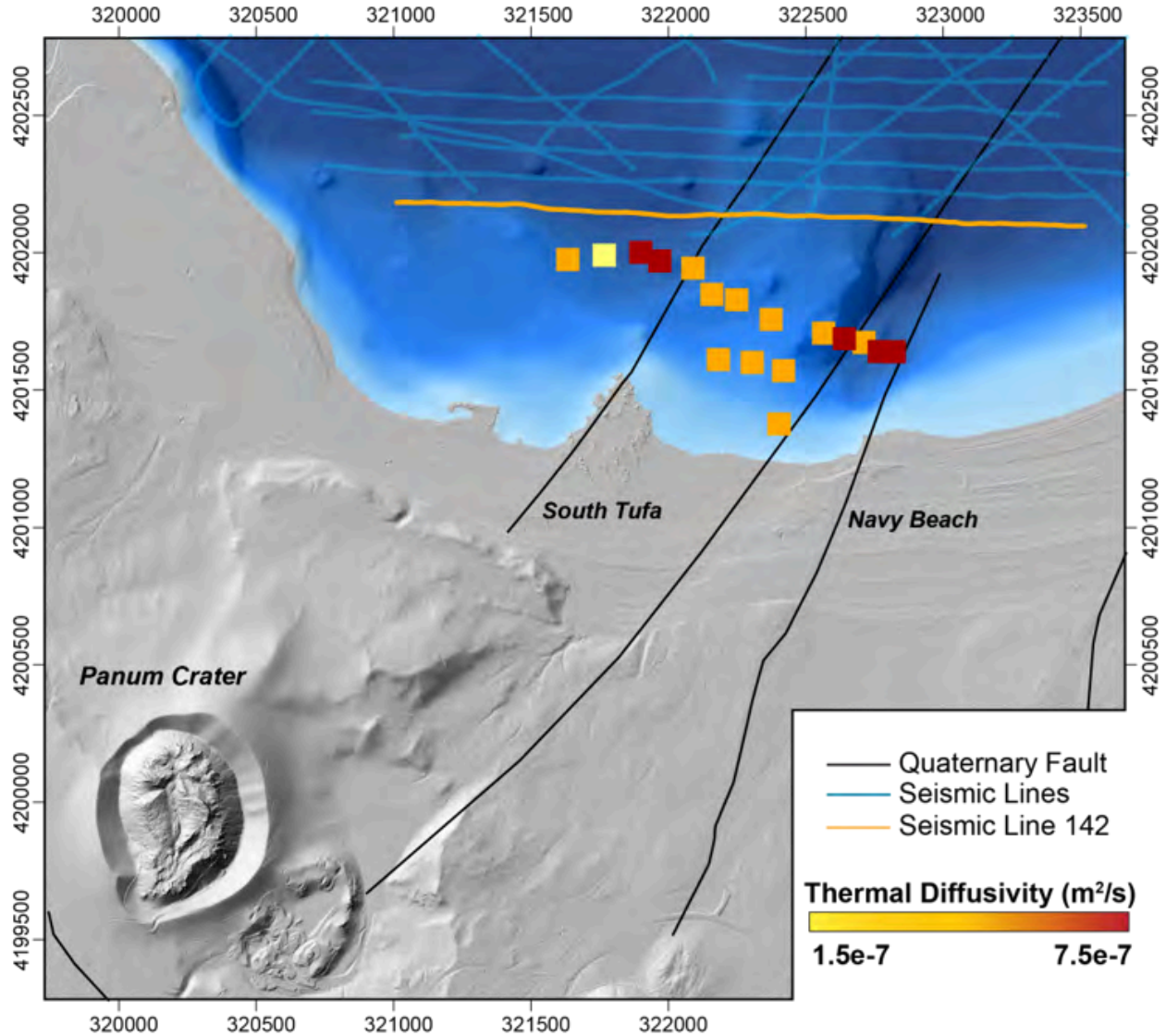


Figure S3.12. Thermal diffusivity near South Tufa estimated from lake-floor temperature profiles. High thermal diffusivity areas are associated with faults where subsurface tufa is actively growing.

Carbonate cement is the simplest explanation. The uncemented sediment at South Tufa ranges from pumice-rich sand near shore to organic-rich mud deeper in the lake. Porous pumice and volcanic glass conduct heat poorly (roughly $0.35\text{--}1.4 \text{ W m}^{-1} \text{ K}^{-1}$), and so does pore water ($0.6 \text{ W m}^{-1} \text{ K}^{-1}$). Calcite conducts heat well, at $3.2\text{--}5.0 \text{ W m}^{-1} \text{ K}^{-1}$ (Clauser & Huenges, 1995; Heap et al., 2023). Replacing pore water with carbonate cement between pumice grains therefore raises both the thermal conductivity and the diffusivity of the sediment. A mixing model for effective conductivity as a function of cementation, porosity, and saturation (Johansen, 1975) reproduces

the observed range. The model also shows that lower-than-normal diffusivity, as low as about $1.5 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$, can arise from gas in the pore space or from carbonate dissolution, both expected where the CO_2 -charged springs emerge.

Thin sections of sand tufa from the Navy Beach area support this interpretation (Figure S3.13). The samples were impregnated with blue epoxy, and pore space, grains, and cement were classified with a support vector machine trained on plane- and cross-polarized light scans (Smith & Griffith, 2022). Cementation is highly heterogeneous. In places it is distributed throughout the sand; in others it is concentrated in thin zones lining the conduits that carried fluid. Similar increases in thermal conductivity accompany microbially induced calcite precipitation in engineered soils (Venuleo et al., 2016) and cementation of planetary regolith (Piqueux & Christensen, 2009).

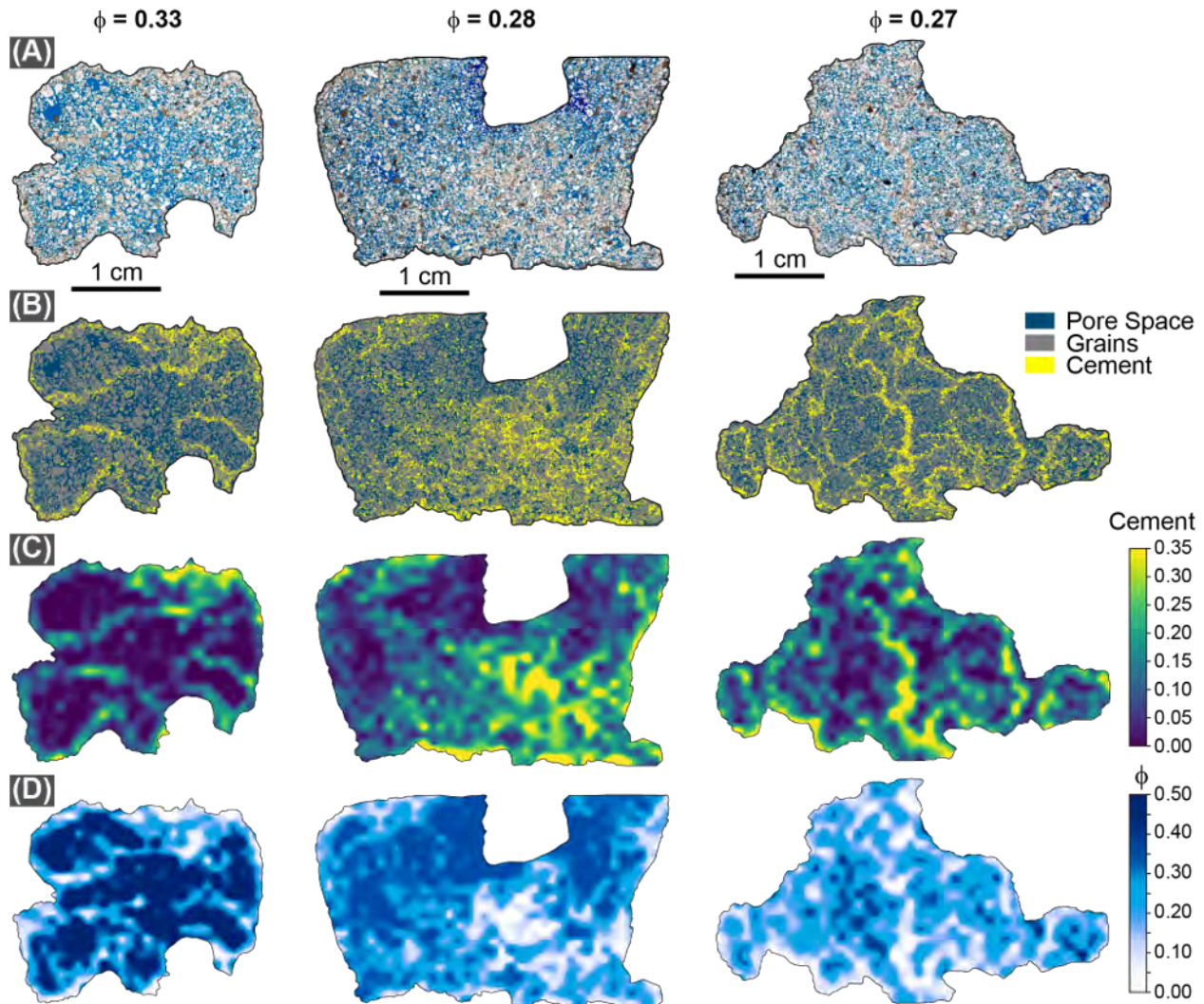


Figure S3.13. Porosity and cementation of sand tufa from the Navy Beach area. (A) Plane-polarized light photomicrographs (blue is epoxy-filled pore space). (B) Machine learning classification of pore space, grains, and cement. Maps of (C) cementation and (D) porosity estimated over 1 mm bins.

Compared with other lake and seafloor settings, South Tufa spans nearly the full range of thermal diffusivity found in natural sediments (Figure 10). Its low end matches non-thermal Yellowstone Lake sediments (Sohn & Harris, 2021). Its high end overlaps sandy tidal flats and low-temperature hydrothermal alteration at the TAG hydrothermal field on the Mid-Atlantic Ridge. All of that variability occurs within a few hundred meters.

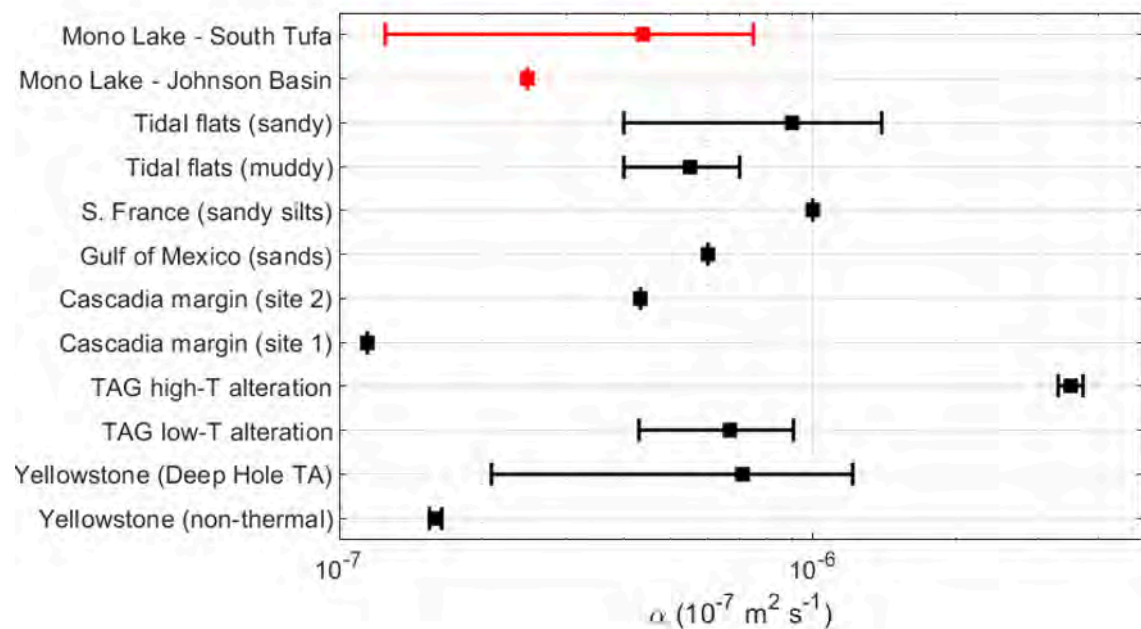


Figure S3.14. Comparison of thermal diffusivity at South Tufa and Johnson Basin, Mono Lake (red), with estimates from other lake and ocean floor environments (black). Squares are best estimates and bars show the reported range.

Heat Flow on Paoha Island

From the shoreline, look north across the lake to Paoha Island. At Hot Springs Cove on its eastern side, the heat flow anomaly is larger and more localized. There, hydrothermal fluids circulate through a fractured lava dome, mix with meteoric water recharged on the island, and discharge through fumaroles and into less than 300 years of lake sediment overlying a landslide deposit. Comparing the two sites separates heat carried by fault-guided groundwater (South Tufa) from heat supplied by a young, cooling intrusion (Paoha).

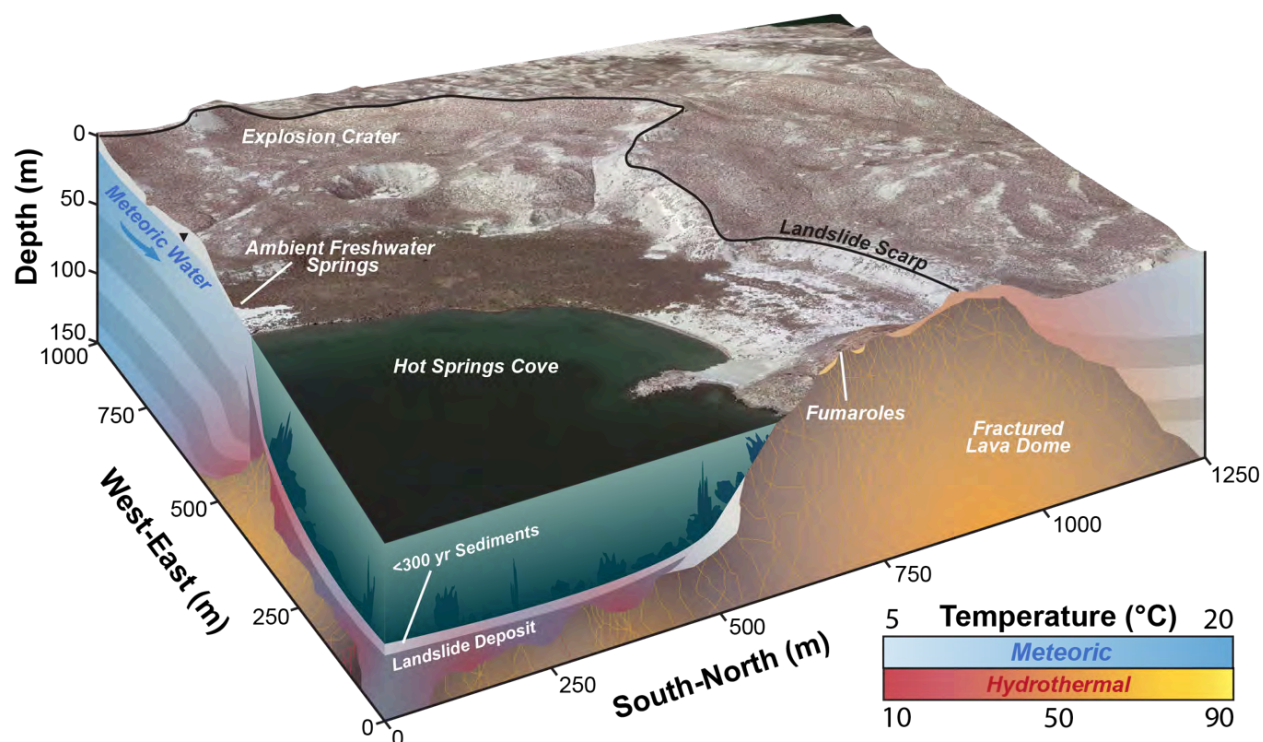


Figure S3.15. Conceptual model showing heat and fluid flow in and around Hot Springs Cove, Paoha Island. Meteoric water and hydrothermal fluids circulating through a fractured lava dome discharge through fumaroles and into young (<300 yr) lake sediments overlying a landslide deposit.

References

- Anderson, M. P. (2005). Heat as a ground water tracer. *Ground Water*, 43(6), 951–968.
<https://doi.org/10.1111/j.1745-6584.2005.00052.x>
- Bergfeld, D., Evans, W. C., Howle, J. F., & Hunt, A. G. (2015). Magmatic gas emissions at Holocene volcanic features near Mono Lake, California, and their relation to regional magmatism. *Journal of Volcanology and Geothermal Research*, 292, 70–83.
<https://doi.org/10.1016/j.jvolgeores.2015.01.008>
- Bischoff, J. L., Stine, S., Rosenbauer, R. J., Fitzpatrick, J. A., & Stafford, T. W., Jr. (1993). Ikaite precipitation by mixing of shoreline springs and lake water, Mono Lake, California, USA. *Geochimica et Cosmochimica Acta*, 57(16), 3855–3865.
[https://doi.org/10.1016/0016-7037\(93\)90339-X](https://doi.org/10.1016/0016-7037(93)90339-X)
- Bredehoeft, J. D., & Papadopoulos, I. S. (1965). Rates of vertical groundwater movement estimated from the Earth's thermal profile. *Water Resources Research*, 1(2), 325–328.
<https://doi.org/10.1029/WR001i002p00325>
- Bursik, M., & Sieh, K. (1989). Range front faulting and volcanism in the Mono Basin, eastern California. *Journal of Geophysical Research*, 94(B11), 15587–15609.
<https://doi.org/10.1029/JB094iB11p15587>

- Clauser, C., & Huenges, E. (1995). Thermal conductivity of rocks and minerals. In T. J. Ahrens (Ed.), *Rock physics and phase relations: A handbook of physical constants* (AGU Reference Shelf, Vol. 3, pp. 105–126). American Geophysical Union. <https://doi.org/10.1029/RF003p0105>
- Constantz, J. (2008). Heat as a tracer to determine streambed water exchanges. *Water Resources Research*, 44, W00D10. <https://doi.org/10.1029/2008WR006996>
- Council, T. C., & Bennett, P. C. (1993). Geochemistry of ikaite formation at Mono Lake, California: Implications for the origin of tufa mounds. *Geology*, 21(11), 971–974.
- Dunn, J. R. (1953). The origin of the deposits of tufa in Mono Lake. *Journal of Sedimentary Petrology*, 23(1), 18–23. <https://doi.org/10.1306/D4269530-2B26-11D7-8648000102C1865D>
- Giambastiani, B. M. S., Colombani, N., & Mastrocicco, M. (2013). Limitation of using heat as a groundwater tracer to define aquifer properties: Experiment in a large tank model. *Environmental Earth Sciences*, 70(2), 719–728. <https://doi.org/10.1007/s12665-012-2157-2>
- Hatch, C. E., Fisher, A. T., Revenaugh, J. S., Constantz, J., & Ruehl, C. (2006). Quantifying surface water–groundwater interactions using time series analysis of streambed thermal records: Method development. *Water Resources Research*, 42, W10410. <https://doi.org/10.1029/2005WR004787>
- Heap, M. J., Wadsworth, F. B., & Jessop, D. E. (2023). The thermal conductivity of unlithified granular volcanic materials: The influence of hydrothermal alteration and degree of water saturation. *Journal of Volcanology and Geothermal Research*, 436, 107775. <https://doi.org/10.1016/j.jvolgeores.2023.107775>
- Hornbach, M. J., Sylvester, J., Hayward, C., Kühn, M., Huhn-Frehers, K., Freudenthal, T., Watt, S. F. L., Berndt, C., Kutterolf, S., Kuhlmann, J., Sievers, C., Rapp, S., Pallapies, K., Gatter, R., & Hoenekopp, L. (2021). A hybrid Lister-outrigger probe for rapid marine geothermal gradient measurement. *Earth and Space Science*, 8(1), e2020EA001327. <https://doi.org/10.1029/2020EA001327>
- Jayko, A. S., Hart, P., Childs, J., Cormier, M., Ponce, D., Athens, N., & McClain, J. S. (2013). Methods and spatial extent of geophysical investigations, Mono Lake, California, 2009 to 2011 (U.S. Geological Survey Open-File Report 2013-1113). <https://doi.org/10.3133/ofr20131113>
- Jellison, R., & Melack, J. M. (1993). Meromixis in hypersaline Mono Lake, California. 1. Stratification and vertical mixing during the onset, persistence, and breakdown of meromixis. *Limnology and Oceanography*, 38(5), 1008–1019. <https://doi.org/10.4319/lo.1993.38.5.1008>
- Johansen, O. (1975). Thermal conductivity of soils (Doctoral dissertation). Norwegian Institute of Technology, Trondheim. (U.S. Army CRREL Draft Translation 637, 1977)

- Kelleher, P. C., & Cameron, K. L. (1990). The geochemistry of the Mono Craters–Mono Lake Islands volcanic complex, eastern California. *Journal of Geophysical Research*, 95(B11), 17643–17659. <https://doi.org/10.1029/JB095iB11p17643>
- Manga, M., Hornbach, M. J., Le Friant, A., Ishizuka, O., Stroncik, N., Adachi, T., et al. (2012). Heat flow in the Lesser Antilles island arc and adjacent back arc Grenada basin. *Geochemistry, Geophysics, Geosystems*, 13, Q08007. <https://doi.org/10.1029/2012GC004260>
- Nawaz, Y., Arif, M. S., Shatanawi, W., & Nazeer, A. (2021). An explicit fourth-order compact numerical scheme for heat transfer of boundary layer flow. *Energies*, 14(12), 3396. <https://doi.org/10.3390/en14123396>
- Piqueux, S., & Christensen, P. R. (2009). A model of thermal conductivity for planetary soils: 2. Theory for cemented soils. *Journal of Geophysical Research*, 114, E09006. <https://doi.org/10.1029/2008JE003309>
- Russell, I. C. (1889). Quaternary history of Mono Valley, California. In *Eighth Annual Report of the U.S. Geological Survey, 1886–1887 (Part 1, pp. 261–394)*. U.S. Geological Survey.
- Smith, Z. D., & Griffith, W. A. (2022). Lithological controls on fault damage zone development by coseismic tensile loading. *Tectonophysics*, 838, 229471. <https://doi.org/10.1016/j.tecto.2022.229471>
- Sohn, R. A., & Harris, R. N. (2021). Spectral analysis of vertical temperature profile time-series data in Yellowstone Lake sediments. *Water Resources Research*, 57, e2020WR028430. <https://doi.org/10.1029/2020WR028430>
- Souza-Egipsy, V., Wierzbos, J., Ascaso, C., & Nealson, K. H. (2005). Mg–silica precipitation in fossilization mechanisms of sand tufa endolithic microbial community, Mono Lake (California). *Chemical Geology*, 217(1–2), 77–87. <https://doi.org/10.1016/j.chemgeo.2004.12.004>
- Stallman, R. W. (1965). Steady one-dimensional fluid flow in a semi-infinite porous medium with sinusoidal surface temperature. *Journal of Geophysical Research*, 70(12), 2821–2827. <https://doi.org/10.1029/JZ070i012p02821>
- Stine, S. (1990). Late Holocene fluctuations of Mono Lake, eastern California. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 78(3–4), 333–381. [https://doi.org/10.1016/0031-0182\(90\)90221-R](https://doi.org/10.1016/0031-0182(90)90221-R)
- Suzuki, S. (1960). Percolation measurements based on heat flow through soil with special reference to paddy fields. *Journal of Geophysical Research*, 65(9), 2883–2885. <https://doi.org/10.1029/JZ065i009p02883>
- Venuleo, S., Laloui, L., Terzis, D., Hueckel, T., & Hassan, M. (2016). Microbially induced calcite precipitation effect on soil thermal conductivity. *Géotechnique Letters*, 6(1), 39–44. <https://doi.org/10.1680/jgele.15.00125>

Villinger, H., & Davis, E. E. (1987). A new reduction algorithm for marine heat flow measurements. *Journal of Geophysical Research*, 92(B12), 12846–12856.
<https://doi.org/10.1029/JB092iB12p12846>

Tufa Formation and Implications for Age Dating

Contributor: Scott Stine, Emeritus California State University, East Bay

Mono Lake is renowned for the abundance and variety of its tufa deposits. Surface accumulations of this calcium carbonate precipitate occur as towers and domes, as beach rock (including “sand tufa”), and as coatings on shoreland cobbles and boulders. Although these features formed in a sublacustrine environment, many of them are exposed today because diversion of the lake’s feeder streams by the City of Los Angeles has lowered the shoreline as much as 45 ft (14 m) below its natural level.

This stop allows us to consider two tufa-related topics: First, it is an opportunity to discuss the problems associated with radiocarbon dating of tufa; and second, it is a chance to compare South Tufa with other Holocene tufa groves in the basin—and in so doing, to establish how very peculiar, and how very young, the South Grove is.

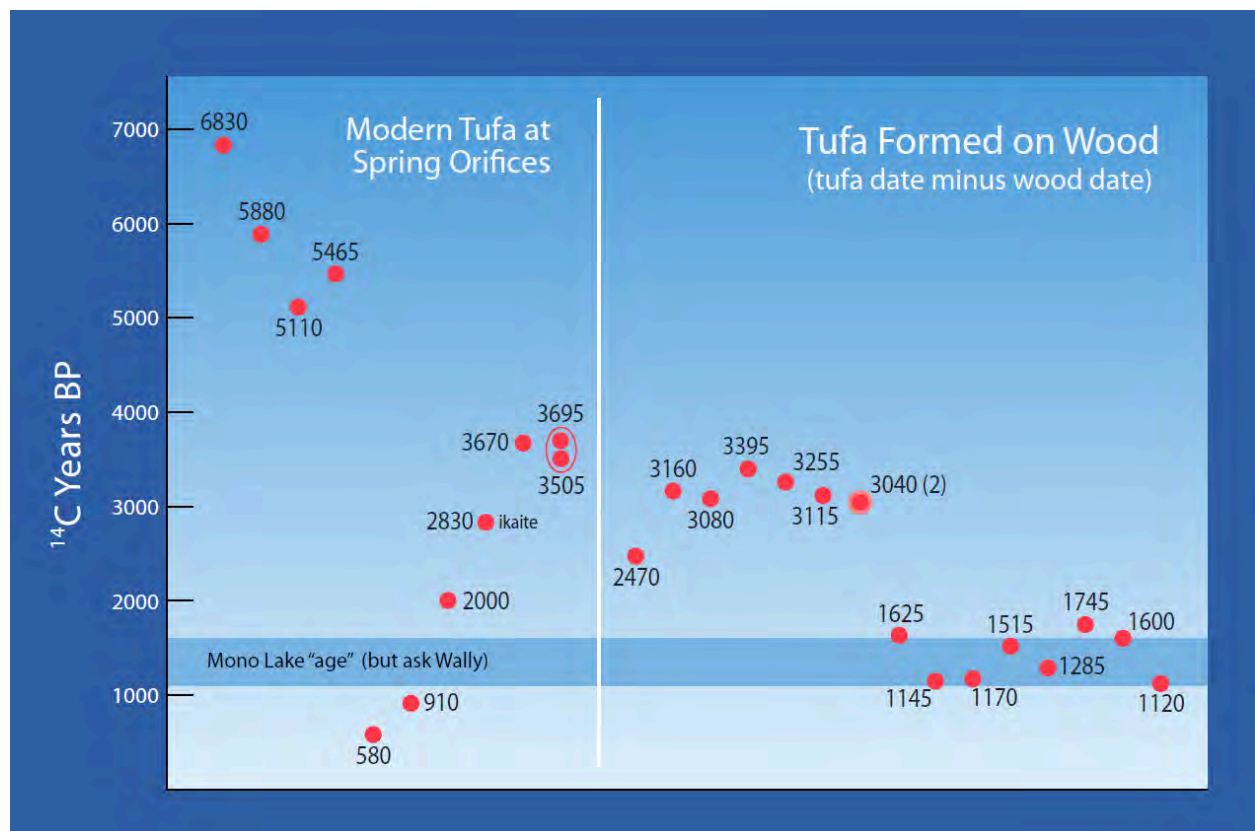


Figure S3.16. Left: Radiocarbon dates on tufa seen to be forming at spring sites on the southern shore of Mono Lake, and at Dechambeau Creek near the lake's northwestern corner. Right: Difference between radiocarbon ages of wood and those of tufa that accumulated on the wood.

Regarding the first topic: While certain characteristics of Mono tufa—including its morphology, its modes of formation (physico- vs. bio-chemical), and the hydrologic settings in which it precipitates—have been studied and reported, the age and history of the deposits were until recently only broadly known, often being categorized as simply “late Pleistocene” or “Holocene.” Development of that history has been stymied by problems associated with radiocarbon dating of the tufa. These problems go beyond mere “reservoir effect,” especially in deposits that were physico-chemically precipitated. We now have radiocarbon dates on nine P-C deposits, each of which was collected at a separate spring orifice, and each of which was actively accumulating at the time of sampling. The dates (all given in yrs. BP) on these nine samples are as follows: 530 $\pm$ 40; 910 $\pm$ 30; 2000 $\pm$ 40; 2830 $\pm$ 90; 3670 $\pm$ 35; 3695 $\pm$ 35; 5465 $\pm$ 35; 5880 $\pm$ 35; and 6830 $\pm$ 35 (see **Stine Fig. S3.16, left panel**). These nine dates on nine thoroughly modern samples span a range of more than 6300 years; only two of the nine overlap; and two of the nine are less than the present-day reservoir age of the lake (approximately 1400 years). Clearly, each of the deposits carries not only a lake-reservoir signature, but a spring-water signature as well. The untenability of any chronology that is based on radiocarbon dating of physico-chemically precipitated tufa at Mono Lake is obvious.



Figure S3.17. One of the many hundreds of examples of wood encased in bio-precipitated tufa that can be found along the shores of Mono Lake. In 2004, when this photo was taken, the tube still contained datable wood.

We inform this problem by radiocarbon dating woody plant stems that were incorporated into tufa at the time of its precipitation. The plants were killed when the lake, transgressing from drought-induced lowstands, inundated the vegetation growing at spring- and seep-sites; dense calcite, precipitated by algae, accumulated on the stems, ultimately forming an encasement that persists to this day (**Stine Fig. S3.17**). Several different generations of

tufa-encased wood, dating from several different post-drought transgressions, can be discerned based on ^{14}C assays. Among the notable findings: Most (but not all) of Mono Lake's large "delta groves" of tufa towers date from two lake transgressions that occurred approximately 800 and 600 years ago. These major spates of tower formation began with the rapid physicochemical precipitation of "punky" porous-lithoid tufa at the vegetated spring sites; colonization of each upwardly growing pinnacle by algae and by alkali flies resulted in the formation of a dense, slowly accumulating outer coat of bio-precipitated tufa. At many towers both the porous core and the dense encasement contain datable branches. Fluctuations of the lake during subsequent centuries resulted in further modification of the towers, most notably through the physicochemical deposition of flat-bottomed flanges that protrude from the tower flanks, and that mark water level at the time of formation (**Stine Figs. S3.17, S3.18**).



Figure S3.18. One of the complex tufa towers located at County Park tufa grove (aka "Dechambeau tufa") near the westernmost point of the Mono shoreline. The (relatively) bright-white tufa near the tower center is primarily biologically precipitated material, whereas the brownish-gray flanges that protrude from the tower flank are primarily a physicochemical (and much more rapidly accumulating) deposit.

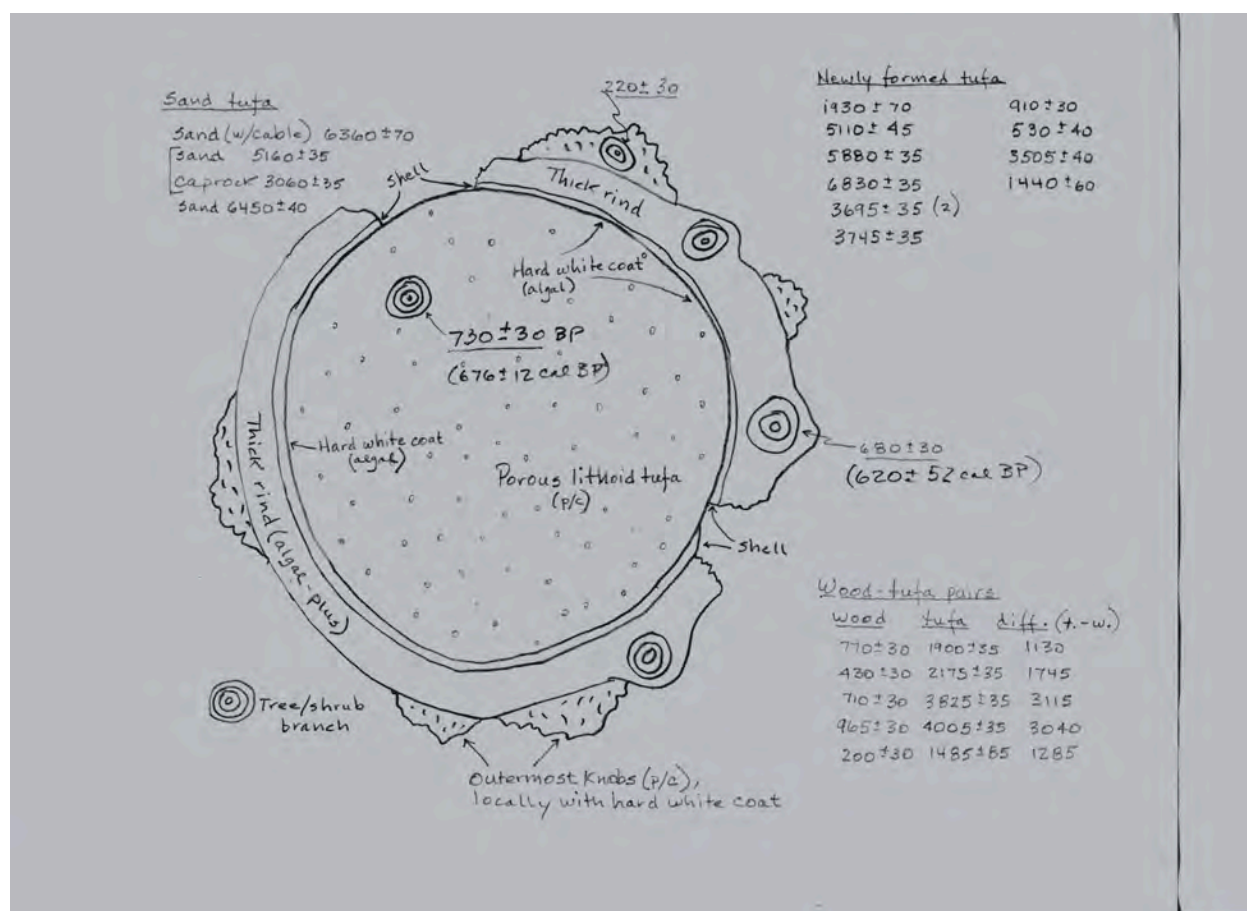


Figure S3.19. Composite cross section based on two broken tufa towers showing the typical layering of bio-precipitated units that encase the physicochemically precipitated porous lithoid tufa of the tower core.

The presence of wood in the tufa allows us not only to radiocarbon date the time of carbonate precipitation, but to contrast the wood date with the apparent ^{14}C age of the tufa coating. Sixteen tufa/wood pairs were assayed. The difference between the wood age and the apparent tufa age for each of the 16 are shown on **Stine Fig. S3.16, right panel**, and will be discussed on-site.

Regarding the peculiarity of the South Tufa grove: South Tufa differs from other Mono Lake groves in several conspicuous ways:

- Broadly speaking, the towers are thinner and taller, and in some places more densely situated, than at any other grove.
- Unlike most other groves, which stand in marshland with active springs and seeps and with hydric vegetation, South Grove occupies xeric land with desert shrubs and few springs or seeps.
- Unlike at most other groves, where the rapidly forming, physicochemically precipitated porous-lithoid core of each tower is encased in a thick, slower growing rind of dense, algally precipitated tufa, the towers at South Grove consist almost entirely of core material, with little sheath.

- Unlike at most other groves, where shrub branches have been incorporated into many of the towers, a careful search of South Grove has turned up not even a single instance of incorporated wood.
- Unlike at most other groves, where the tower bases extend downward into the ground at least to the level of the North Mono tephra, at South Tufa the tower bases stand above not only the North Mono tephra, but above the tephra associated with the volcanic eruptions from the northeast corner of Paoha Island.

Such differences speak to the youthfulness of the South Grove towers, and raise the question as to the source, and the fate, of the water that caused the towers to form. These peculiarities are explained by the diversion of Rush Creek water onto the highly permeable surface of Pumice Valley starting around 1920, as described at our preliminary stop (see **Stine Fig. S1.2**). Water applied to the irrigated land flowed lakeward, subsurface, through the "Panum gap," emerging from the vegetation-free lake floor under as much as 60 ft (18.3 m) of lake water, then streaming upward through the dense Mono brine. Unlike tower formation at other groves, where upward growth was influenced by the rate of lake rise, the towers at South Grove could rapidly lengthen from the lake floor to the lake surface, where, in some cases, they formed "serpent heads" (these will be pointed out in the field). Irrigation of Pumice Valley became intermittent after 1947, and ceased after the 1960s, resulting in the desiccation of the South Tufa lands.

References

- Bischoff, J.L., Fitzpatrick, J.A., and Rosenbauer, R.J., 1993a, The solubility and stabilization of ikaite ($\text{CaCO}_3 \cdot 6\text{H}_2\text{O}$) from 0° to 25 °C: Environmental and paleoclimatic implications for thinolite tufa: *The Journal of Geology*, v. 101, p. 21–33.
- Bischoff, J.L., Stine, S., Rosenbauer, R.J., Fitzpatrick, J.A., and Stafford, T.W., Jr., 1993b, Ikaite precipitation by mixing of shoreline springs and lake water, Mono Lake, California, USA: *Geochimica et Cosmochimica Acta*, v. 57, p. 3855–3865, [https://doi.org/10.1016/0016-7037\(93\)90339-X](https://doi.org/10.1016/0016-7037(93)90339-X).
- Broecker, W.S., Wanninkhof, R., Mathieu, G., Peng, T.-H., Stine, S., Robinson, S., Herczeg, A., and Stuiver, M., 1988, The radiocarbon budget for Mono Lake: An unsolved mystery: *Earth and Planetary Science Letters*, v. 88, p. 16–26, [https://doi.org/10.1016/0012-821X\(88\)90042-8](https://doi.org/10.1016/0012-821X(88)90042-8).
- Cloud, P., and Lajoie, K.R., 1980, Calcite-impregnated defluidization structures in littoral sands of Mono Lake, California: *Science*, v. 210, p. 1009–1012, <https://doi.org/10.1126/science.210.4473.1009>.
- Council, T.C., and Bennett, P.C., 1993, Geochemistry of ikaite formation at Mono Lake, California: Implications for the origin of tufa mounds: *Geology*, v. 21, p. 971–974, [https://doi.org/10.1130/0091-7613\(1993\)021<0971:GOIFAM>2.3.CO;2](https://doi.org/10.1130/0091-7613(1993)021<0971:GOIFAM>2.3.CO;2).
- Dunn, J.R., 1953, The origin of the deposits of tufa in Mono Lake: *Journal of Sedimentary Petrology*, v. 23, p. 18–23, <https://doi.org/10.1306/D4269530-2B26-11D7-8648000102C1865D>.
- Russell, I.C., 1889, Quaternary history of Mono Valley, California: U.S. Geological Survey Eighth Annual Report, pt. 1, p. 261–394.

- Scholl, D.W., and Taft, W.H., 1964, Algae, contributors to the formation of calcareous tufa, Mono Lake, California: *Journal of Sedimentary Petrology*, v. 34, p. 309–319, <https://doi.org/10.1306/74D71041-2B21-11D7-8648000102C1865D>.
- Shearman, D.J., McGugan, A., Stein, C., and Smith, A.J., 1989, Ikaite, $\text{CaCO}_3 \cdot 6\text{H}_2\text{O}$, precursor of the thinolites in the Quaternary tufas and tufa mounds of the Lahontan and Mono Lake Basins, western United States: *Geological Society of America Bulletin*, v. 101, p. 913–917, [https://doi.org/10.1130/0016-7606\(1989\)101<0913:ICOPOT>2.3.CO;2](https://doi.org/10.1130/0016-7606(1989)101<0913:ICOPOT>2.3.CO;2).
- Sieh, K., and Bursik, M., 1986, Most recent eruption of the Mono Craters, eastern central California: *Journal of Geophysical Research*, v. 91, no. B12, p. 12,539–12,571, <https://doi.org/10.1029/JB091iB12p12539>.
- Stine, S., 1987, Mono Lake: The past 4,000 years [Ph.D. dissertation]: Berkeley, University of California, 615 p.
- Stine, S., 1990, Late Holocene fluctuations of Mono Lake, eastern California: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 78, p. 333–381, [https://doi.org/10.1016/0031-0182\(90\)90221-R](https://doi.org/10.1016/0031-0182(90)90221-R).
- Stine, S., 1991a, Extent of riparian vegetation on streams tributary to Mono Lake, 1930–1940: An assessment of the streamside woodlands and wetlands, and the environmental conditions that supported them: Report to the California State Water Resources Control Board and Jones & Stokes Associates, Sacramento, 73 p. plus appendices (Mono Basin EIR Auxiliary Report No. 1).
- Stine, S., 1991b, Geomorphic, geographic, and hydrographic basis for resolving the Mono Lake controversy: *Environmental Geology and Water Sciences*, v. 17, p. 67–83.
- Stine, S., 1992, Past and future toppling of tufa towers and sand tufa at Mono Lake, California: Report to Jones & Stokes Associates, Sacramento, 20 p. (Mono Basin EIR Auxiliary Report No. 9).
- Stine, S., 1994, Extreme and persistent drought in California and Patagonia during mediaeval time: *Nature*, v. 369, p. 546–549, <https://doi.org/10.1038/369546a0>.
- Vorster, P., 1985, A water balance forecast model for Mono Lake, California [M.A. thesis]: Hayward, California State University, Hayward.

Day 1, S4: Panum Crater Block and Ash Flow

FIELD TRIP STOP CONTRIBUTORS: Scott Stine, Emeritus California State University, East Bay

LOGISTICS: Park along the road and join the group for science talk.

WALKING DISTANCE: Optional hike to look at block and ash flow deposit and shorelines

LOCATION: Panum Crater Block and Ash Flow (37.94677, -119.05140)

DIRECTIONS: [South Tufa to Panum Crater Block](#)



Panum Crater Block and Ash Flow and Mono Lake Volcanism

Contributor: Scott Stine, Emeritus California State University, East Bay

One of the final events of the North Mono eruption sequence of 1325-1350 CE was the emplacement of what Sieh and Bursik (1986) called the Panum block-and-ash flow. An aerial photograph (Fairchild Aviation, December 1929, north up) centered on that feature is shown on **Stine Fig S4.1**. Notice there that the very toe of the flow is separated from the lake by a thin white strip of vegetation-free surface. The landward (i.e., upper elevation) edge of that strip marks Mono Lake's "Historic Highstand" of 6428 ft (1959.25 m)—46 vertical ft (14 m) above the present-day lake. The shoreline reached that elevation on July 18, 1919, then began to decline in response to the onset of the "dustbowl drought"; by the time the aerial photo was taken in late 1929 the lake had dropped to 6421 ft (1957.1 m).

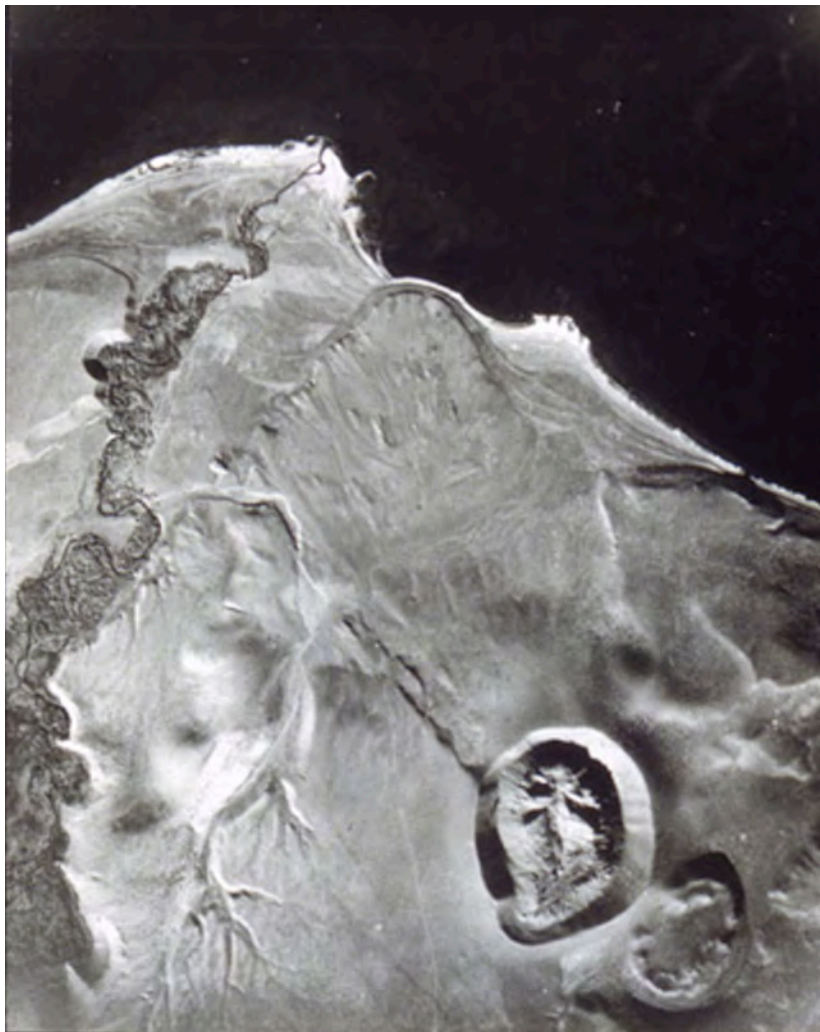


Figure S4.1. Aerial photo (Fairchild Aviation, north up) taken in December, 1929 (lake ele. 6421 ft = 1957 m) showing Panum Dome and Crater at lower right, and the Panum block-and-ash flow just above the latitudinal center line. Note that the very toe of the b&a flow is separated from the lake by a thin white strip of (largely) vegetation-free surface. The landward (i.e., upper elevation) edge of that strip marks Mono Lake's 1919 "Historic Highstand" of 6428 ft (1959.25 m)—46 vertical ft (14 m) above the present-day lake. Notice that to both the right and left of the b&a flow are three higher shorelines, the highest of which marks the "Clover Ranch Highstand" (ele. 6456 ft = 1968 m) of roughly 400 years ago. That CR shoreline can be traced around nearly the entirety of Mono Lake.

Focus next on the eastern (right-most) edge of the flow. Apparent there are three higher shorelines, the highest of which lies at 6456 ft (1967.8 m)—28 ft above the Historic Highstand. That elevation defines the "Clover Ranch Highstand" of roughly 350-400 years ago (think Little Ice Age). That stranded shoreline represents the highest level attained by the lake in nearly 4000 years (Stine dissertation, 1987; Stine, 1990, P^3). Notice on the photo that those same three high shorelines can be traced to the western margin of the flow. We will examine those high shorelines in the field and discuss some of their intricacies, including the differences between their erosional and their depositional expressions.

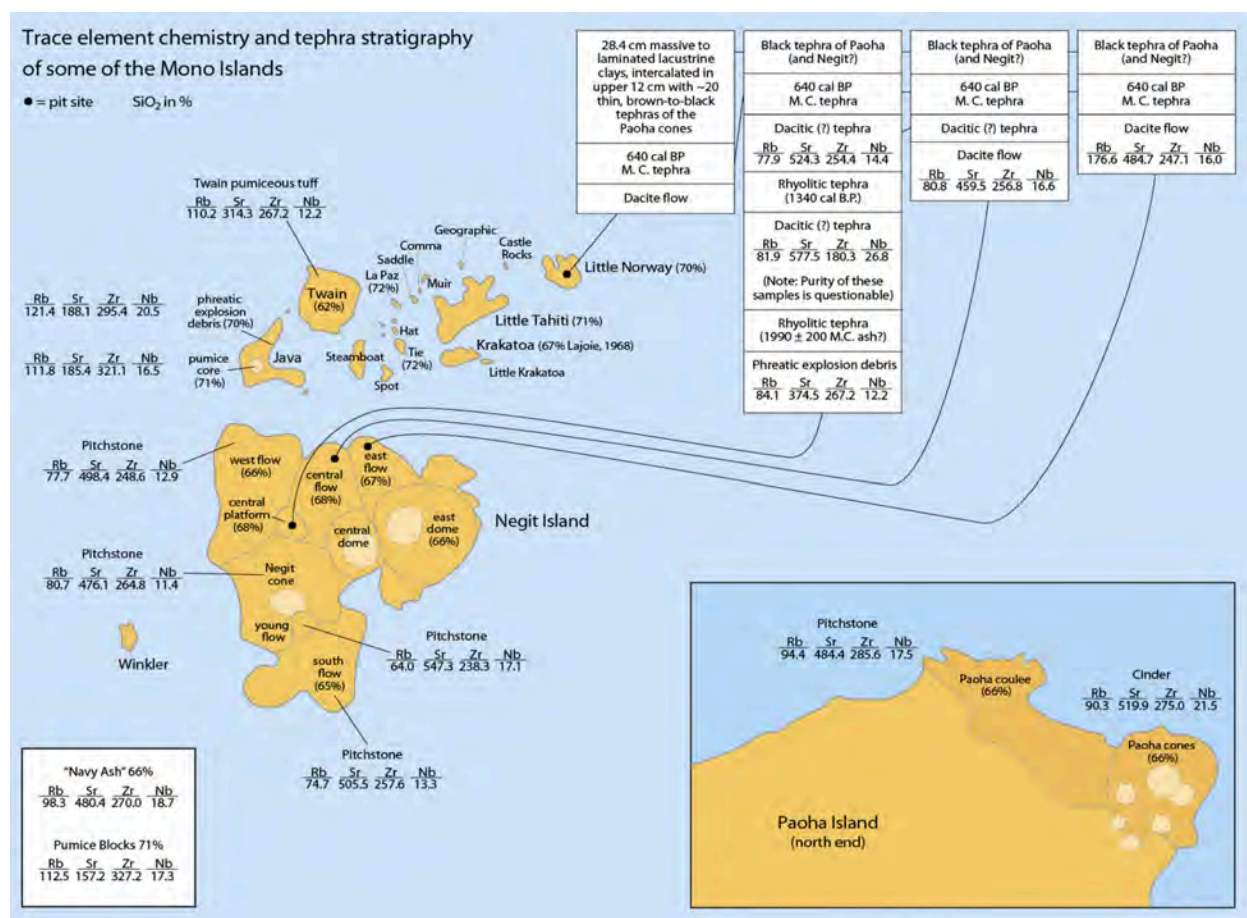


Figure S4.2. Trace element chemistry and tephra stratigraphy of the Mono Islands.

The Clover Ranch shoreline can be traced around almost the entire lake. Moreover, it encircles nearly the entirety of Negit Island, being absent only on the youngest of its lava flows (this strongly suggests that that youngest Negit flow post-dates Clover Ranch time.) We will say more about the ages of this and other portions of Negit Island at our field stop (see **Stine Fig. S4.2**).

What about the age of Paoha? **Stine Fig S4.3** illustrates the island as it appeared in December, 1929. (Note: north is toward the left.) Notice the conspicuous cliff and the adjoining wave-cut platform along the western edge of the island, and the littoral embankments (i.e., depositional beachlines) on the island's northwest corner. These stranded shore features mark the Historic Highstand of 1919, at 6428 ft. Remarkably, neither on the photo nor in the field can a stranded shoreline, erosional or depositional, be found on Paoha above that elevation—not on the explosion pits and lava flows of the island's northern end, not on the hard rock of "hot spring cove," not on the lake sediments that compose the bulk of the island. The absence of a higher shoreline would seem to point to a post-Clover Ranch emergence of Paoha.

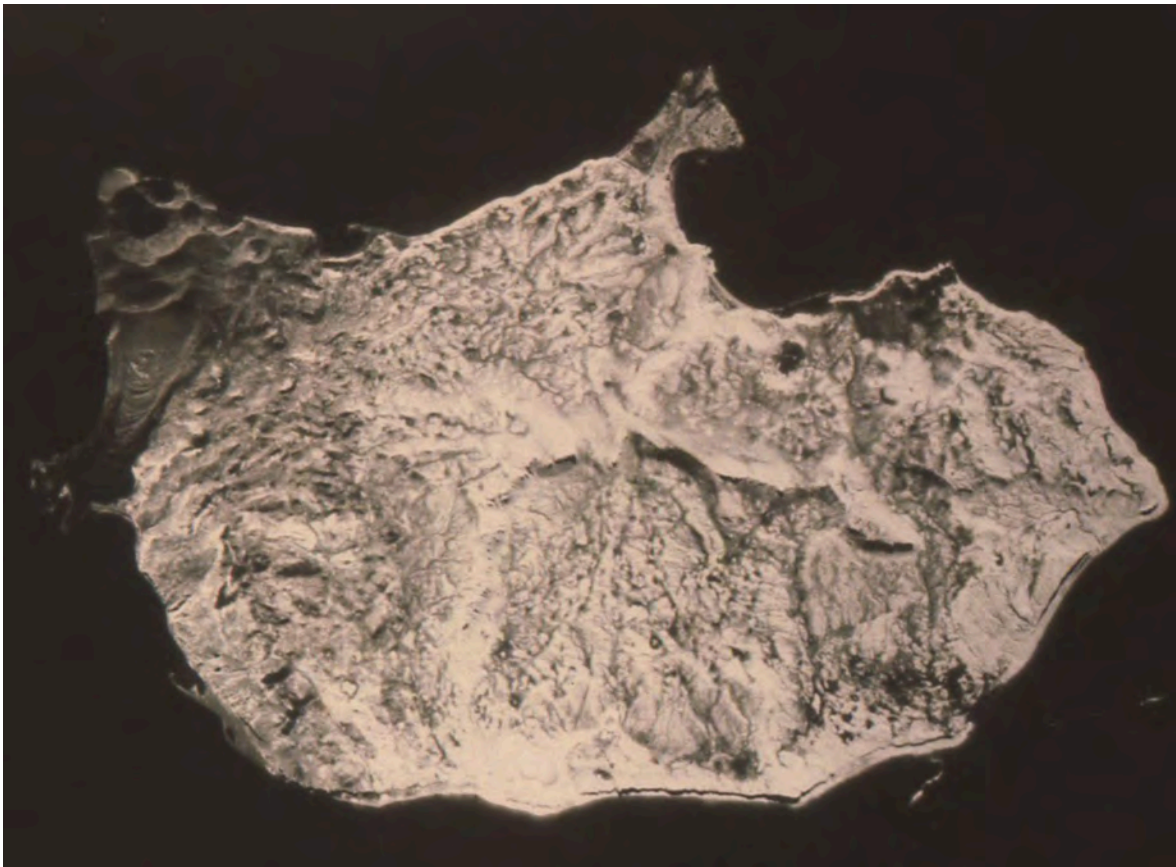


Figure S4.3. Aerial photo (Fairchild Aviation, north to left) showing Paoha Island as it appeared in December, 1929 (lake ele. 6421 ft = 1957 m). The shoreline associated with the historical highstand of 6428 feet is marked on the island by a cliff and wave-cut platform, especially apparent on its western margin. Notice that there appears to be no higher shoreline anywhere on the island, which is to say, the Clover Ranch shoreline, nearly omnipresent around the Mono Lake, seems to be lacking on Paoha. Field examination bears this out.

At our stop we will discuss further evidence—tephro-stratigraphic (**Stine Figs. S4.4, S4.5**), geomorphic, bathymetric, and historic—that corroborates a very recent (post-North Mono eruption, post-Clover Ranch Highstand) emergence of Paoha Island.



Figure S4.4. Stratigraphy of the research pit excavated on "Little Norway" of the Negit Archipelago in 1986. The basal unit is composed of tephra from the North Mono eruption of 600-625 cal BP. It is overlain by ~15-16 cm of laminated biogenic clay, which in turn is overlain by 12-13 cm of silt intercalated with 15-20 brown-to-black tephras of the Paoha cones. The silt is presumed to be sediment that was shed from Paoha at the time of its emergence. Note that the clay is in the process of oxidizing from black to green, indicating that it had only recently been exposed to the atmosphere.



Figure S4.5. Stratigraphy exposed in the excavation on Little Norway as of 1989. Notice that the clay unit has oxidized to white and has shrunk, creating desiccation cracks--again, pointing to subaerial exposure for the first time since its deposition.

References

- Sieh, K., and Bursik, M., 1986, Most recent eruption of the Mono Craters, eastern central California: *Journal of Geophysical Research*, v. 91, no. B12, p. 12,539–12,571, <https://doi.org/10.1029/JB091iB12p12539>.
- Stine, S., 1987, Mono Lake: The past 4,000 years [Ph.D. dissertation]: Berkeley, University of California, 615 p.
- Stine, S., 1990, Late Holocene fluctuations of Mono Lake, eastern California: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 78, p. 333–381, [https://doi.org/10.1016/0031-0182\(90\)90221-R](https://doi.org/10.1016/0031-0182(90)90221-R).

Day 1, S5: Panum Crater Block and Ash Flow

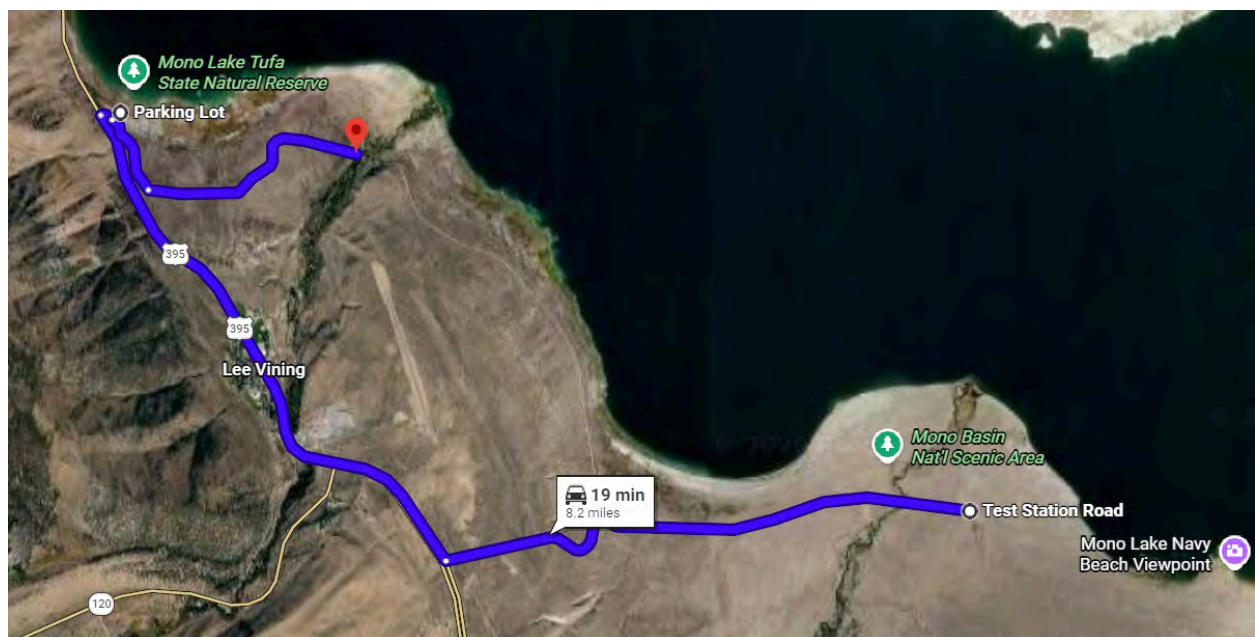
FIELD TRIP STOP CONTRIBUTORS: Zachary Smith (University of California, Berkeley); Scott Stine (Emeritus California State University, East Bay)

LOGISTICS: This is the final stop of the day. Depending on time, we will either go to the Lee Vining Delta Boat Launch or the Old Marina. For those needing assistance with accessibility please coordinate with the organizers and you will be able to drive down to the boat launch.

WALKING DISTANCE: 1.2 miles roundtrip from road to Mono Lake Boat Launch.

LOCATION: Panum Crater Block and Ash Flow (37.93867, -119.02710)

DIRECTIONS: [Panum Crater Block and Ash Flow to Mono Lake Boat Launch](#)



If time is short we will hold presentations at the Mono Lake State Natural Reserve (Old Marina) and anyone who is interested can hike to the boat launch.



Pumice Rafts and Lake Floor Mass Flow Deposits

Contributor: Natalie Homyk (University of California, Berkeley), Zachary Smith (University of California, Berkeley), Michael Manga (University of California Berkeley), Matthew Hornbach (Southern Methodist University)

At this stop we continue our discussion of volcanic unrest in the Mono Basin, this time from the perspective of the lake itself. Strewn along the northwestern shore of Mono Lake are thousands of large blocks of pale, finely vesicular pumice, many of them a meter or more across (Fig S5.1). These blocks did not fall here from an eruption column. They erupted beneath the lake, floated to the surface, drifted across it as pumice rafts, and were stranded when they grounded on the shallow, muddy littoral (Hildreth and Fierstein, 2023). The pumice is therefore a record of a sublacustrine eruption, of the wind that blew on the days that followed it, and of the level of the lake at that time.



Figure S5.1. Pumice blocks around Mono Lake. (A) Pumice blocks and the Mono Lake Boat Launch, and (B) on the southern flank of Paoha Island.

The pumice is one of several ways that recent volcanism has changed the lake floor. The rise of Paoha Island only a few centuries ago steepened the flanks of the central lake basin, triggered landslides, and sent mud flows across the lake floor (Colman et al., 2014). Paoha is still deforming today (Smith et al., 2024). Below we summarize the pumice deposit and its source, look at why giant pumice floats and eventually sinks, and then turn to the lake-floor evidence for mass flows and ongoing deformation around Paoha Island.

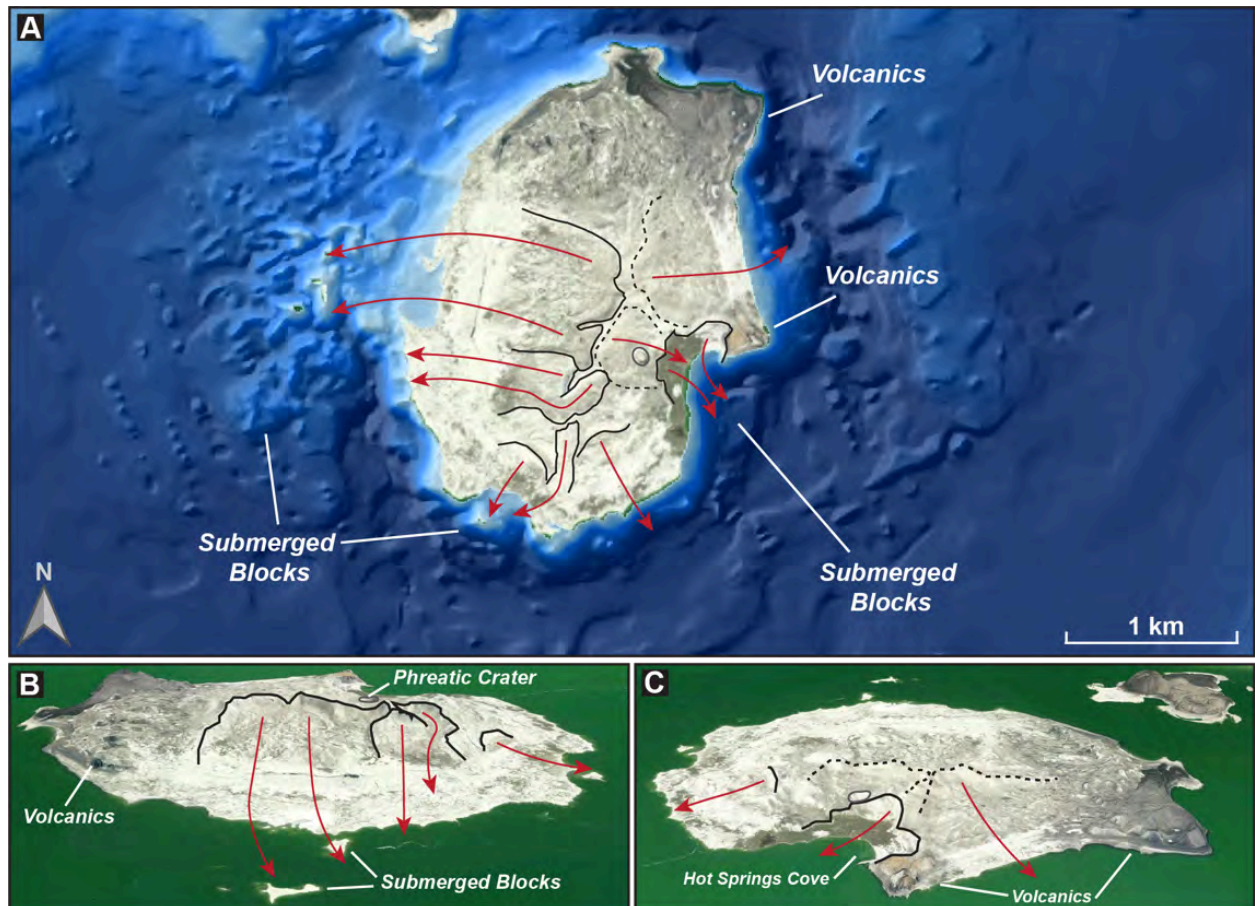


Figure S5.2. Landslide scarps on Paoha Island and landslide blocks partially submerged within Mono Lake. (A) Overhead, (B) east-looking, and (C) west-looking views of Paoha Island.

Giant Pumice Raft

Pumice layers that formed when blocks floated across a lake from an eruption within it are known from only a few places, including La Primavera in Mexico (Clough et al., 1981), Aniakchak in Alaska, Taupo in New Zealand, and Mono Lake (Hildreth and Fierstein, 2023). The beached accumulation of pumice at Mono Lake is 30 to 200 m wide and nearly continuous for 19 km, and in most places it is only one block thick. The blocks are 1 to 12 m across, but few stand more than 1 m tall (Hildreth and Fierstein, 2023). Their flat, tabular shapes and their single-block-thick layout show that this is a raft deposit, not a fall or flow deposit. The blocks were stranded one by one as the raft came into shallow water.

Because the pumice is found only on the northwest shore, the eruption was probably short-lived and the wind direction changed little while the rafts were adrift (Hildreth and Fierstein, 2023). Modern pumice rafts give a sense of how fast this can happen. The raft from the 2012 Havre submarine eruption in the southwest Pacific covered more than 400 km² within

a day, and it drifted at 0.02 to 0.19 m/s, or about 2 to 17 km per day (Jutzeler et al., 2014). At similar speeds, wind-driven pumice could have crossed the 5 to 8 km from its source to the northwest shore of Mono Lake in a day or two.

Source of Pumice Raft

Java Islet has previously been suggested as the source of the pumice. However, the dense lavas and more evolved compositions exposed on Java do not match the beached blocks. Hildreth and Fierstein (2023) instead identified Twain Islet, less than 200 m east of Java, as the source. Twain's outcrops are phenocryst-poor biotite rhyodacite that is similar in composition to the lakeshore blocks and, like them, micropumiceous. Rhyodacite makes up only about 12% of the islet's exposed surface. The rest is reworked sand and pumice.

The best exposure on Twain is a 10-m-thick bedded pile of coarse rhyodacitic ash that dips 10 to 20° northwest or southwest toward the lake. The section is mostly plane-bedded and poor in fine ash. Most beds are 0.3 to 4 cm thick, but a few beds with lapilli are 5 to 20 cm thick, and rarely as thick as 50 cm. Most granules and lapilli, which rarely exceed 5 cm, are micropumiceous, and about 1% are dense black glassy rhyodacite. There are no lithic fragments from older rocks. Free blocks of micropumiceous rhyodacite 1 to 5 m across are enclosed in the bedded ash, along with dikes 1 to 3 m wide and irregular intrusions up to 7 m across. The absence of bomb sags and oxidation points to an eruption beneath the lake. Hildreth and Fierstein (2023) interpret dikes intruding the growing pile of ejecta as the source of the pumice blocks, which were released into the lake and then driven 5 to 8 km north and west by the wind.

This picture of a shallow, dike-fed, largely effusive vent matches what has been learned from deep-marine silicic eruptions. At Havre in 2012, buoyant rhyolite was extruded onto the seafloor, where it rose, quenched, and broke apart into clasts up to several meters across. Clasts that kept their gas rose to form rafts, and those that did not settled as giant pumice on the seafloor (Manga et al., 2018). Other studies have shown that silicic magma extruded into water can make highly vesicular pumice without a strongly explosive eruption, either when parcels of magma foam detach from the vent (Rotella et al., 2013) or through the effect of water pressure at submarine domes (Allen et al., 2010). Twain Islet, where the vent was probably less than a few tens of meters deep, may be a lake-sized version of these processes.

Pumice on Paoha Island

Blocks of pumice up to about 0.5 m across also lie on top of Paoha Island. It is not yet known whether they come from the Twain raft, from a separate eruption, or from vents on Paoha itself. Because Paoha consists largely of lake sediments that were raised above the lake only in the last few centuries (Stine, 2023), one possibility is that these are raft blocks that sank or grounded on the lake floor and were later lifted up with it. If geochemistry shows that the Paoha blocks

match the Twain rhyodacite, the pumice would become a marker bed for measuring how much Paoha was uplifted. If the blocks do not match, they would record another, so far unrecognized, eruption in the lake. We invite discussion on how best to tell these possibilities apart.

Lake-floor mass flow deposits

Because Paoha was uplifted so rapidly, its flanks are unstable. Landslide scarps cut the island, and landslide blocks that have moved down off its slopes now sit partly submerged around its shoreline (Fig. S5.3). High-resolution CHIRP seismic-reflection profiles show that the uplift of Paoha in the last few centuries caused slope failures that produced a thick, fluid mud flow, which spread radially across the lake floor (Colman et al., 2014).

Sediment cores record these events directly. In core CRAD-MONO07-2H, collected southwest of Paoha (Fig. S5.5), a massive, structureless mass flow deposit contains a block of laminated, organic-rich lake sediment that was faulted and tilted during transport (Fig. S5.3). The same interval holds rhyolitic, dacitic, and basaltic glass shards (Fig. S5.4), and the mixture of compositions reflects the range of volcanic sources around the lake. The rhyolitic shards are consistent with the Mono Craters and Paoha Island, and the dacitic shards with the island lavas (Kelleher and Cameron, 1990; Bray et al., 2017). Deposits like these are how lake-floor sediments record volcanic unrest in the lake, even when the source eruptions are small or happen beneath the lake surface.

Figure S5.3. Core CRAD-MONO07-2H-1k-7 from 25-29 cm showing massive mass flow deposit with suspended block of faulted laminated organic rich lake sediments.



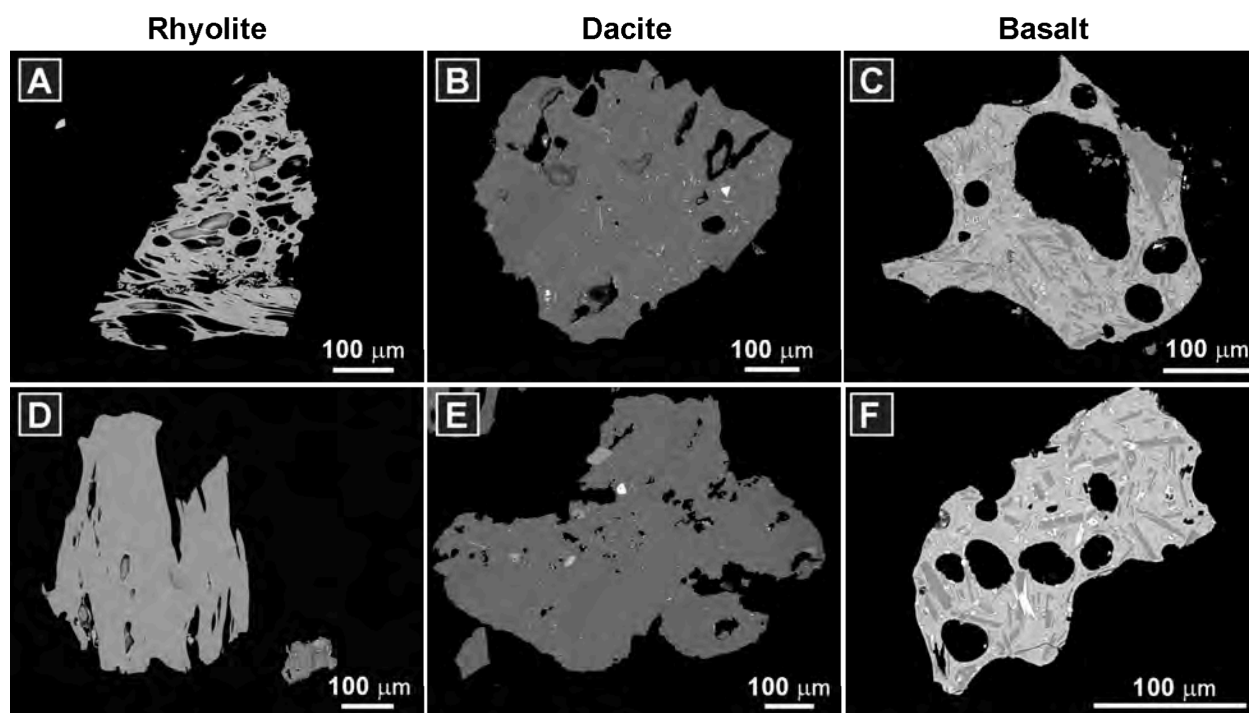


Figure S5.4. Rhyolite, dacite, and basalt tephra from within core CRAD-MONO07-2H-1k-7 from 25-29 cm.

Origin of Deep Holes Around Paoha Island

On the northeast and southwest sides of Paoha Island there are depressions that are twice the mean depth of the lake (Fig S5.5). We hypothesize that the holes were created by 1) volcanic eruption, or 2) post-Paoha Island formation subsidence. InSAR data from 1992 to the present show that the northern third of the island is subsiding at about 1 cm/yr. Airborne LiDAR and field mapping show that this deformation is confined to the rhyodacite and dacite of the northern end of the island and does not extend into the lake. Therefore it's not clear if the depressions are related to recent ground deformation or record some older process. Current work is being carried out to constrain the origin of the depressions in Mono Lake.

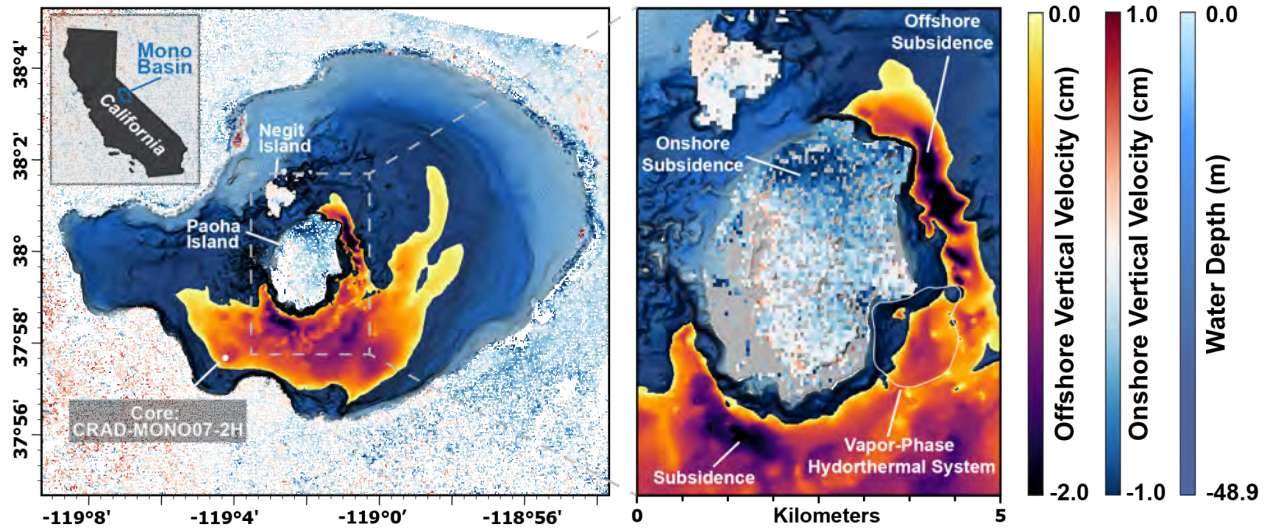


Figure S5.5. Onshore and offshore subsidence in Mono Lake. Onshore is mapped with InSAR and offshore subsidence is constrained with seismic reflection (Coleman et al., 2014; Jayko et al. 2014)

References cited

- Allen, S.R., Fiske, R.S., and Tamura, Y., 2010, Effects of water depth on pumice formation in submarine domes at Sumisu, Izu-Bonin arc, western Pacific: *Geology*, v. 38, p. 391–394, <https://doi.org/10.1130/G30500.1>.
- Bergfeld, D., Evans, W.C., Howle, J.F., and Hunt, A.G., 2015, Magmatic gas emissions at Holocene volcanic features near Mono Lake, California, and their relation to regional magmatism: *Journal of Volcanology and Geothermal Research*, v. 292, p. 70–83, <https://doi.org/10.1016/j.jvolgeores.2015.01.008>.
- Bray, B., Stix, J., and Cousens, B., 2017, Mafic replenishment of multiple felsic reservoirs at the Mono domes and Mono Lake islands, California: *Bulletin of Volcanology*, v. 79, article 54, <https://doi.org/10.1007/s00445-017-1123-y>.
- Bursik, M., and Sieh, K., 1989, Range front faulting and volcanism in the Mono Basin, eastern California: *Journal of Geophysical Research*, v. 94, no. B11, p. 15,587–15,609, <https://doi.org/10.1029/JB094iB11p15587>.
- Clough, B.J., Wright, J.V., and Walker, G.P.L., 1981, An unusual bed of giant pumice in Mexico: *Nature*, v. 289, p. 49–50, <https://doi.org/10.1038/289049a0>.
- Colman, S.M., Hemming, S.R., Stine, S., and Zimmerman, S.R.H., 2014, The effects of recent uplift and volcanism on deposition in Mono Lake, California, from seismic-reflection (CHIRP) profiles: *Journal of Geophysical Research: Solid Earth*, v. 119, p. 3955–3970, <https://doi.org/10.1002/2013JB010726>.

- Fauria, K.E., Manga, M., and Wei, Z., 2017, Trapped bubbles keep pumice afloat and gas diffusion makes pumice sink: *Earth and Planetary Science Letters*, v. 460, p. 50–59, <https://doi.org/10.1016/j.epsl.2016.11.055>.
- Hildreth, W., 2004, Volcanological perspectives on Long Valley, Mammoth Mountain, and Mono Craters: Several contiguous but discrete systems: *Journal of Volcanology and Geothermal Research*, v. 136, p. 169–198, <https://doi.org/10.1016/j.jvolgeores.2004.05.019>.
- Hildreth, W., and Fierstein, J., 2023, Giant pumice deposit from Twain Islet: East of Java; Mono Lake, California: *Geological Society of America Abstracts with Programs*, v. 55, no. 4, <https://doi.org/10.1130/abs/2023CD-387148>.
- Jayko, A.S., Hart, P., Childs, J., Cormier, M., Ponce, D., Athens, N., and McClain, J.S., 2013, Methods and spatial extent of geophysical investigations, Mono Lake, California, 2009 to 2011: U.S. Geological Survey Open-File Report 2013-1113, 18 p., <https://doi.org/10.3133/ofr20131113>.
- Jayko, A., Hart, P.E., Childs, J., and Bursik, M.I., 2023, Late Quaternary deformation along the Mono Lake, Trench Canyon and other offshore faults, Mono Lake Basin, California: *Geological Society of America Abstracts with Programs*, v. 55, no. 4, <https://doi.org/10.1130/abs/2023CD-387109>.
- Jutzeler, M., Marsh, R., Carey, R.J., White, J.D.L., Talling, P.J., and Karlstrom, L., 2014, On the fate of pumice rafts formed during the 2012 Havre submarine eruption: *Nature Communications*, v. 5, article 3660, <https://doi.org/10.1038/ncomms4660>.
- Kelleher, P.C., and Cameron, K.L., 1990, The geochemistry of the Mono Craters–Mono Lake Islands volcanic complex, eastern California: *Journal of Geophysical Research*, v. 95, no. B11, p. 17,643–17,659, <https://doi.org/10.1029/JB095iB11p17643>.
- Manga, M., Fauria, K.E., Lin, C., Mitchell, S.J., Jones, M.P., Conway, C.E., Degruyter, W., Hosseini, B., Carey, R., Cahalan, R., Houghton, B.F., White, J.D.L., Jutzeler, M., Soule, S.A., and Tani, K., 2018, The pumice raft-forming 2012 Havre submarine eruption was effusive: *Earth and Planetary Science Letters*, v. 489, p. 49–58, <https://doi.org/10.1016/j.epsl.2018.02.025>.
- Peacock, J.R., Mangan, M.T., McPhee, D., and Ponce, D.A., 2015, Imaging the magmatic system of Mono Basin, California, with magnetotellurics in three dimensions: *Journal of Geophysical Research: Solid Earth*, v. 120, p. 7273–7289, <https://doi.org/10.1002/2015JB012071>.
- Rotella, M.D., Wilson, C.J.N., Barker, S.J., and Wright, I.C., 2013, Highly vesicular pumice generated by buoyant detachment of magma in subaqueous volcanism: *Nature Geoscience*, v. 6, p. 129–132, <https://doi.org/10.1038/ngeo1709>.
- Smith, Z., Hornbach, M., Manga, M., Barth, A., and Adourian, S., 2023, Magma-tectonic structural controls on sublacustrine hydrothermal systems in Mono Lake, California: *Geological Society of America Abstracts with Programs*, v. 55, no. 4, <https://doi.org/10.1130/abs/2023CD-387286>.
- Smith, Z., Manga, M., Hornbach, M., Lindsay, D., Homyk, N., DeMarines, J., and Bürgmann, R., 2024, Constraints on the distribution and mechanism of active subsidence on Paoha Island

in Mono Lake, California: Geological Society of America Abstracts with Programs, v. 56, no. 5, <https://doi.org/10.1130/abs/2024AM-403208>.

Stine, S., 1990, Late Holocene fluctuations of Mono Lake, eastern California: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 78, p. 333–381, [https://doi.org/10.1016/0031-0182\(90\)90221-R](https://doi.org/10.1016/0031-0182(90)90221-R).

Stine, S., 1994, Extreme and persistent drought in California and Patagonia during mediaeval time: *Nature*, v. 369, p. 546–549, <https://doi.org/10.1038/369546a0>.

Stine, S., 2023, Late Holocene island volcanism at Mono Lake, California: Geological Society of America Abstracts with Programs, v. 55, no. 4, <https://doi.org/10.1130/abs/2023CD-388067>.

Whitham, A.G., and Sparks, R.S.J., 1986, Pumice: *Bulletin of Volcanology*, v. 48, p. 209–223, <https://doi.org/10.1007/BF01087675>.

Mono Lake Boat Launch Immediately West of Lee Vining Creek

Contributor: Scott Stine, Emeritus California State University, East Bay

From where we park we will walk approximately 0.6 mi to the Mono shoreline. In doing so, notice the changes in ground color and vegetation that we encounter along the way; note too the littoral embankments that we cross—the first at elevation 6428 ft (the Historic Highstand of CE 1919) then several at around 6416-6418 ft (~1956 m), where the lake stood between 1938 and 1947. The drop in lake level that occurred after 1947 (ultimately to the "Historic Lowstand" of 6372 ft = 1942.2 m in late December 1981) is attributable to diversion of Mono Basin stream water southward to Los Angeles.



Figure S5.6. The Lee Vining Creek delta in the early 1980s, near the time of the historical lowstand of 6372 ft (1942.2 m). The drop in lake level to that low elevation, caused by Los Angeles's stream diversions, forced the stream to incise its delta. That incision exhumed large, buoyant pumice blocks that had floated to shore following their eruption from one or two of the Negit islets—"Twain" and/or "Java." Notice a line of the light-colored pumice blocks can be seen running R-L across the stream bed.

As we near the lake we will see dozens of large tufa-covered blocks that are scattered along the beach. At places where the tufa has been broken away the blocks can be seen to be composed of pumice (toss a piece onto the lake surface—it floats!). Tens of thousands of these buoyant, biotite-bearing rhyodacite blocks litter the lake's western and southwestern shorelands at elevations at and below 6392

ft (1948.3 m). They are the product of a sublacustrine eruption centered at Twain and Java islets, immediately north of Negit Island. Once expelled, the floating blocks were blown toward shore (then at 6392 ft), where they shoaled and developed a rind of tufa.



Figure S5.7. One of the pumice blocks on the stream bed of Lee Vining Creek, here shown in stratigraphic context. The cobble gravel that immediately overlies the block contained a fresh-looking pinecone that

radiocarbon dated (at 2 sigma) to 1390-1530 cal BP, thus providing a minimum age on the block eruption.

Just west of our present position flows Lee Vining Creek. Over the centuries following the deposition of the pumice blocks, the creek prograded its delta, burying the blocks in as much as 12 ft (3.6 m) of sediment. The recent diversion-induced drop in lake level forced the creek to incise its delta, which exhumed the blocks (**Stine Fig S5.6**), and provided views of them in stratigraphic context (**Stine Fig S5.7**). Notice that the block shown in **Fig S5.7** is overlain by two Mono Craters tephras—the North Mono ash of 600-625 cal BP (here reworked and accumulated by the stream) and the South Mono ash of 1302-1356 cal BP (here exposed as a slightly disrupted airfall sequence). (2-sigma dates from the Digital Data Base of Bursik and Sieh, 2013.) Clearly the blocks pre-date the South Mono eruption.

One of the blocks exposed in the (river-right) wall of the incised channel was in direct contact with overlying, fluvially deposited, woody debris including pinecones, the freshest one of which radiocarbon dated (at 2 sigma) to 1390-1530 cal BP, thus further constraining the minimum possible age of the Twain/Java eruption. One of my research pits excavated low on the shorelands at the Mono County Marina (~1.4 mi west of our present position) exposed a layer of pumice gravels expelled during that eruption. These gravels apparently lie above the tephra from the Northwest Coulee eruption of 1520-1658 cal BP, and unequivocally lie above the Wilson Butte tephra of 1611-1710 cal BP (Bursik and Sieh, 2013). For now it seems safe to bracket the timing of the Twain/Java pumice eruption with the outermost constraining dates of 1390 cal BP (pinecones) and 1710 cal BP (Wilson Butte tephra).

Just a few dozen steps to the west are good examples of wood that was incorporated into bio-precipitated tufa; a few dozen steps to our east is one of the trees that was killed when Mono Lake rose from one of its lowstands associated with the Medieval Climatic Anomaly of Stine (1994, *Nature*). I'm hoping we can visit both before walking back to the cars.

References

- Bursik, M., and Sieh, K., 2013, Digital database of the Holocene tephras of the Mono-Inyo Craters, California: U.S. Geological Survey Data Series 758, <https://doi.org/10.3133/ds758>.
- Stine, S., 1994, Extreme and persistent drought in California and Patagonia during mediaeval time: *Nature*, v. 369, p. 546–549, <https://doi.org/10.1038/369546a0>.

FOP 2026 DAY 2



Day 2, S6: Scenic Overlook

FIELD TRIP STOP CONTRIBUTORS: Michael Oskin (University of California Davis), William Hammond (Nevada Geodetic Lab., Nevada Bureau of Mines and Geology, University of Nevada, Reno), Fred Phillips (New Mexico Tech), Chris Wills (retired)

LOGISTICS: This stop is at a scenic overlook along Highway 395.

WALKING DISTANCE: Minimal.

LOCATION: Scenic Overlook along Highway 395 (37.518887, -118.604094)

DIRECTIONS: [Brown's Owens River Campground to 395 Vista Point](#)



Dextral Slip along the Sierra Nevada Range Front north of Owens Valley

Contributors: Michael Oskin, Yvonne Leon, Eduardo Alatorre-Acevedo, Sofia Franck, Elias Kai Yepishin, Ryan Lynch, Loïs Papin, and Erik Perkins (Department of Earth and Planetary Sciences, University of California, Davis)

We report new mapping and geochronology documenting active dextral strike-slip faulting along the Sierra Nevada Range front at Round Valley and Long Valley. Our preliminary results, combined with earlier work to the north from Bursik and Sieh (1989), show that a component of Walker Lane dextral shear remains localized along the Sierra Nevada range front north of the northern terminus of the Owens Valley fault. This slip helps to reconcile kinematic models for the Mina Deflection, confirming dextral slip along the Sierra Nevada range front predicted by Delano et al. (2019).

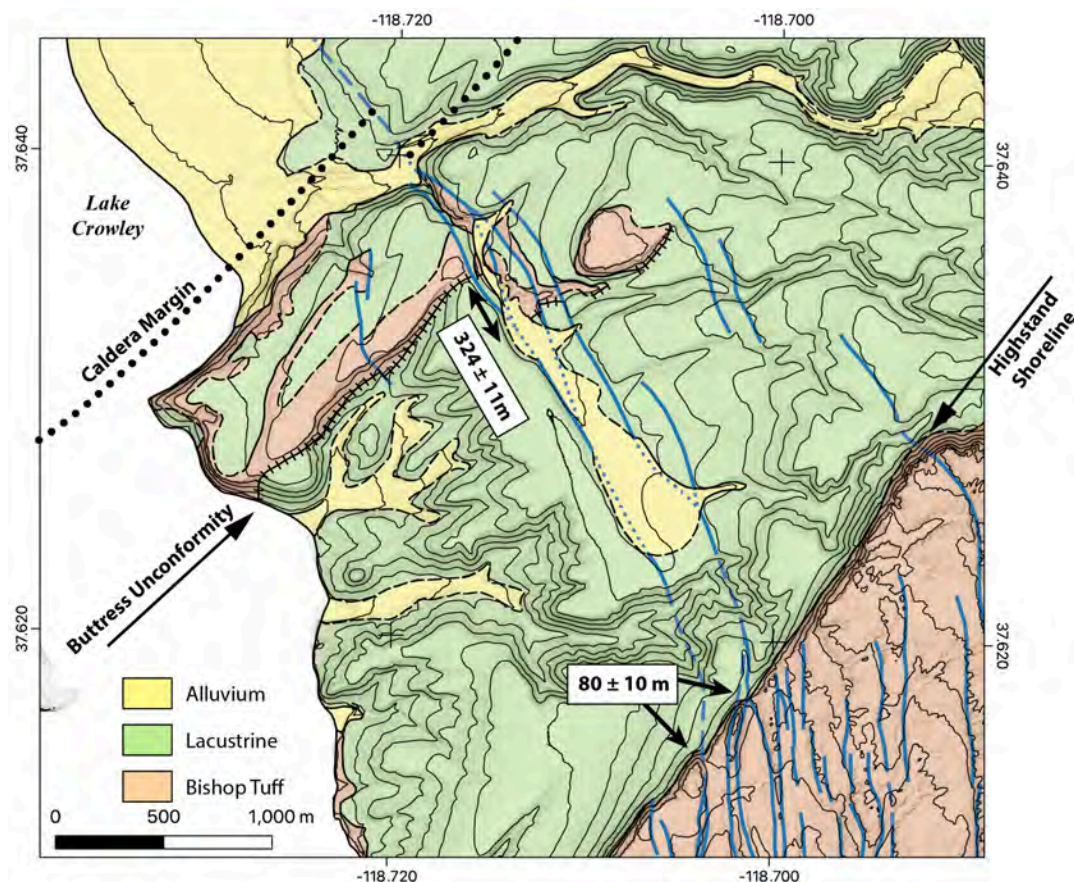


Figure S6.1. Geologic map of the Long Valley fault adjacent to Lake Crowley Reservoir. The Long Valley fault dextrally offsets a buttress unconformity interpreted to have formed in association with collapse of Long Valley Caldera at 767.1 ± 0.9 ka. Dextral offsets also occur at the shoreline cliff formed by Long Valley Lake highstand, dated 159 ± 23 ka.

The Long Valley fault (Figure S6.1) cuts the southwest margin of the Long Valley Caldera immediately east of Lake Crowley. Here we document 324 ± 11 m of dextral offset of a linear buttress unconformity between Bishop Tuff and overlying lacustrine strata of Pleistocene Long Valley Lake. The buttress is clearly exposed by shoreline erosion by Lake Crowley and consists of a southeast-facing escarpment of Bishop Tuff, overlain by tufa-mantled tuff boulders and cobbles, and capped by lacustrine marl and tufa-cemented sand and gravel beds. A thin veneer of pebbly littoral sediment overlies the Bishop Tuff northwest of the buttress. We interpret the buttress to be an antithetic fault scarp formed during the collapse of the Long Valley Caldera during or shortly after the eruption of the Bishop Tuff. The buttress was subsequently flooded by Long Valley Lake and preserved beneath lacustrine sediment. Using the 767.1 ± 0.9 ka age of the Bishop Tuff (Crowley et al., 2007) as the formation age of the buttress, we calculate a long-term dextral slip rate of 0.42 ± 0.02 mm/yr for the Long Valley fault.

Along strike to the southeast, the Long Valley fault splits into two strands that exhibit dextral offset of the highstand shoreline cliff of Long Valley Lake (Figure S6.1). This lake occupied the Long Valley Caldera up until the penultimate glacial maximum, when it overtopped an embayment within the caldera margin to reoccupy the tuff-filled paleovalley of the ancestral Owens River (Hildreth and Fierstein, 2016). We used a ^{10}Be depth profile and a co-located IRSL sample to date a sand and gravel spit formed at the highstand shoreline, and combined this with published constraints to estimate the age of lake highstand to be 159 ± 23 ka (Leon, 2025). From offsets of the highstand cliff-face, we estimate a combined 80 ± 10 m of dextral slip after correction for apparent dextral offset due to the normal component of fault slip since the lake highstand. The combined offset and refined lake age yield a dextral slip rate of 0.5 ± 0.1 mm/yr for the Long Valley fault, consistent with its long-term rate.

We also find evidence of dextral slip along the Round Valley fault, which bounds the steep Sierra Nevada range front between Bishop Pass and Long Valley. We analyze lidar-derived topography to delineate moraine crests at Horton Creek and Pine Creek (Figure S6.2). Both sites exhibit subtle dextral offsets of moraine crests in addition to normal slip. Figure 3 shows a map of moraines at Horton Creek, refined using the lidar topography from previous mapping by Phillips et al. (2009). Here, the north (left) Tioga stage 2 lateral moraine (Figures S6.3 and S6.4) exhibits 7 ± 2 m normal offset and 15 ± 3 m of dextral offset. Using an age range of 20 to 25 ka from Phillips et al. (2009), these offsets yield a dip-slip rate of 0.3 ± 0.1 mm/year and a dextral slip rate of 0.7 ± 0.2 mm/year. Similar fault strike and dextral slip rates suggest that the Round Valley fault and Long Valley fault may be connected by faulting through the Owens River gorge area. Such a connection is not yet confirmed.

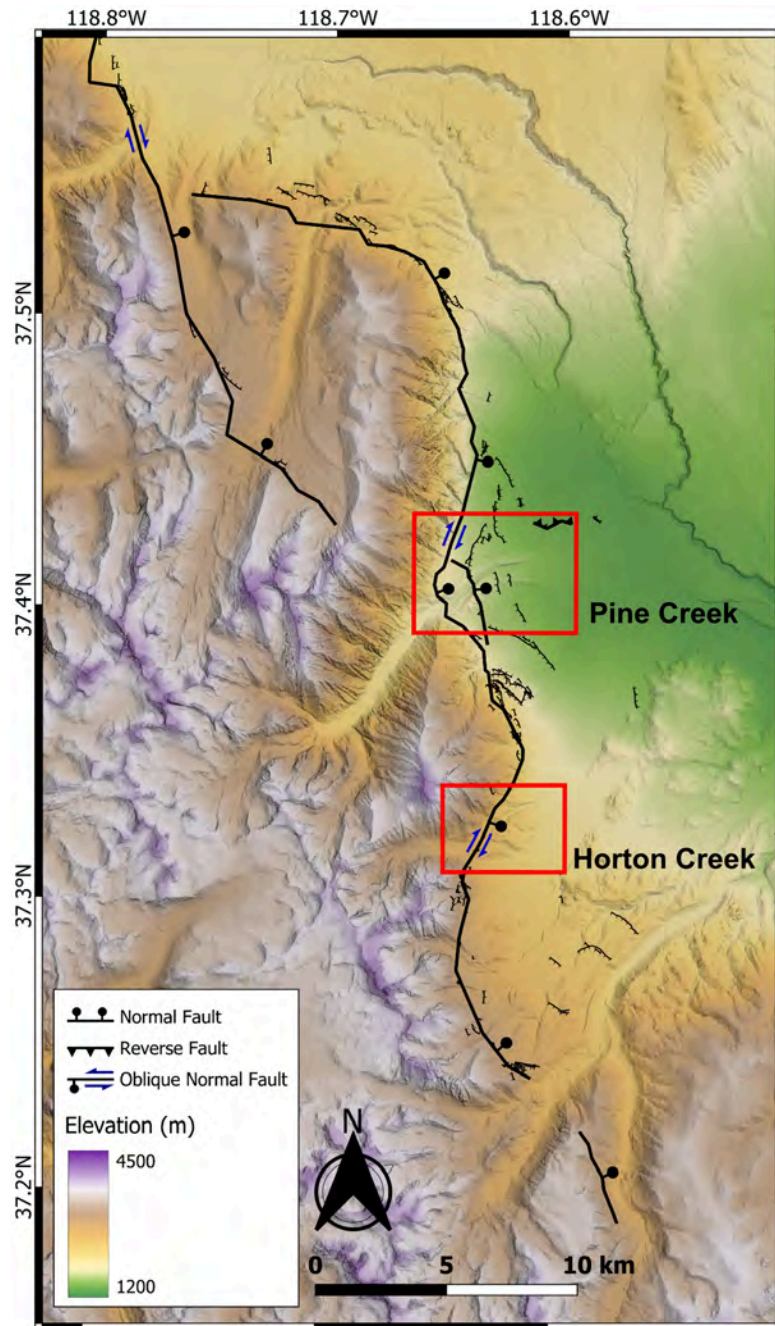


Figure S6.2. Active traces of the Round Valley fault and southern Hilton Creek fault. Evidence for dextral slip occurs where moraine crests are offset at Pine Creek and Horton Creek.

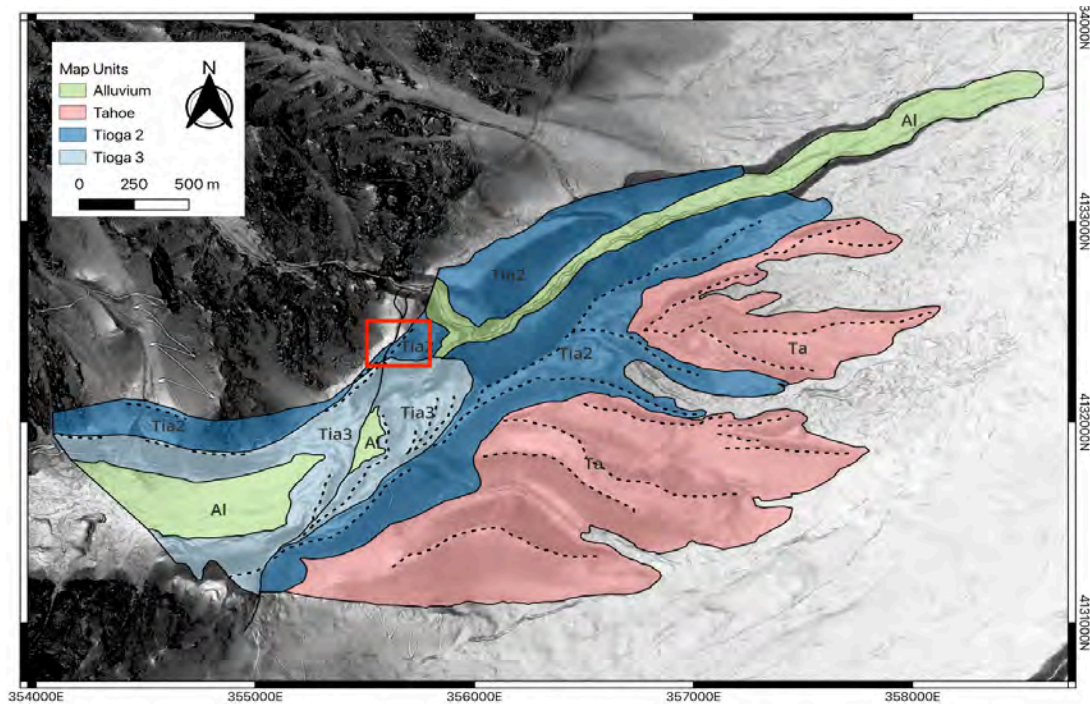


Figure S6.3. Map of Tahoe and Tioga stage 2 and 3 moraines at Horton Creek. Modified from Phillips et al. (2009) on the basis of lidar topography and field observations.

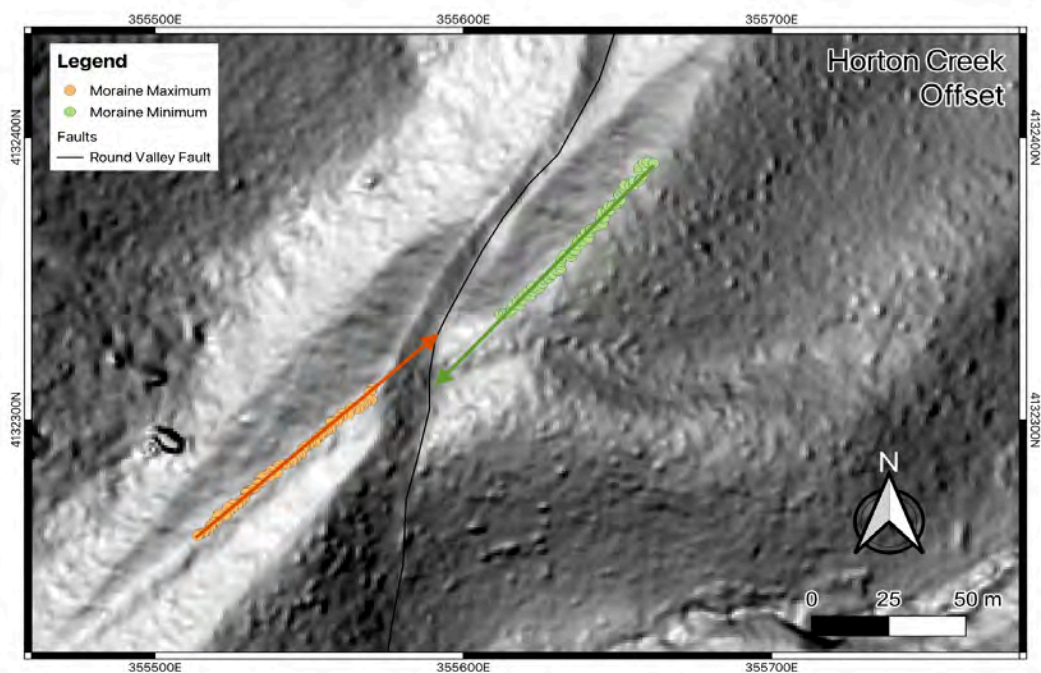


Figure S6.4. Lidar-derived hillshade image of north (left) lateral moraine in Tioga stage 2 deposits at Horton Creek. Points automatically extracted from serial swath profiles of moraine crest. These points define best-fit vectors we projected to the fault with 60° dip. The offset of this moraine crest indicates 7 ± 2 m of normal slip and 15 ± 3 m of dextral slip since 20 to 25 ka.

References

- Bursik, M., and Sieh, K., 1989. Range front faulting and volcanism in the Mono Basin, eastern California, *Journal of Geophysical Research: Solid Earth*, v. 94(B11) pp. 15587-15609.
- Crowley, J. L., Schoene, B., & Bowring, S. A., 2007. U-PB dating of zircon in the Bishop Tuff at the millennial scale. *Geology*, 35(12).
- DeLano, K., Lee, J., Roper, R., Calvert, A., 2019. Dextral, normal, and sinistral faulting across the eastern California shear zone–Mina deflection transition, California-Nevada, USA. *Geosphere* v. 15, 1206–1239.
- Hildreth, W., and Fierstein, J., 2016. Long Valley Caldera Lake and reincision of Owens River Gorge: U.S. Geological Survey Scientific Investigations Report 2016–5120, 63 p.
- Leon, Y.J., 2025. Slip Rate and Hazard Potential of the Long Valley Fault Zone near Mammoth, CA, Masters Thesis, University of California, Davis, 81p.
- Phillips, F.M., Zreda, M., Plummer, M.A., Elmore, D., Clark, D.H., 2009. Glacial geology and chronology of Bishop Creek and vicinity, eastern Sierra Nevada, California. *Geological Society of America Bulletin* v. 121, 1013–1033.

A Reevaluation of Moraine Chronology at Pine Creek, Inyo County, California

Contributors: Fred M. Phillips (New Mexico Tech), Jeffery Lee (Central Washington University and Colorado School of Mines) and Kimberly Blisniuk (San Jose State University)

While writing a GSA Field Guide article that used offset moraine crests to quantify fault displacements, we became interested in the Pine Creek moraines because they are offset by the Round Valley fault. We reassessed the chronology of the moraines in Lee et al. (in review). This contribution is a brief summary of that work. Please see the guidebook article, once published, for additional details on the work.

Pine Creek (Inyo County, northern Owens Valley) possesses one of the most prominent sets of moraines on the eastern front of the Sierra Nevada. They are the only moraines in the Owens Valley to extend a significant distance onto the valley floor, beyond the mountain front. The moraines indicate that the glacier advanced 4.5 km over the valley floor and the maximum moraine height is about 200 m above the valley. Several episodes of glacial advance at different times are evident from inspection of the moraines.

Given their prominence, the chronology of the Pine Creek moraines is still poorly known. They were initially mapped by Bateman (1965) as part of a mapping effort that extended over several 15 minute quadrangles. He estimated where they fit into the conventional Sierra moraine chronology of his day, but did not perform any measurements to support these assignments. Subsequently, Robinson (1980) and Berry (1994) measured relative weathering indicators at numerous locations. Robinson (1980) focused on only two indicators, boulder frequency and percent of fresh boulders, at a large number of measurement sites (up to 75), whereas Berry (1994) utilized 11 weathering indicators, but at far fewer locations. Finally, Bach and Elliott-Fisk (1996) measured soil profile development index (PDI) at 15 sites. These weathering studies yielded only relative age sequences. The only quantitative moraine chronology available has been five ^{10}Be and ^{26}Al ages measured as part of a regional study by Nishiizumi et al. (1993). The ages were calculated using now-obsolete cosmogenic production rates. We have recalculated these using modern production rates. Additionally, Dr. Nishiizumi had measured three unpublished cosmogenic samples that he graciously allowed us to use.

We used the well-established glacial sequence chronology for the eastern Sierra Nevada to classify the Pine Creek deposits. These glacial stages and their ages are as follows: Tahoe 170-146 ka, Tioga 4 ~28 ka, Tioga 2/3 22-17 ka, Tioga 4 17-16 ka. Minor glacial advances subsequent to 16 ka are found only high in the mountains. See Lee et al. (in review) for supporting information on the sources of information for this chronological sequence.

The recalculated and new cosmogenic ages and our remapping of the moraine sequence are shown in sketch-map form in Figure 1. They are from the same perspective as Stop 6 on Day 2 of this field trip. Our moraine mapping, although independent, is very similar to that of Bateman (1965). Our moraine designations are founded on the correspondence of Bateman's relative sequence estimates



Figure S6.5. Map showing delineation of glacial units at Pine Creek and locations of cosmogenic nuclide samples. Ages are in ka and are measured using ^{10}Be except one indicated as ^{26}Al . Cosmogenic data were measured by Dr. K. Nishiizumi. Dashed line illustrates the portion of the moraine that we designate as $\text{Qti}_{2/3}$ that was mapped as Qtay by Bateman (1965). Note that image perspective is looking approximately south.

with Nishiizumi's cosmogenic ages, as can be verified by comparing the cosmogenic ages shown on the map with the moraine chronology listed above.

There is one exception to the generally good correspondence between Bateman's estimated sequence assignments and the cosmogenic ages. This is for the right-lateral moraine area that we have mapped as $\text{Qti}_{2/3}$. Bateman designated the area south of the dotted line in

Figure S6.5 as Qtay, or ‘Younger Tahoe’. This is a major discrepancy, since the generally accepted age of the Tahoe glaciation is seven times older than the Tioga 2/3 (i.e., >145 ka versus ~20 ka). We have chosen to accept the single cosmogenic age rather than Bateman’s estimate, but this discrepancy deserves further examination.

The large amount of relative dating data collected at Pine Creek offers the opportunity for a comparison of the relative and cosmogenic data and to address the right-lateral age discrepancy. The measures of extent of weathering performed by Robinson, Berry, and Bach and Elliot-Fisk yielded incommensurate data having a wide variety of units. To make them comparable, we converted them to Z-scores. The Z-score of an individual sample out of a set of samples of a particular parameter is the sample value minus the mean value for all samples, divided by the standard deviation of all samples. A sample whose measurement value is the same as the mean for all samples will have a Z-score of zero, one with a measurement value equal to the mean plus one standard deviation will have a Z-score of +1, and one equal to the mean minus one standard deviation will have a Z-score of -1. For the Pine Creek weathering data set, relatively unweathered samples will have negative Z-scores and strongly weathered ones will have positive Z-scores. We lumped all of the Tioga substages into one “Ti” category. The measurements performed on the portion of the moraine sequence mapped differently by us and Bateman we term “intermediate”.

In Figure S6.6 we have plotted the difference of the mean Z values for the Tahoe minus the Tioga samples on the X-axis versus the Z score of each unit on the Y-axis. Samples plotting toward the bottom of the graph thus show little weathering while those near the top are indicating strong weathering. A small difference between the Tahoe and Tioga Z scores (i.e., left side) indicates that the parameter is insensitive to weathering, while a large difference (right side) shows that the parameter changes a lot with weathering and is a good indicator.

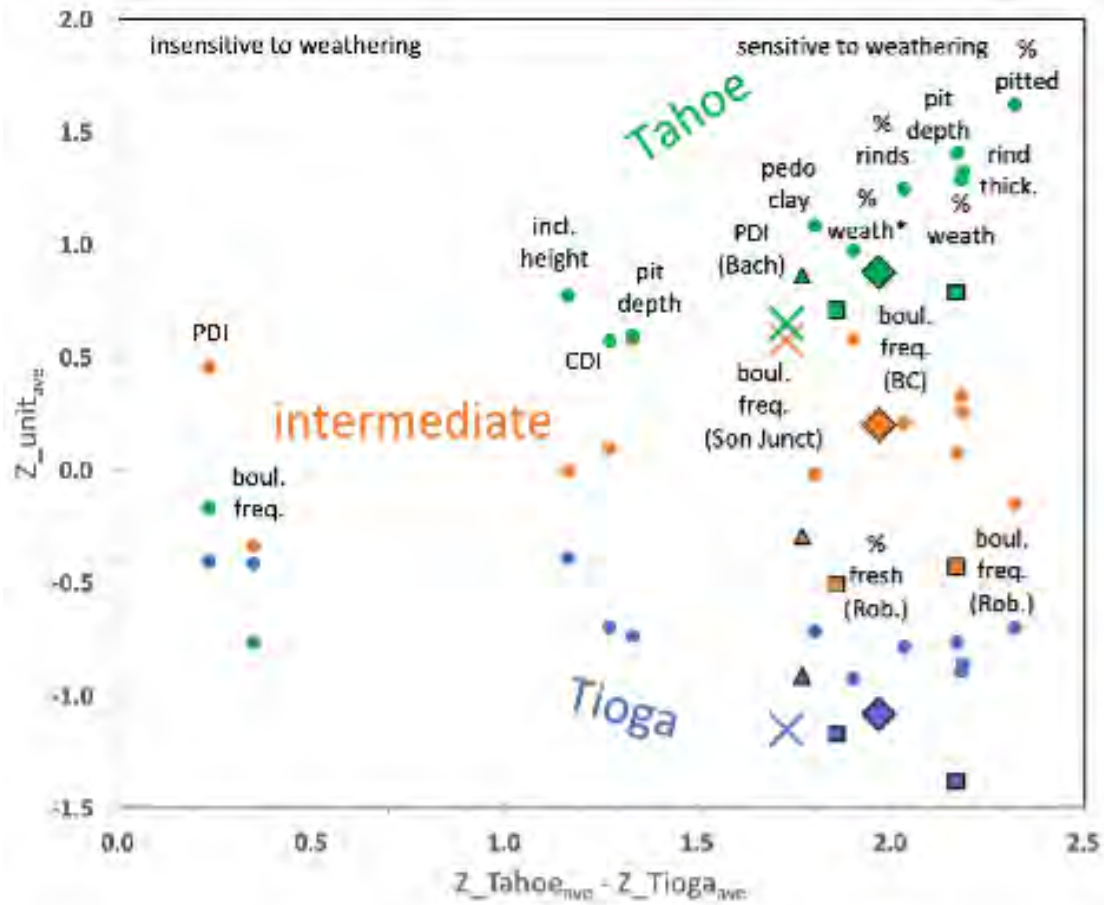


Figure S6.6. Average Z-scores for weathering parameters measured on moraines assigned to standard glacial stages at Pine Creek, as a function of the difference between the Tahoe and the Tioga average values. Parameters falling to the left show little variation with age and those on the right increase with age and are therefore useful for evaluating moraine age. Tioga are in blue, Tahoe in green, and the orange data labeled ‘intermediate’ are from the lower, inner, right lateral moraine, which has a ^{10}Be age much younger than that assigned by Bateman (1965), and whose age is therefore in question. See text for information on calculations. Data are from Robinson (1990) (squares), Berry (1994) (small dots), and Bach and Elliot-Fisk (1996) (X symbol). Abbreviations: PDI = soil profile development index, CDI = clast disintegration index, bould. freq. = boulder frequency, incl. height = inclusion height, pedo clay = pedogenic clay mass, % weath* = percent weathered by observation, % weath = percent weathered by hammer blow, rind thick. = rind thickness, Son Junct = Sonora Junction from Clark (1976), BC = Bloody Canyon from Sharp and Birman (1963), Rob. = from Robinson (1980), Bach = from Bach and Elliot-Fisk (1996).

The data from the Tahoe moraines is in green and plots near the top of the graph. The data from the Tioga moraines is in blue and groups near the bottom of the graph. The weathering data thus corroborate the cosmogenic sequence assignments. The data from the

moraine portion alternatively mapped as Qtay by Bateman and $Qti_{2/3}$ by us are in orange and do indeed plot between the Tahoe and Tioga samples. Is the positioning of these samples consistent with them having an actual Tioga 2/3 age? To address this question, we compared the positions of the 'intermediate' samples from Pine Creek with those from two other moraines sequences that historically were assigned 'Tenaya' designations (i.e., between Tahoe and Tioga), but when cosmogenically dated had clearly Tioga ages. These were from Sonora Junction and Bloody Canyon (BC). Our 'intermediate' samples span most of the range between the Tahoe and Tioga samples. If those from the other studies with numerous cosmogenic ages are close to the Tioga samples, lower than most of our intermediate data points, this would cast doubt on the single cosmogenic Tioga age for the moraine. But if they are closer to the Tahoe moraine, higher than most of our intermediate samples, then the weathering data are likely consistent with the cosmogenic age. The latter is the case. We therefore conclude that the weathering data support the Tioga age designation.

With this corroboration, we feel that the moraine age designations in (Lee et al., in review), and shown here, are fairly secure.

References

- Bach, A. J., and Elliott-Fisk, D. L., 1996, Soil development on late Pleistocene moraines at Pine Creek, east-central Sierra Nevada, California: *Physical Geography*, v. 17, p. 1–28.
- Bateman, P. C., 1965, Geology and tungsten mineralization of the Bishop District, California: U.S. Geological Survey Professional Paper 470, Professional Paper 470, 208 p. p.
- Berry, M. E., 1994, Soil-geomorphic analysis of late-Pleistocene glacial sequences in the McGee, Pine, and Bishop Creek drainages, east-central Sierra Nevada, California: *Quaternary Research*, v. 41, p. 160–175.
- Clark M. M. (1967) Pleistocene glaciation of the drainage of the West Walker River, Sierra Nevada, California: Ph.D., Stanford University, Palo Alto, California, 170 pp.
- Lee, J., Phillips, F. M., and Blisniuk, K., in review, Late Oligocene to Present-day slip along the southern Sierra Nevada normal fault, California (USA): Implications for uplift of the Sierra Nevada, *in* Jones, C. H., Henry, C. D., Cassel, E. J., and Wakabayashi, J., eds., *Uplift History of the Sierra Nevada, USA: Field Guide*, Geological Society of America.
- Nishiizumi, K., Kohl, C. P., Arnold, J. R., Dorn, R., Klein, J., Fink, D., Middleton, R., and Lal, D., 1993, Role of in situ cosmogenic nuclides ^{10}Be and ^{26}Al in the study of diverse geomorphic processes: *Earth Surface Processes and Landforms*, v. 18, p. 407–425.
- Robinson, O. G., 1980, The Pleistocene glacial sequence of Pine Creek, Round Valley, California [M.S. Thesis]: California State University at Northridge, 97 p.
- Sharp, R. P. and Birman, J. H. (1963) Additions to classical sequence of Pleistocene glaciations, Sierra Nevada, California: *Geological Society of America Bulletin*, v. 74, p. 1079-1086.

Active tectonics and fault slip rates from GNSS measurements around the Mono Basin, Walker Lane, western Great Basin

Contributor: William C. Hammond (Nevada Geodetic Lab., Nevada Bureau of Mines and Geology, University of Nevada, Reno)

Active faults in and around the Mono Basin in the central Walker Lane are part of a complex system of crustal deformation composed of many dextral, sinistral and normal faults that work in concert to relieve accumulating transtensional strain. We seek to understand the strain rates and associated fault slip rates in order to relate them to longer-term development of the landscape and to quantify seismic hazards.

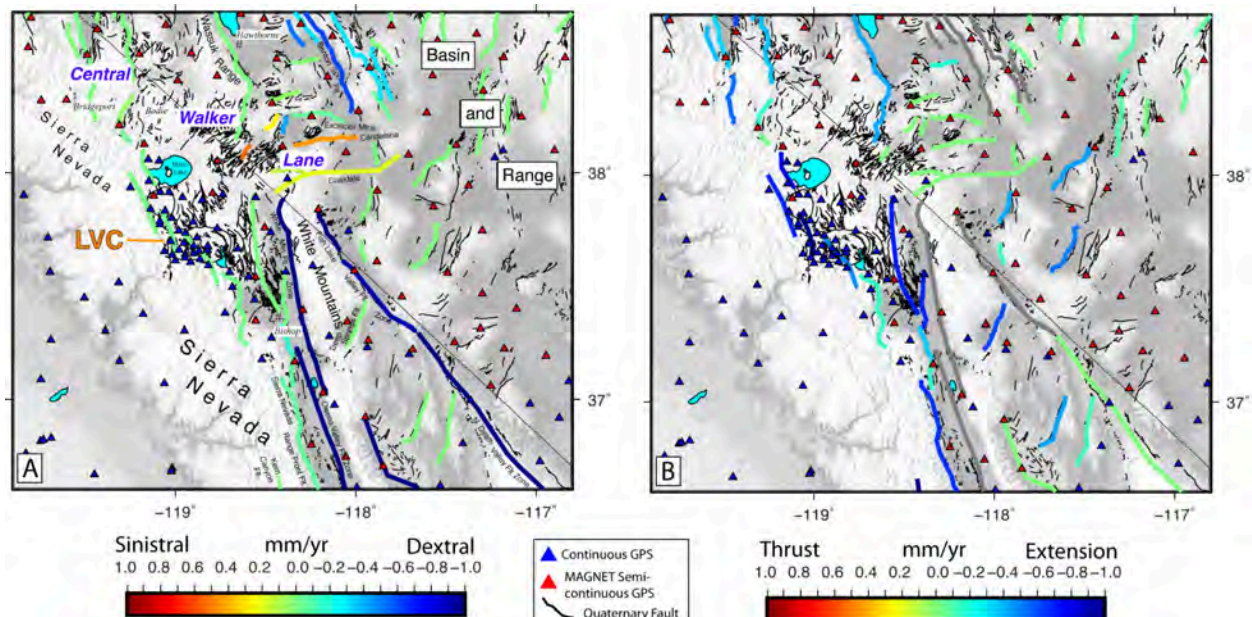


Figure S6.7A and S6.7B. A) Strike slip rates superimposed on topography of Sierra Nevada and Walker Lane. Triangles show locations of continuous GNSS stations (blue) and semi-continuous MAGNET GNSS stations (red). Faults with thick colored lines are those in the NSHM (Hatem et al., 2023). Color scales are clipped at ± 1 mm/yr to emphasize small slip rate variations. LVC indicates Long Valley Caldera, selected faults, towns and geographic features are named. A) same as B) except showing normal components of slip rates. Figure is a combination of Fig. 5 of Hammond et al., (2024) and Fig. 1 of Hammond et al., (2019).

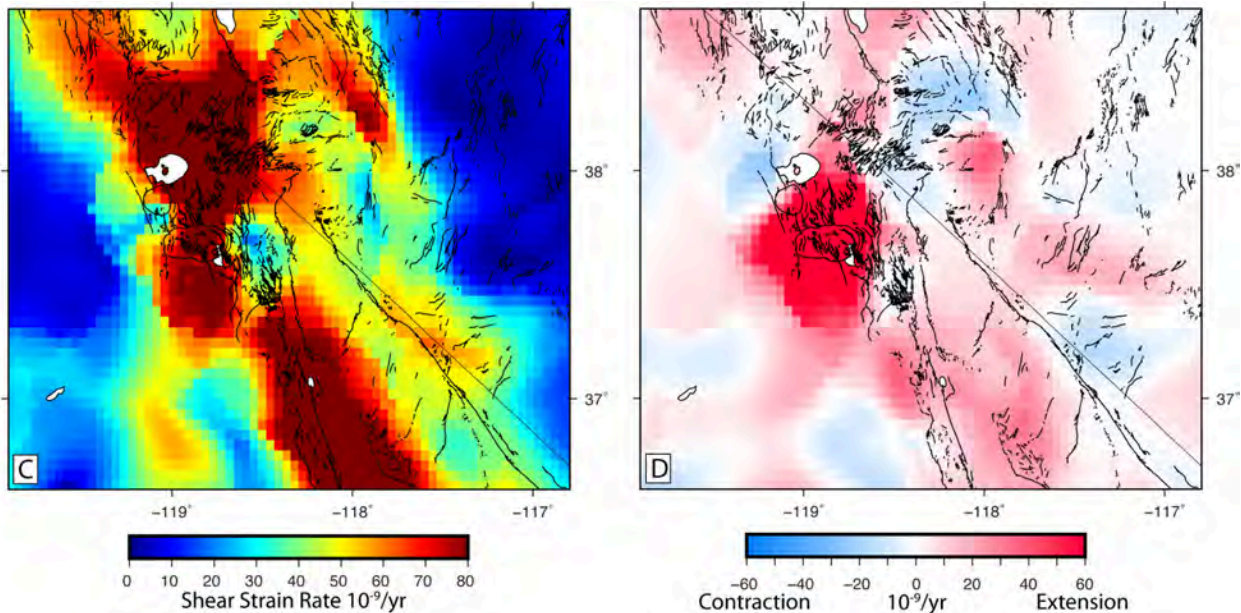
The Sierra Nevada to the west behaves as a mostly rigid boundary condition, with its northwest translation driving strain accumulation on Walker Lane faults. The Southern Walker Lane consists of dextral strike slip, normal and oblique faults in the Owens Valley, Fish Lake Valley, and Sierra Nevada range front. The Mina Deflection is a group of sub-parallel left lateral fault

systems that facilitate a right step of strain rates to the east of Mono Basin (Wesnousky, 2005; DeLano et al., 2019). The Central Walker Lane consists of a series of en echelon asymmetric basins with faults accommodating shear and extension (Wesnousky et al., 2012; Pierce et al., 2021). These diverse yet interconnected tectonic domains make a complicated kinematic framework that motivates us to develop system-scale models of the deformation.

We use geodetic GNSS networks to obtain ultra-precise positioning which over time captures the active and current motion of Earth's crust in a synoptic way. The data measure the broad-scale tectonic driving movements of plates that generate stresses around fault systems which will ultimately be released in earthquakes. With precision near 0.1 mm/yr they can be used to make maps of strain rates which help identify which faults are being loaded most rapidly and in what way (Figure S6.7A and S6.7B).

To estimate the relative motion between crustal blocks we model their movements with the GNSS observations as constraints. However, the complexity of the system makes development of bespoke models onerous. Databases include many faults that are short, overlapping, incompletely mapped, with enigmatic interconnections, making it a challenge to translate geodetic velocities and strain rates (Figure S6.7C and S6.7D) to fault slip rates. The modeling study of Hammond et al., (2024) addresses this problem by developing new ensemble block modeling algorithms to facilitate robust block motion and slip rate imaging.

Figure S6.7 is an amalgamation of results from two studies, a Walker Lane-scale block modeling analysis (Hammond et al., 2024) and a study of tectonic and magmatic deformation around the Long Valley Caldera (LVC) (Hammond et al., 2019). The LVC is an active magmatic system that drives inflation, centered on the resurgent dome (Montgomery-Brown, 2015). The magmatic system generates a radial extension signal in the GNSS networks that overlaps with the tectonic deformation, further complicating solutions for slip rates. These studies evaluate the overlapping signals of strain from Walker Lane tectonics and transient, time-variable deformation from LVC magmatism.



Figures S6.7C and D. A) Shear component of Walker Lane strain rate derived from GNSS measurements and B) dilatational component of strain rate (area growth is positive, contraction negative). Stations locations contributing GNSS velocity data are shown in A and B, above. Black line segments are fault traces from the USGS Quaternary Fault and Fold Database (USGS and NBMG, 2011). The strongest red zone in D is the radial growth from the Long Valley Caldera magmatic inflation.

The results in Figure S6.6A and 1B show that while strike slip and normal slip are partitioned into different zones, they both tend to follow the areas with high strain rates. In this model the Mono Lake Fault has a normal slip rate of 0.8 mm/yr, resolved to horizontal extension. Slip rate estimates for all USGS National Seismic Hazard Map (NSHM) fault sources in the Walker Lane are available in Hammond et al., (2024).

References

- DeLano, K., Lee, J., Roper, R., & Calvert, A. (2019). Dextral, normal, and sinistral faulting across the eastern California shear zone—Mina deflection transition, California-Nevada, USA. *Geosphere*, 15(4), 1206-1239.
- Hatem, A.E., Collett, C.M., Briggs, R.W., Gold, R.D., Angster, S.J., Powers, P.M., Field, E.H., et al., (2021). Earthquake geology inputs for the U.S. National Seismic Hazard Model (NSHM) 2023 (western U.S.) (ver. 3.0, December 2023): U.S. Geological Survey data release. <https://doi.org/10.5066/P9AWINWZ>.
- Hammond, W.C., C. Kreemer, G. Blewitt, (2024). Robust imaging of fault slip rates in the Walker Lane and Western Great Basin from GPS data using a multi-block model approach, *J. Geophys. Res. - Solid Earth*, 129, e2023JB028044. <https://doi.org/10.1029/2023JB028044>.

- Hammond, W.C., C. Kreemer, I. Zaliapin, G. Blewitt, (2019). Drought-triggered magmatic inflation, crustal strain and seismicity near the Long Valley Caldera, Central Walker Lane, *J. Geophysical Res. - Solid Earth*, 124(6), p. 6072–6091. <https://doi.org/10.1029/2019JB017354>.
- Montgomery-Brown, E. K., C. W. Wicks, P. F. Cervelli, J. O. Langbein, J. L. Svarc, D. R. Shelly, D. P. Hill, and M. Lisowski (2015). Renewed inflation of Long Valley Caldera, California (2011 to 2014), *Geophys. Res. Lett.*, 42, 5250–5257. <https://doi.org/10.1002/2015GL064338>.
- Pierce, I.K.D., Wesnousky, S.G., Owen, L.A., Bormann, J.M., Li, X., Caffee, M., (2021). Accommodation of Plate Motion in an Incipient Strike-Slip System: The Central Walker Lane. *Tectonics* 40, e2019TC005612. <https://doi.org/10.1029/2019TC005612>.
- U.S. Geological Survey and Nevada Bureau of Mines and Geology, (2011). Quaternary Fault and Fold Database for the United States, accessed April, 2011, at: <https://www.usgs.gov/natural-hazards/earthquake-hazards/faults>.
- Wesnousky, S.G., (2005). Active faulting in the Walker Lane. *Tectonics* 24, 35 pp. <https://doi.org/10.1029/2004TC001645>.
- Wesnousky, S.G., Bormann, J.M., Kreemer, C., Hammond, W.C., Brune, J.N., (2012). Neotectonics, geodesy, and seismic hazard in the Northern Walker Lane of Western North America: Thirty kilometers of crustal shear and no strike-slip? *Earth and Planetary Science Letters* 329–330, 133–140. <https://doi.org/10.1016/j.epsl.2012.02.018>.

Earthquake Scenarios for the Long Valley Caldera-Mono Lake Area

Contributor: Chris Wills (retired)

This viewpoint looking out along the Sierra front is a good place to discuss the processes and rates that have led to this magnificent landscape. It's also a good place to think about how these forces will manifest themselves as future earthquakes and volcanic eruptions. How will future earthquakes and volcanic eruptions affect the society and infrastructure of this region? How do we as geologists inform society of the potential impacts and help prepare for these events?

One format that has been developed for helping society, and particularly emergency managers, understand the scale and impact of major earthquakes and other natural disasters is the Planning Scenario, developed in the 1980's and 1990's by the California Geological Survey (CGS) and California Office of Emergency Services (beginning with Davis et al, 1982). Planning Scenarios were further developed to include more analysis of social and economic impacts in the USGS/CGS ShakeOut Scenario (Jones et al, 2008) and expanded to other hazards in the ARKStorm Scenario (Porter et al, 2011). Following these, the USGS California Volcano Observatory (CalVO) planned a similar multi-hazards project to develop potential earthquake and volcanic eruption scenarios for the Long Valley Caldera-Mono Lake area. For this project, CGS would prepare earthquake scenarios, CalVO would prepare volcanic eruption scenarios and then work with emergency managers and engineers to estimate the impacts on society and infrastructure from these potential events.

To start this project, CalVO contracted with CGS to develop the earthquake scenarios. The report describing these (Chen, et al, 2014) was published by CGS and USGS in 2014. A general description of volcanic hazards and potential impacts was published by USGS in 2019 (Mangan et al, 2019). To the best of my knowledge, the original vision of planning scenarios describing the potential impact of future earthquakes and volcanic eruptions has not been realized.

The following text and figures are excerpted from the 2014 report, which is available on the USGS publications web site.

Abstract

As part of the U.S. Geological Survey's (USGS) multi-hazards project in the Long Valley Caldera-Mono Lake area, the California Geological Survey (CGS) developed several earthquake scenarios and evaluated potential seismic hazards, including ground shaking, surface fault rupture, liquefaction, and landslide hazards associated with these earthquake scenarios. The results of these analyses can be useful in estimating the extent of potential damage and economic losses because of potential earthquakes and in preparing emergency response plans.

The Long Valley Caldera-Mono Lake area has numerous active faults. Five of these faults or fault zones are considered capable of producing magnitude ≥ 6.7 earthquakes. These five faults are the Fish Slough, Hartley Springs, Hilton Creek, Mono Lake, and Round Valley Faults. Earthquake scenarios are intended to depict the potential consequences of significant earthquakes. They are not necessarily the largest or most damaging earthquakes possible. Earthquake scenarios are both large enough and likely enough that emergency planners should consider them in

regional emergency response plans. Each scenario is illustrated with maps of seismic shaking potential and fault displacement, liquefaction, and landslide potential.

Scenario Earthquake Hazard Results

Round Valley M7.0 Scenario

An **M7.0** earthquake on the Round Valley Fault would produce strong ground shaking in the southern Long Valley, Round Valley, and Bishop Creek areas. Instrumental intensity, median PGA, and median SA at 1.0 second for this scenario are shown in Figure S6.8.

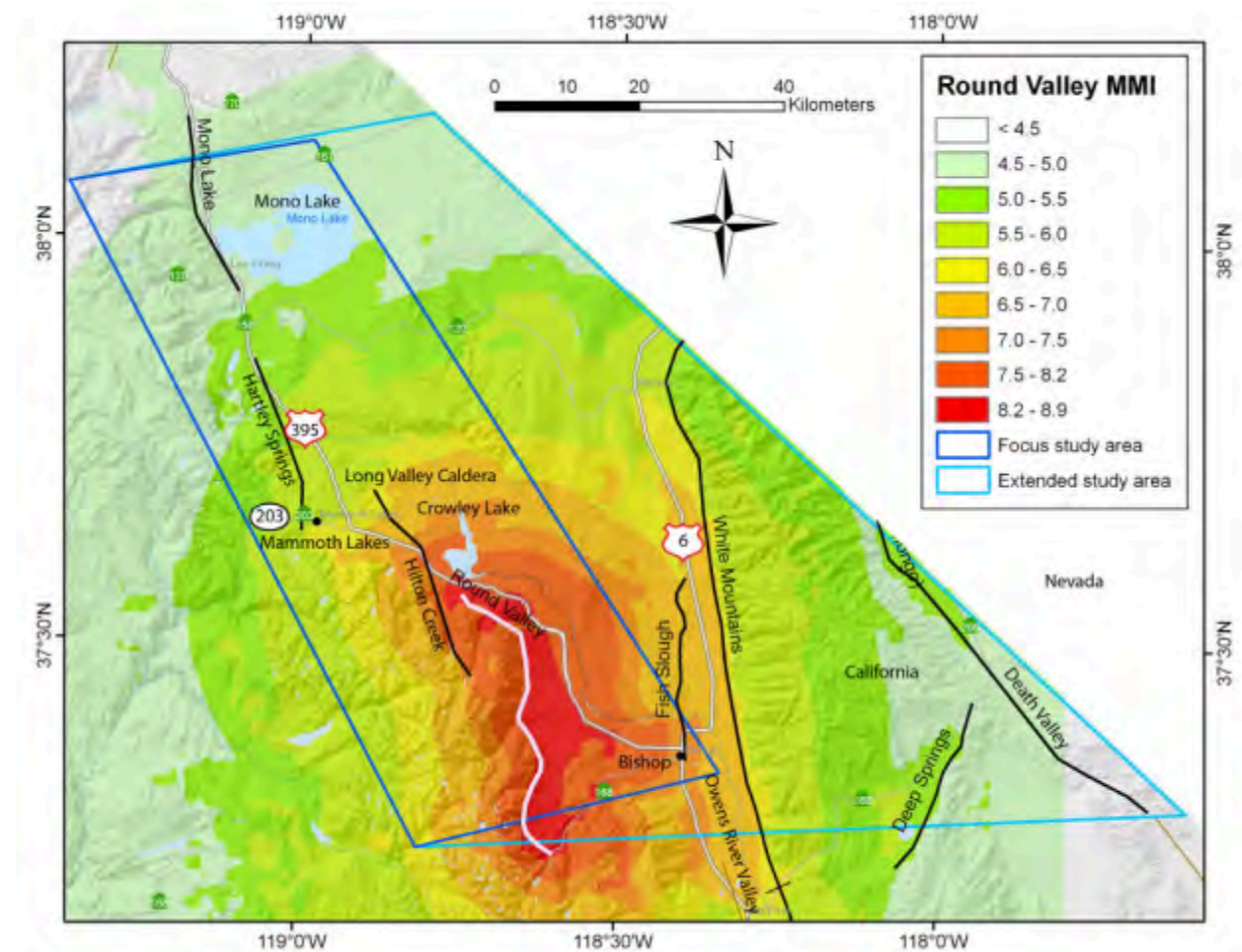


Figure S6.8. Maps showing ground motion hazards of an M7.0 earthquake on Round Valley Fault (a. intensity, b. median peak ground acceleration, and c. median spectral acceleration at 1.0 second). Only a. shown here, see Chen et al (2014) for b. and c.

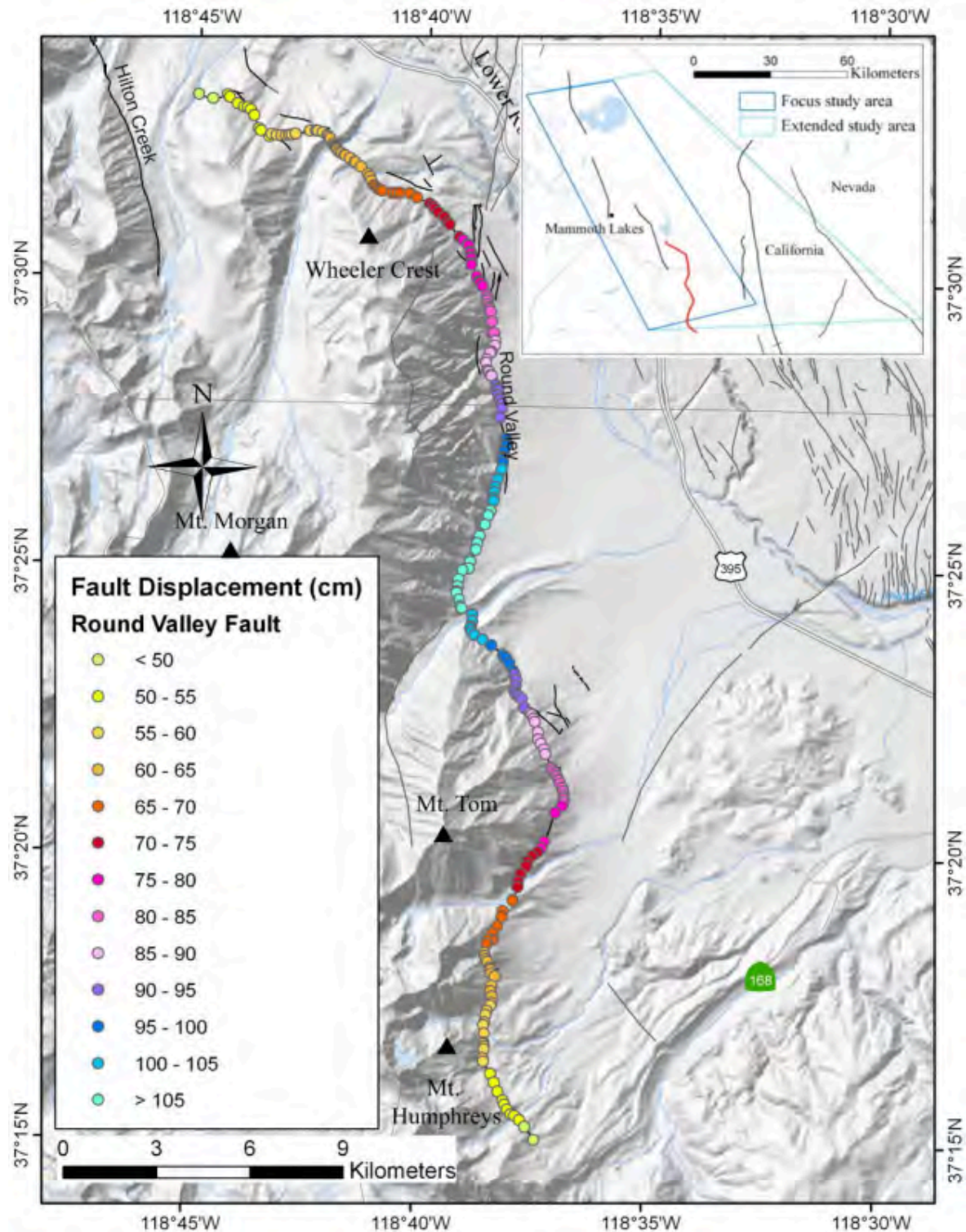


Figure S6.9. Map showing median deterministic principal fault displacement along fault strike for the Round Valley M7.0 earthquake scenario. Dots represent simplified fault trace and are color-coded by calculated fault displacement.

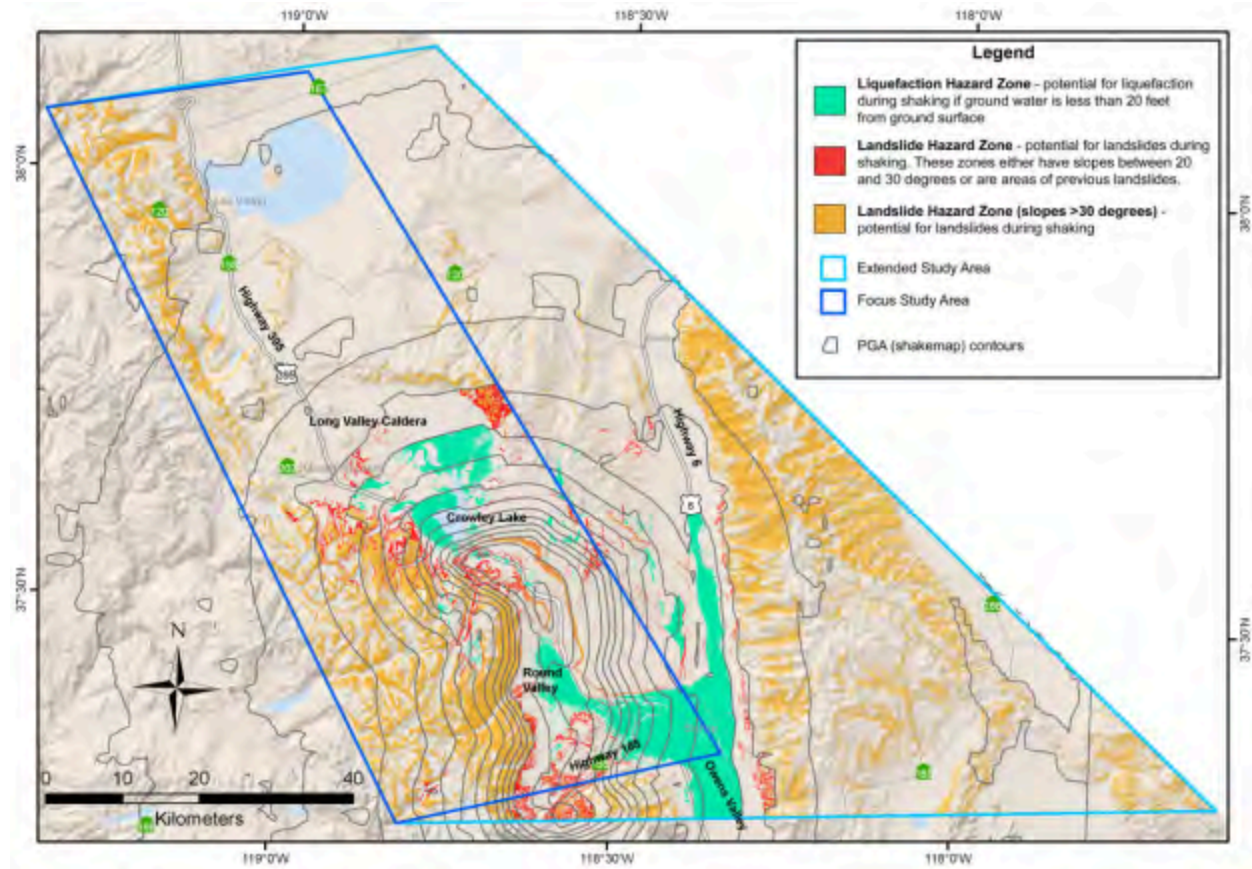


Figure S6.10. Map showing potential liquefaction and landslide areas for the Round Valley M7.0 scenario. All areas with slopes greater than 30 degrees are considered at risk during even minimum ground shaking. These areas are shown as orange. Additional landslide hazard zones due to combined effects of scenario ground motion, slope, and shear strength of the geologic materials are shown as red.

References (for this guidebook, see Chen et al, 2014 for references for that study)

- Chen, R., Branum, D.M., Wills, C.J., and Hill, D.P., 2014, Scenario earthquake hazards for the Long Valley Caldera-Mono Lake area, east-central California: U.S. Geological Survey Open-File Report 2014-1045, and California Geological Survey Special Report 233, 92 p., <http://dx.doi.org/10.3133/ofr20141045>.
- Davis, J.F., Bennett, J.H., Borchardt, G.A., Kahle, J.E., Rice, S.J., and Silva M.A., 1982, Earthquake Planning Scenario for a Magnitude 8.3 Earthquake on the San Andreas Fault in Southern California: California Department of Conservation, Division of Mines and Geology Special Publication 60, 121 p.

- Jones, L.M., Bernknopf, R., Cox, D., Goltz, J., Hudnut, K., Mileti, D., Perry, S., Ponti, D., Porter, K., Reichle, M., Seligson, H., Shoaf, K., Treiman, J., and Wein, A., 2008, The ShakeOut Scenario (Version 1.0): U.S. Geological Survey Open-File Report 2008-1150, iv, 308 p., <https://doi.org/10.3133/ofr20081150>
- Mangan, M., Ball, J., Wood, N., Jones, J.L., Peters, J., Abdollahian, N., Dinitz, L., Blankenheim, S., Fenton, J., and Pridmore, C., 2019, California's exposure to volcanic hazards (ver. 1.1, December 2019): U.S. Geological Survey Scientific Investigations Report 2018-5159, 49 p., <https://doi.org/10.3133/sir20185159>.
- Porter, K., Wein, A., Alpers, C., Baez, A., Barnard, P., Carter, J., Corsi, A., Costner, J., Cox, D., Das, T., Dettinger, M., Done, J., Eadie, C., Eymann, M., Ferris, J., Gunturi, P., Hughes, M., Jarrett, R., Johnson, L., Dam Le-Griffin, H., Mitchell, D., Morman, S., Neiman, P., Olsen, A., Perry, S., Plumlee, G., Ralph, M., Reynolds, D., Rose, A., Schaefer, K., Serakos, J., Siembieda, W., Stock, J., Strong, D., Sue Wing, I., Tang, A., Thomas, P., Topping, K., and Wills, C.; Jones, Lucile, Chief Scientist, Cox, Dale, Project Manager, 2011, Overview of the ARkStorm scenario: U.S. Geological Survey Open-File Report 2010-1312, 183 p. and appendixes.

Normal Fault displacement along the eastern Sierra Nevada frontal fault system

Contributor: Ken Hudnut (retired, former USGS)

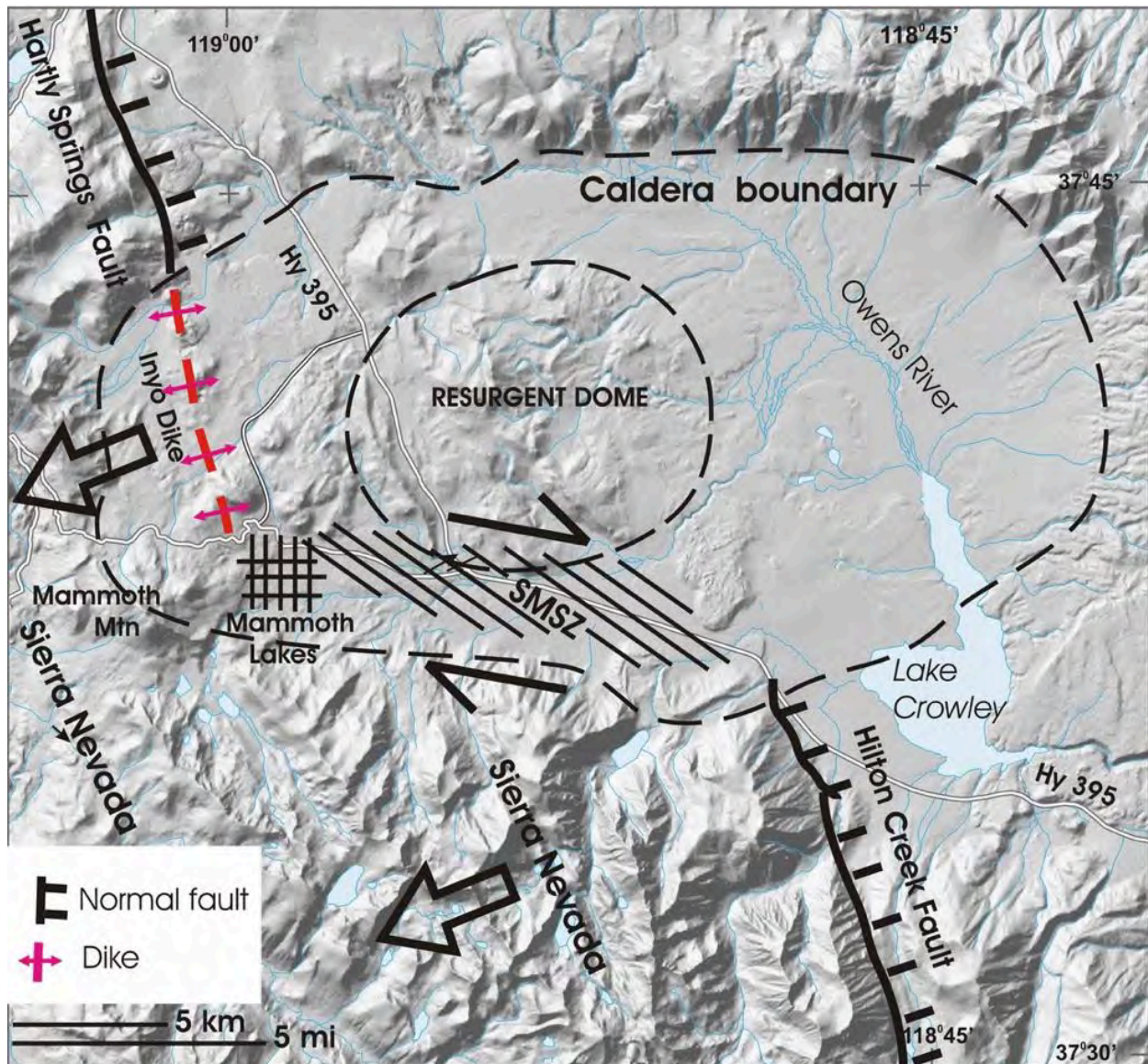


Figure S6.11. LiDAR image of the Hilton Creek and Hartly Springs faults along the eastern Sierra Nevada

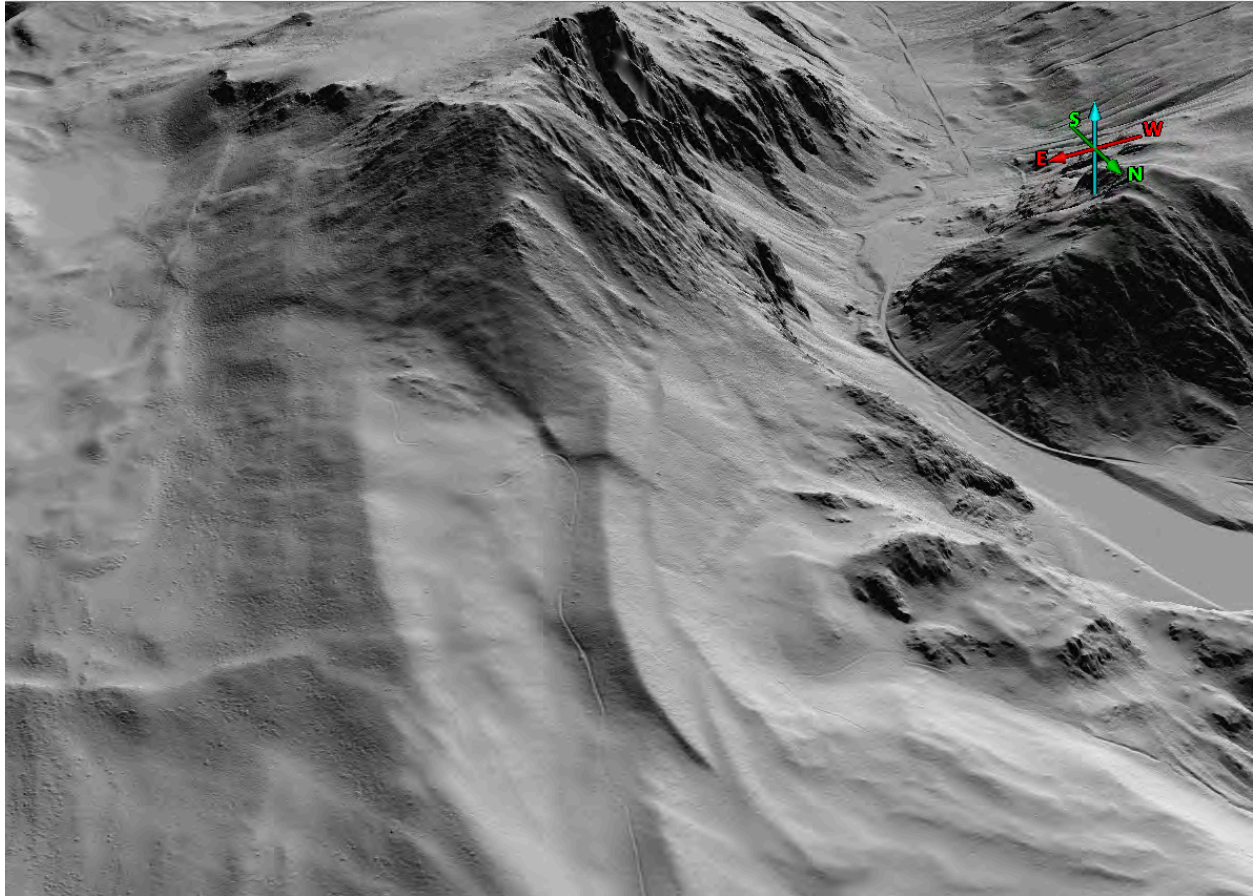


Figure S6.12. Aerial LiDAR view of the Hartly Springs Faults offsetting a moraine.

Day 2, S7: Bloody Canyon

FIELD TRIP STOP CONTRIBUTORS: Fred Phillips (New Mexico Tech)

LOGISTICS: This stop will focus on the glacial history in Bloody Canyon

WALKING DISTANCE: Optional hike around the moraines

LOCATION: Bloody Canyon (37.894927, -119.127370)

DIRECTIONS: [395 Vista Point to Bloody Canyon](#)



Day 2, S8: Tioga Pass Stop 1

FIELD TRIP STOP CONTRIBUTORS: Greg Stock (National Park Service, Yosemite National Park)

LOGISTICS: Park on the right gravel shoulder of Highway 120, about 1/2 mile north of the Tioga Pass entrance station (37.917069 lat, -119.256292 lon). Watch for traffic and cross the road carefully. Round trip walking distance is ~1/4 mile along the highway shoulder and down an embankment to a relatively flat meadow surface. The purpose of this stop is to note the U-shape of Tioga Pass and observe glacial striations and bedforms indicating the direction of ice flow near the pass.

WALKING DISTANCE: 0.4 miles

LOCATION: Tioga Pass Stop 1 (37.912317, -119.257987)

DIRECTIONS: [Bloody Canyon to Tioga Pass Stop 1](#)



Last Glacial Maximum (Tioga) ice flow across Tioga Pass: Glacial striations north Tioga Pass

Contributor: Greg Stock (National Park Service, Yosemite National Park)

Tioga Pass is somewhat unusual among Sierra passes in its east-west alignment. From this stop, the view south towards the pass shows that—contrary to the idealized precipitous mountain pass—Tioga Pass sits within a broad U-shaped valley. This morphology strongly suggests ice flow across the pass during the Last Glacial Maximum (Tioga) glaciation circa 21-18 ka. Glaciers crossed over the Sierra crest in many places, carving wide valleys and even forming perched lake basins on the crest (one striking example is Peeler Lake, 30 km northwest of Tioga Pass, which has two outlets draining to both the Pacific and the Great Basin). In most cases, glaciers flowed from the western slope of the Sierra, where ice thickness was generally greater, up and over the crest into east-side canyons. Does that relationship hold at Tioga Pass?



Figure S8.1. Glacial striations in metavolcanic bedrock above Tioga Pass indicating northwest (54°) flow of Tioga-age ice. The steep, blocky surface in the lower left likely reflects quarrying on the down-ice side of the outcrop.

The bedrock here is mapped as pelitic hornfels, part of a larger complex of metamorphosed sedimentary and volcanic rocks of the Saddlebag Lake pendant (Bateman et al., 1983). Glacial

striations on this outcrop constrain ice flow either southward (190-195°) or northward (10-15°), but tapering striations and small asymmetric bedforms support the latter, i.e., ice flow from Tioga Pass northward towards Lee Vining Canyon. This flow direction is further supported by striations and bedforms on the walls above the pass that generally indicate converging northward ice flow across the pass (Figure S8.1).

Indeed, a 1987 USGS map depicting the extent of the Tioga glaciation in Yosemite (Alpha et al., 1987) showed ice flowing northward across Tioga Pass, a northern extension of large westward-flowing ice streams originating from the flanks of Mts. Dana and Gibbs. However,

shortly after that map's publication, this interpretation was challenged by field evidence directly observable at Tioga Pass—our next stop!

References Cited

- Alpha, T.R., Wahrhaftig, C., and Huber, N.K. (1987). Oblique map showing maximum extent of 20,000-year-old (Tioga) glaciers, Yosemite National Park, central Sierra Nevada, California: U.S. Geological Survey Miscellaneous Investigations Series Map I-1885, <https://pubs.usgs.gov/imap/i1885/>
- Bateman, P.C., Kistler, R.W., Peck, D.L., and Busacca, A. (1983) Geologic map of the Tuolumne Meadows quadrangle, Yosemite National Park, California: U.S. Geological Survey Geologic Quadrangle Map GQ-1570, <https://pubs.usgs.gov/publication/gq1570>

Day 2, S9: Tioga Pass Stop 2

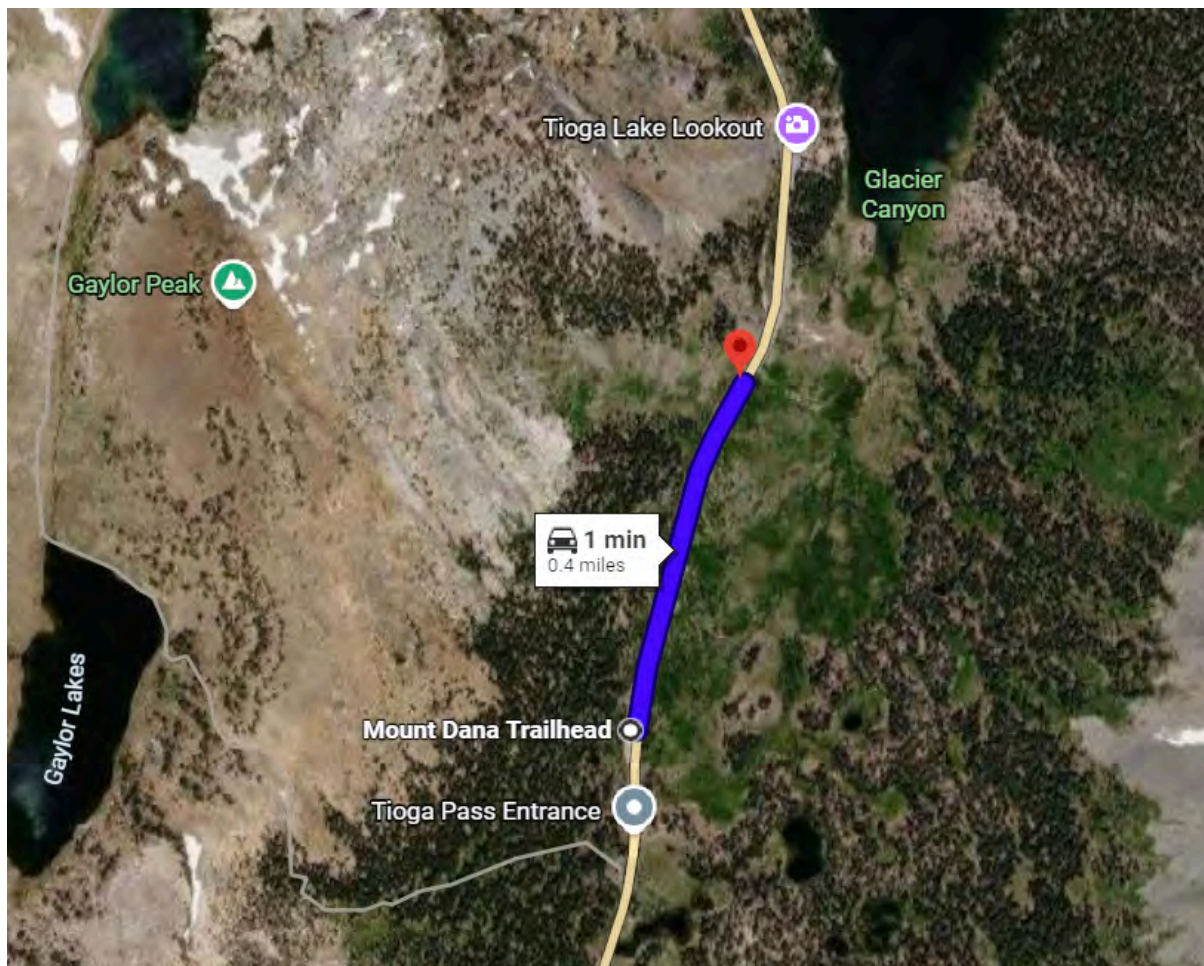
FIELD TRIP STOP CONTRIBUTORS: Greg Stock (National Park Service, Yosemite National Park), Elizabeth Haddon (US Geological Survey)

LOGISTICS: Park on the right gravel shoulder of Highway 120 (Tioga Road) just before (~0.1 mile north of) the Yosemite National Park entrance station at Tioga Pass (37.912317 lat, -119.257987 long). Be careful of traffic while walking along and crossing the highway. Round trip walking distance is ~1/2 mile, first along the highway shoulder and then on a trail across relatively level, uneven ground. The goals of this stop are to observe additional evidence for glacier flow direction(s) across Tioga Pass and to discuss deglaciation and surficial geology of the Tuolumne Meadows region.

WALKING DISTANCE: Minimal.

LOCATION: Tioga Pass Stop 2 (37.917664, -119.255884)

DIRECTIONS: [Tioga Pass Stop 1 to Tioga Pass Stop 2](#)



Rates and patterns of LGM (Tioga) deglaciation at Tuolumne Meadows and Tioga Pass

A. Exotic erratics at Tioga Pass

Contributors: Field Stop Leader: Greg Stock (National Park Service, Yosemite National Park), with contributions from Alexander Laub and Peter Clark (Oregon State University) and John Ljung (retired geologist)

Exotic granitic boulders at Tioga Pass

Tioga Pass, elevation 3,031 m (9,945 ft), is the highest highway pass in California and sits on the drainage divide between the Tuolumne River watershed and the Mono Basin. The broad, U-shaped valley forming the pass is mantled with glacial till and hummocky moraines composed of a mix of granitic, metavolcanic, and metasedimentary clasts.

As described in Huber (2007), shortly after publication of the LGM glacier map of Alpha et al. (1987) showing northward ice flow across Tioga Pass, co-author King Huber of the USGS was



asked by a local Yosemite naturalist about the many boulders of Cathedral Peak Granite there (Figure S9.1); if the glacier flowed northward across the pass, how did those boulders get there?

Figure S9.1. Boulder of Cathedral Peak Granite (CPG) at Tioga Pass, showing distinctive potassium feldspar “megacrysts”.

The Cathedral Peak Granite (CPG) is the most easily identifiable granitic rock in Yosemite, characterized by large, pink-colored orthoclase “megacrysts”. The eastern boundary of CPG bedrock is well west of Tioga Pass (Figure. S9.2). Given the prominent medial moraines extending west towards Tuolumne Meadows from the flanks of Mts. Dana and Gibbs, bearing a large component of metamorphic clasts derived from those peaks (Wahrhaftig et al., 2019), it is difficult to conceive of a scenario in which the CPG boulders could have arrived at Tioga Pass by northward flow as depicted in Alpha et al. (1987).

We mapped >3,000 CPG erratic boulders between Tioga Lake to the north and the downstream end of Dana Meadows to the southwest (Figure S9.2). The mapping reveals relatively high

concentrations of CPG boulders at Tioga Pass and to the north, with numbers decreasing southward through Dana Meadows and ultimately disappearing about 4.5 km southwest of Tioga Pass, more than 1 km upslope of the CPG contact near Tuolumne Meadows (Figure S9.2). CPG boulders are not present in the Gaylor and Granite lakes basins directly west of Tioga Pass. Because the eastern CPG contact extends north of Dana Meadows and crosses over the Sierra crest, a large area of CPG bedrock is exposed east of the crest in the vicinity of Mt. Conness (Figure S9.2). This is the most likely source for the CPG boulders at Tioga Pass, indicating southward ice flow across the pass (Huber, 2007).

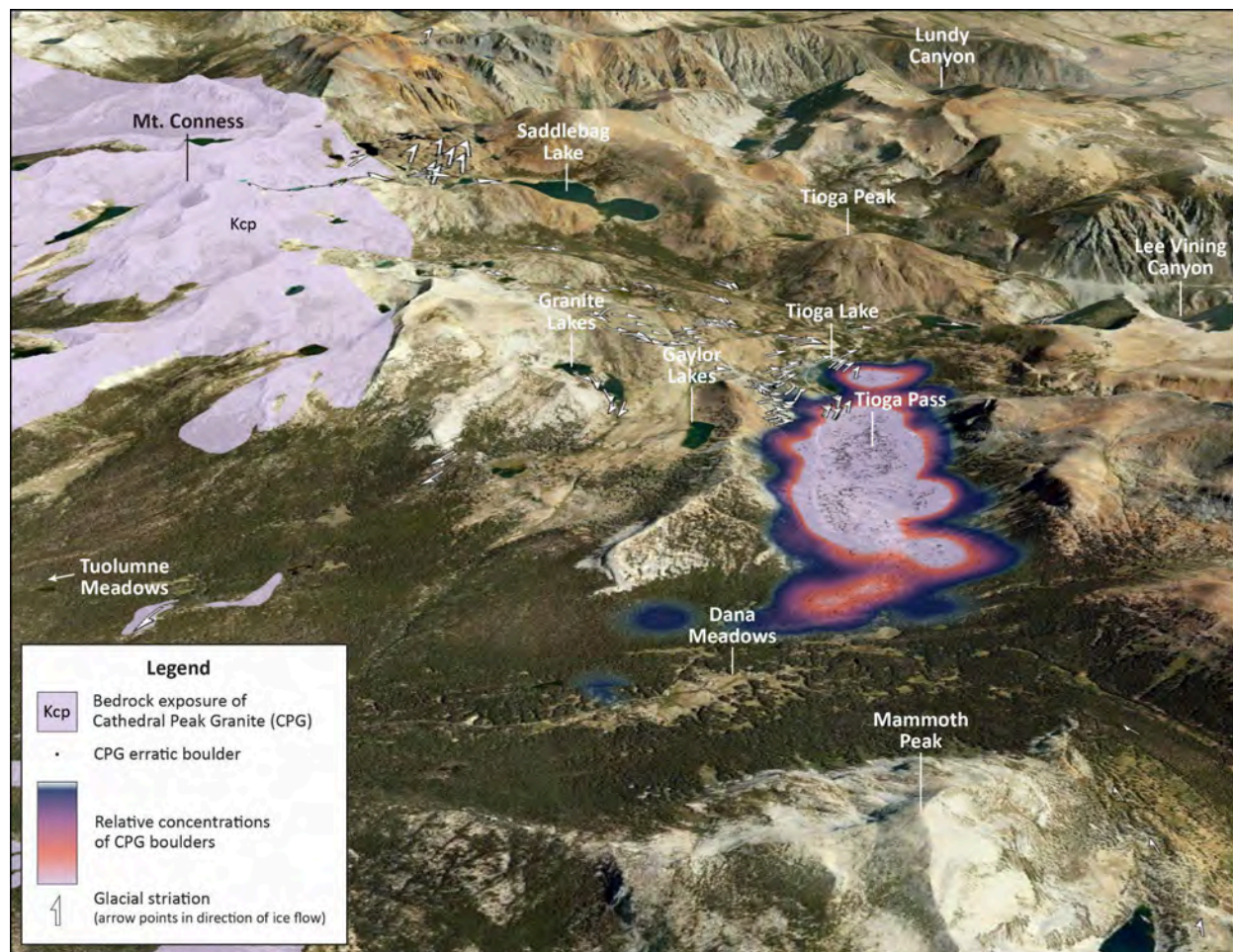


Figure S9.2. Oblique map looking north showing the spatial distribution of >3,000 CPG erratic boulders near Tioga Pass, CPG bedrock contact west-northwest of Tioga Pass, and orientation of glacial striations north of the pass.

The highest mapped CPG boulder at Tioga Pass is at 3,133 masl, indicating a minimum ice thickness of ~100 m. A presumed Tioga-age till deposit on the northwest ridge of Tioga Peak, 4.5 km north of Tioga Pass near Saddlebag Lake, is ~200 m higher than the minimum glacier

surface at Tioga Pass. Thus, the glacier surface sloped southward from the Saddlebag Lake area towards Tioga Pass during the Tioga maximum. Although most of the ice that originated from the Mt. Conness/Saddlebag Lake area flowed eastward down Lee Vining Canyon, the southward slope of the glacier surface drove an arm of ice to flow up and over Tioga Pass.

Although CPG boulders at Tioga Pass require southward ice flow across the pass, this conflicts with striations north of the pass that generally indicate northward flow (Figure S9.2). The discrepancy may be explained by shifts in ice flow direction over time. Several outcrops on the west wall of Tioga Pass preserve two sets of cross-cutting striations (Figure S9.3): a sparsely preserved set trending south-southeast (150–175°), and a more prominent set trending northeast (50–78°). Spatial relations point to the sparsely preserved south-southeast trending

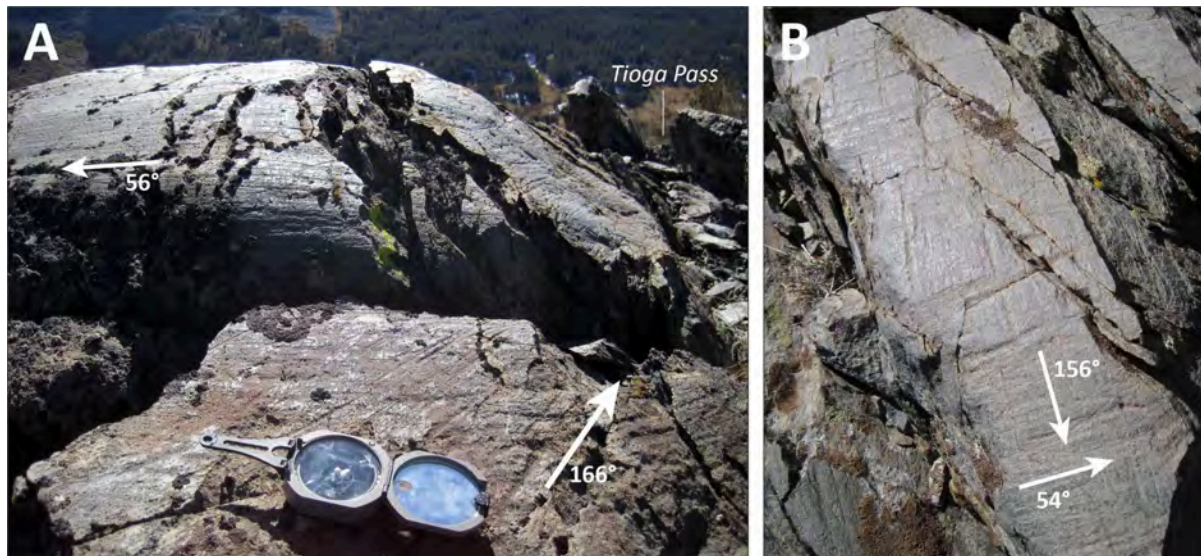


Figure S9.3 Cross-cutting glacial striations on the west wall of Tioga Pass, indicating a shift in the ice flow direction.

set as being the older of the two. This suggests a more complicated scenario in which ice flowed southward across Tioga Pass during LGM (Tioga) maximum conditions, then reversed direction on the north side as ice thinned and stagnated at the pass. The presence of hummocky heaps of glacial till and kettle ponds (Figure S9.4) indicate that, rather than retreating steadily across Tioga Pass, the glacier there thinned, stagnated, and melted in place (e.g., Benn and Evans, 2010). This occurred at 15.1 ± 0.3 ka, based on four ^{10}Be exposure ages of granitic boulders at Tioga Pass (Table S9.1), consistent with the broader story of deglaciation in the Tuolumne Meadows region (see discussions for Stops S9 B and C below).



Figure S9.4. View southeast from Gaylor Peak, showing the broad U-shape of Tioga Pass, as well as probable kettle ponds filling depressions left by the melting of stagnant ice blocks buried in till. Mt. Gibbs is the peak in the upper left.

Table S9.1. Cosmogenic ^{10}Be exposure age data for erratic boulders at Tioga Pass

Sample	Latitude	Longitude	Elevation (m)	^{10}Be concentration (atm/g qtz)	^{10}Be exposure age (kyr)
TP-10Be2	37.908162	-119.2559	3036	$472,284 \pm 12,741$	15.6 ± 0.5
TP-10Be3	37.908766	-119.2540	3038	$453,426 \pm 12,374$	15.0 ± 0.5
TP-10Be4	37.907939	-119.2544	3038	$450,276 \pm 13,197$	14.9 ± 0.5
TP-10Be5	37.907422	-119.2537	3039	$458,962 \pm 12,515$	15.1 ± 0.5

In summary, the available evidence indicates the following sequence of glacial events at Tioga Pass:

1. Glacier ice flowed from north to south over Tioga Pass during peak LGM (Tioga) conditions $\sim 20\text{-}18$ ka, transporting CPG boulders derived from the Mt. Conness area across the pass (Figure S9.5).
2. Ice thinned and stagnated at Tioga Pass ~ 15.1 ka, leaving behind CPG boulders, hummocky till deposits, and kettle ponds.

3. As ice on either side of the pass decoupled, changes in glacier surface slope redirected ice flow away from Tioga Pass in both directions, leaving northward directed striations on the north side of the pass.

At the urging of King Huber, the updated map of Yosemite's LGM glacier extent (Wahrhaftig et al., 2019) now shows ice flow southward across Tioga Pass during the Tioga maximum.

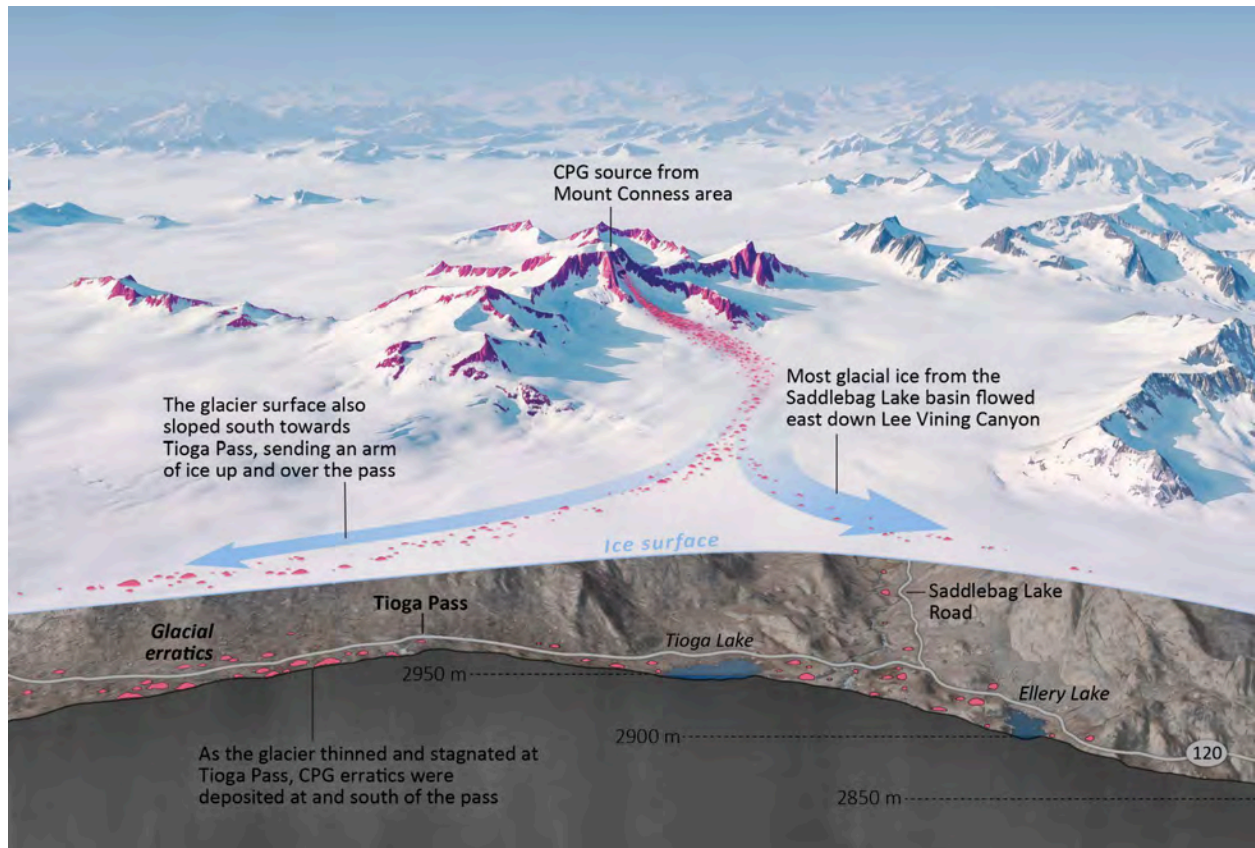


Figure S9.5. Schematic illustration looking northwest showing ice flow and transport of CPG boulders from the Mt. Conness area primarily eastward down Lee Vining Canyon but also southward across Tioga Pass. Illustration by Eric Knight.

B. Evidence for rapid deglaciation between Tuolumne Meadows and Tioga Pass

Contributors: Elizabeth Haddon (US Geological Survey) and Greg Stock (National Park Service, Yosemite National Park)

The remarkably broad subalpine basin system extending from Tuolumne Meadows to Tioga Pass was sculpted by multiple, converging Tioga-age glaciers. The distinctive arrangement of glacial landforms in this area, and their sedimentary structures, informs a fundamental question regarding local ice-retreat dynamics: did LGM (Tioga) deglaciation in the Sierra Nevada occur rapidly by in situ melting and disintegration of ice, or slowly through gradual retreat (e.g., Clark, 1976)? Reconstructing the LGM deglaciation of the Tuolumne Meadows region requires a precise understanding of geomorphic adjustments recorded by patterns of erosion and sedimentation. We conducted new surficial geologic mapping by integrating field observations, analysis of lidar topographic data, and multi-proxy geochronologic investigations (^{10}Be , IRSL, Tephra, ^{14}C) to evaluate the exceptionally well-preserved glacial landforms and investigate the pace of deglaciation.

Our mapping of the Tuolumne Meadows region, including its tributaries the Dana and Lyell forks, shows hummocky till deposits fringing and mantling polished bedrock slabs and domes, an extensive sequence of prominent medial moraines, well-preserved kettle holes and ponds, and abundant erratics. Relatively diminutive recessional moraines and short glaciofluvial terraces (10 to 100-m-scale) that slope towards the river are preserved in western parts of Tuolumne Meadows. Subsurface exposures of glacial till are limited to roadcuts between Lembert Dome and Dana Meadows, with the best cross-section view at the “Little Blue Slide,” an unstable ~110-meter-long exposure across the core of a double-crested medial moraine east of Tuolumne Meadows. Glacial erratics, partially buried by glaciofluvial and fluvial deposits but clearly anchored in glacial drift, are exposed in stream channels and floodplain surfaces.

We targeted glacial deposits for age constraint, including outwash and erratics deposited with glacial till as the ice melted, such that their surface exposure or burial ages effectively date the abandonment of that part of the glaciated landscape. We sampled erratics for ^{10}Be -exposure dating between the downstream end of Tuolumne Meadows and Tioga Pass (Figure S.9.6). The mean ^{10}Be -exposure age for 13 erratics along a 10-km-long west-east transect in Tuolumne and Dana meadows shows glacial occupation as late as 16.0 ± 0.7 ka ($n = 13$) (Figure S.9.6). This population of ^{10}Be -exposure ages lacks a clear progressive spatiotemporal pattern that would indicate gradual, systematic ice retreat across the basin. Instead, the dataset overlaps within analytical uncertainty with ^{10}Be -exposure ages from Tioga Pass, ~3.5 km to the north, which record final abandonment of the drainage divide by 15.1 ± 0.3 ka. We also sampled fine sand filling terraced glacial outwash deposits for infrared stimulated luminescence (IRSL) dating. Our

first preliminary IRSL age indicates burial at 16.6 ± 1.1 ka for a basal sediment sample in a ~1-m-thick sequence of laminated fine-grained sediment deposited in a depression on glacial drift north of the Dana Fork. Taken together, these similar results imply relatively synchronous, rapid deglaciation.

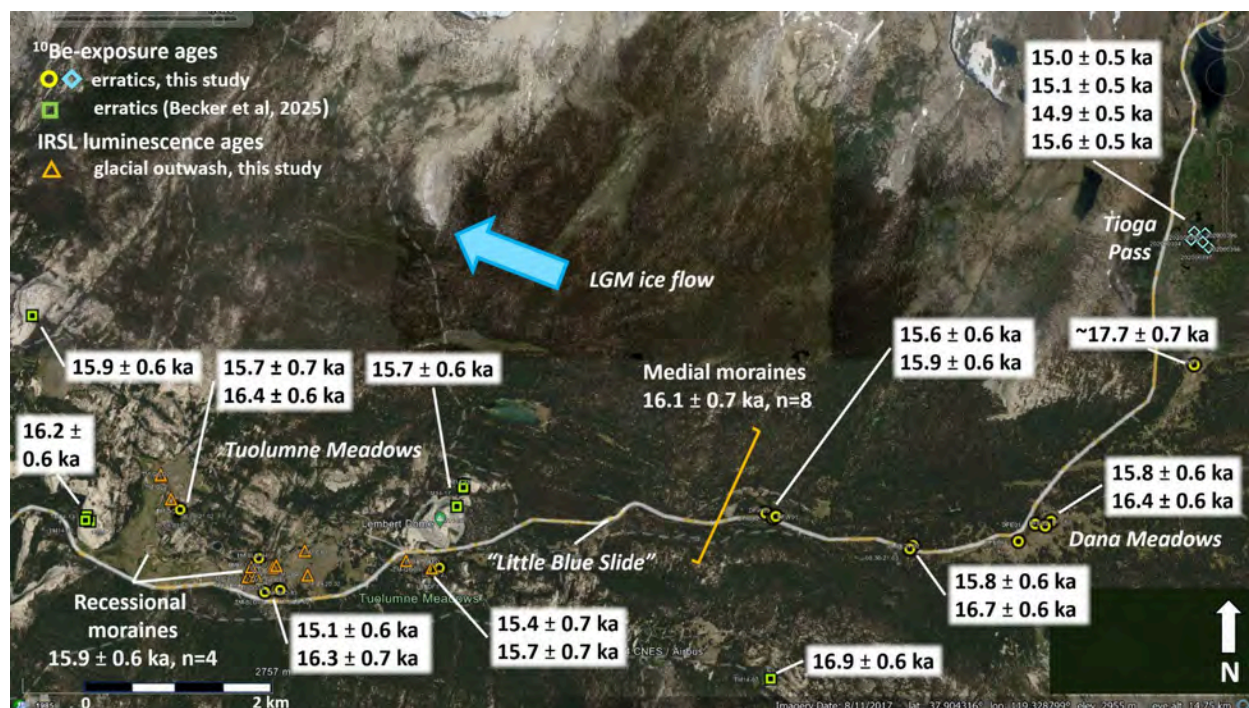


Figure S9.6. Spatial distribution of ^{10}Be cosmogenic exposure ages. Sample sites, located between Tuolumne Meadows and Dana Meadows (yellow circles) and Tioga Pass (blue diamonds), shown relative to the Tioga Road. Basemap is Google Earth satellite imagery collected in 2017. Green squares show previously published ^{10}Be -exposure ages for erratics reported by Becker et al. (2025). Weighted mean ^{10}Be -exposure age for the 13 erratics (yellow circles) along the 10-km-long transect between Tuolumne and Dana Meadows is 16.0 ± 0.7 ka. Also shown are inferred ages for recessional (15.9 ± 0.6 ka, $n=4$) and medial (16.1 ± 0.7 ka, $n=8$) moraines, approximated using ^{10}Be -exposure ages for erratics deposited on either side of the crested moraines.

In Clark's (1976) conceptual framework, the linear geometry of the medial moraines records sediment concentration due to active convergence of ice flows from the flanks of Mts. Dana, Gibbs, and the Kuna Crest (Wahrhaftig et al., 2019); however, their preservation as continuous, crested landforms indicates passive downwasting of ice rather than retreat (e.g., Benn and Evans, 2010). The "Little Blue Slide" roadcut through one of these medial moraines provides a valuable window into its sedimentary structure, consisting of massive, uncompacted, matrix-supported diamict. The large volume of fine-grained sand and rock flour in this moraine reflects local bedrock lithologies predisposed to abrasive glacial erosion (e.g., Duhnforth et al.,

2010). Slow systematic ice retreat at this location would likely have at least partially eroded and dissected this matrix-rich till via relatively high-energy meltwater, channelized along ice-debris contacts (e.g., Benn and Evans, 2010; Marren and Toomath, 2013).

The geomorphic assemblage from western Tuolumne Meadows to Tioga Pass provides macro-scale evidence of adjacent ice-contact depositional environments, consistent with a transition from active ice flow to quasi-retreat of the ice-margin followed by widespread ice stagnation and disintegration. The spatial distribution of landforms in western Tuolumne Meadows downslope of the recessional moraines—including a depositional drumlin of well-sorted clasts, onlapping outwash terraces, and a sinuous esker—indicate a period of active ice flow and a confined, subglacial hydraulic system consistent with relatively thick glacial ice (e.g., Shreve, 1985; Shaw and Sharpe, 1987). Low recessional moraines in Tuolumne Meadows record gradual thinning and retreat punctuated by periods of ice stabilization (e.g., Benn and Evans, 2010). Conversely, extensive hummocky topography beginning in eastern Tuolumne Meadows and the increasing extent of kettle hole basins farther upslope require successive ice thinning and the cessation of ice flow (e.g., Ives and Andrews, 1963; Clayton, 1967; Andersson, 1998), triggering the collapse of debris-laden ice and its inferred conduit system (e.g., Eyles, 1983). As large, isolated blocks of sediment-buried ice slowly melted (e.g., Schomacker, 2008), supraglacial and englacial debris likely slumped, slid, and chaotically piled into clusters of mounds separated by closed depressions (e.g., Benn, 1992; Eyles et al., 1999). If the LGM ice had maintained dynamic retreat then, given the gentle gradient in surface slope, high-energy ice-marginal meltwater streams presumably would have progressively eroded the hummocks and infilled the adjacent kettle holes with stratified outwash gravels (e.g., Maizels, 1983; Marren, 2005).

The lack of recessional chronologies for glacial deposits across a 10-km transect, combined with the near pristine preservation of diagnostic landform assemblages, strongly support a rapid transition to widespread ice stagnation and in situ disintegration at 16.0 ± 1.3 ka. Applying a similar multi-proxy chronostratigraphic approach to other settings in the Sierra for comparison with our calibrated baseline will allow geologists to more reliably reconstruct LGM deglaciation in the Sierra Nevada.

C. Rapid deglaciation of the Tuolumne Meadows region linked to glacier hypsometry

Contributors: Andrew Jones, Marianne Haseloff, and Shaun Marcott (University of Wisconsin-Madison), Elizabeth Haddon (US Geological Survey), and Greg Stock (National Park Service, Yosemite National Park)

During the LGM deglaciation (19–11 ka), episodes of glacier retreat and stabilization in the western United States are often attributed to abrupt climate change events via changes in Atlantic Meridional Overturning Circulation (e.g., Becker et al. 2025; McGee et al. 2018; Phillips 2016). However, there is considerable variability in the magnitude and timing of glacier retreat across the western US that is difficult to reconcile with a uniform climate forcing (Marcott et al. 2019), indicating either heterogeneous climate change or the influence of local non-climatic variables, such as glacier hypsometry (Young et al. 2011). Prior work from basal lake radiocarbon ages (Gillespie and Clark 2011) and cosmogenic nuclide dating (Phillips 2016; Phillips et al. 2009; Rood et al. 2011) established that Sierra Nevada glaciers receded from LGM moraines to north-facing cirques between 18 and 15 ka, with most deglaciation occurring at 15.8 ± 0.5 ka. As discussed above, geomorphic evidence supports abrupt glacier stagnation and in situ melting in the Tuolumne Meadows region.

We investigated whether the abrupt deglaciation of the Tuolumne Meadows region was primarily driven by climate or topography using a one-dimensional glacier flowline mass-balance model. We conducted a series of temperature and precipitation step-change experiments over two bed geometries: the current bed profile of the Tuolumne glacier derived from a 10-m DEM, and a generalized linear bed profile. In each test, the model is spun up to match the known glacier size at LGM, which is based on the 20.0 ± 2.0 ka lateral moraine at Camp Mather ~100 km from Mt. Lyell, and trimlines that limit ice elevation to ~3800 m (Wahrhaftig et al. 2019).

In the temperature step-change experiments, temperature is raised by 2, 4, 6, 8, and 10° C in five separate simulations over the two profiles and held for 3 kyr (Figure S9.7). The same temperature change produces two different length responses depending on the bed profile. Over the linear bed, each increasing step change produces a proportionate response. For example, 4 °C of warming causes twice as much retreat as 2 °C of warming (Figure S9.7). Over the actual bed profile, the relationship is non-linear. Warming of 2 °C causes a ~40% reduction in length as the glacier retreats over a low-gradient portion of the bed. Warming of 4 and 6 °C each cause only an additional retreat of ~10% as the glacier recedes over steeper terrain. Following 8 °C of warming, the glacier recedes over a second low-gradient section and loses an

additional 30% in length. Additionally, the +8° C warming scenario ends with a small remnant cirque glacier over the actual bed profile, but not over the linear bed profile.

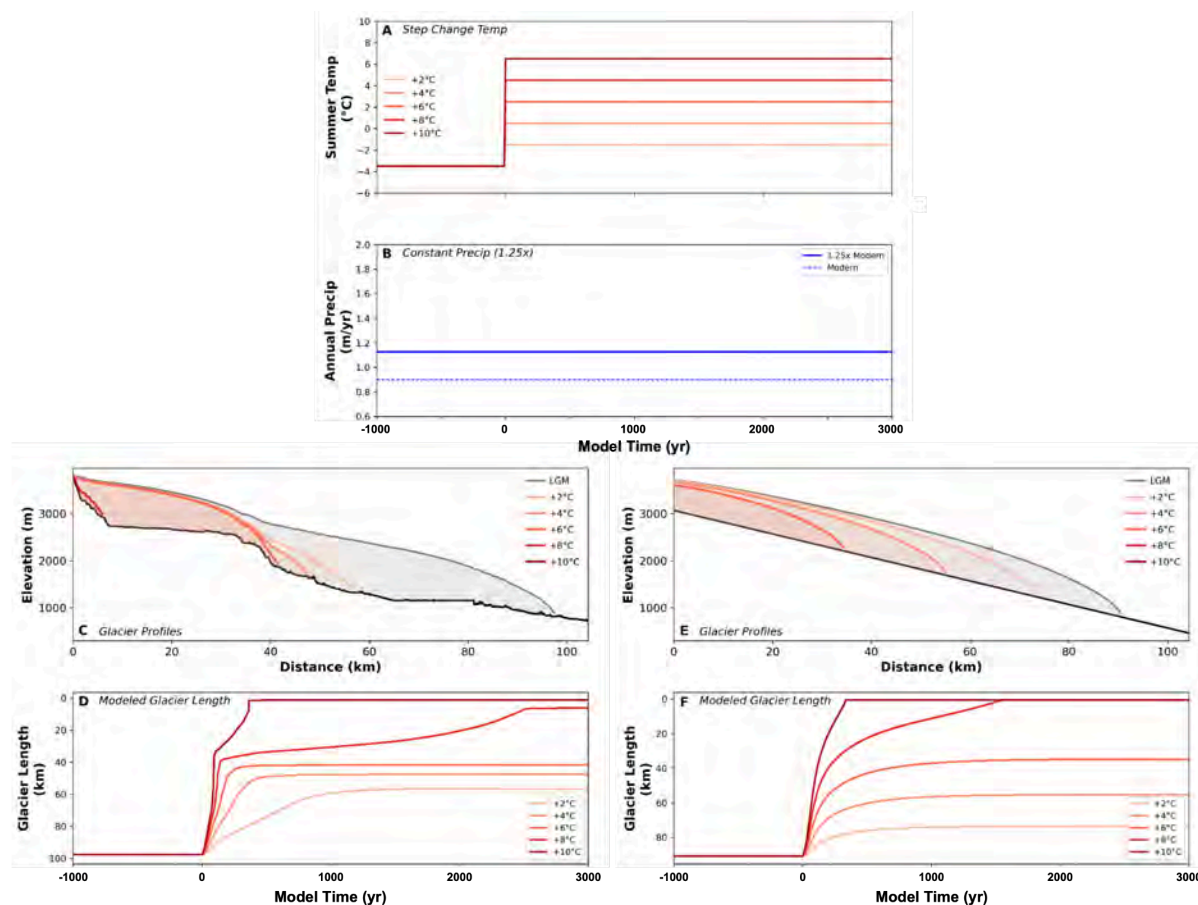


Figure S9.7. Temperature step-change experiments. Glacier flowline model simulations over the current Tuolumne River canyon profile and a linear bed. A. Temperature is increased in increments of 2° C from an initial temperature calibrated to LGM glacier size. B. Precipitation is held at 1.25x modern values (modern precipitation is shown for reference). C, E. Glacier profiles 3 ky after the step-change were imposed. D, F. Modeled glacier length. Time is in model years, where years -1000 to 0 represent the last 1,000 years of model spin-up to LGM size, and year 0 is when the step-change is imposed.

The precipitation step-change experiments vary precipitation while holding temperature constant (Figure S9.8). The initial precipitation for these experiments is 250% of modern, based on the maximum estimate of Barth et al. (2016). Precipitation is decreased in steps from 250% to 200%, 150%, 100%, and 50% of modern values. The precipitation step-change experiments show the same pattern, where the glacier is unstable over the flat part of the bed and resilient to change over steeper areas.

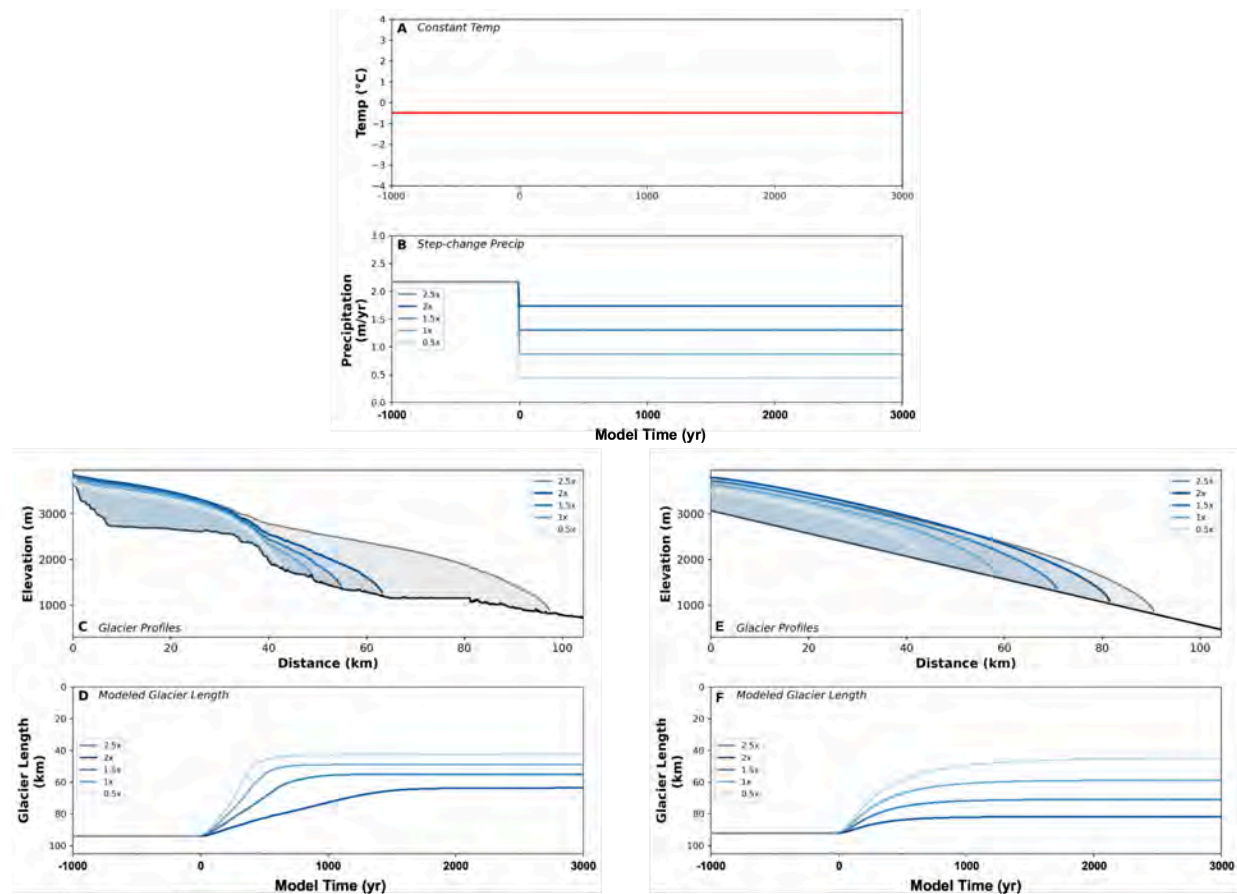


Figure S9.8. Precipitation step-change experiments. Glacier flowline model simulations over the current Tuolumne River canyon profile and a linear bed. A. Temperature scenario (constant). B. Precipitation is decreased from 2.5x modern precipitation levels to 0.5x modern levels. C, E. Glacier profiles 3 ky after the step-change was imposed. D, F. Modeled glacier length. Time is in model years, where years -1000 to 0 represent the last 1,000 years of model spin-up to LGM size, and year 0 is when the step-change is imposed.

Our experiments indicate that bed geometry is an important component of rapid deglaciation. When the glacier recedes to a low-slope or plateau-like area of the bed, it disintegrates rapidly and disproportionately to warming. This geometry-dependent sensitivity offers a mechanistic explanation for the heterogeneous deglacial record of the western US—adjacent valleys subjected to similar climate forcing but underlain by different bed profiles would be expected to deglacierate at markedly different rates (Young et al. 2011). As the glacier's Equilibrium Line Altitude rises above a plateau, ice thins and the glacier surface descends into warmer air, amplifying melt through ice-elevation feedbacks that can sustain rapid mass loss without

additional external forcing (Leonard et al. 2023). Rapid deglaciation of the Tuolumne Meadows region at ~16 ka may therefore reflect the internal instability of a large, low-gradient ice body responding non-linearly to gradual warming, rather than being the product of abrupt climate change.

References Cited

- Alpha, T.R., Wahrhaftig, C., and Huber, N.K. (1987). Oblique map showing maximum extent of 20,000-year-old (Tioga) glaciers, Yosemite National Park, central Sierra Nevada, California: U.S. Geological Survey Miscellaneous Investigations Series Map I-1885, <https://pubs.usgs.gov/imap/i1885/>
- Andersson, G. (1998). Genesis of hummocky moraine in the Bolmen area, southwestern Sweden. *Boreas*, 27(1), 55-67. <https://doi.org/10.1111%2Fj.1502-3885.1998.tb00867.x>.
- Barth, C., Boyle, D.P., Hatchett, B.J., Bassett, S.D., Garner, C.B., and Adams, K.D. (2016). Late Pleistocene climate Inferences from a water balance model of Jakes Valley, Nevada (USA). *Journal of Paleolimnology* 56 (2): 109–22. <https://doi.org/10.1007/s10933-016-9897-z>
- Becker, R.A., Barth, A.M., Marcott, S.A., Tikoff, B., and Caffee, M.W. (2025). Deglaciation of the Sierra Nevada (USA) during Heinrich Event 1. *EGUsphere*, <https://doi.org/10.5194/egusphere-2025-1370>
- Benn, D. I. (1992). The genesis and significance of ‘hummocky moraine’: evidence from the Isle of Skye, Scotland. *Quaternary Science Reviews*, 11(7-8), 781-799. [https://doi.org/10.1016/0277-3791\(92\)90083-K](https://doi.org/10.1016/0277-3791(92)90083-K).
- Benn, D. I., and Evans, D. J. A. (2010). *Glaciers and Glaciation* (2nd ed.). Routledge.
- Clark, M.M. (1976). Geomorphic significance of stagnant glaciers in the Sierra Nevada, California. *Geological Society of America Abstracts with Programs*, 8(6), 813–814.
- Clayton, L. (1967). Stagnant-glacier features of the Missouri Coteau in North Dakota. *Glacial Geology of the Missouri Coteau: North Dakota Geological Survey Miscellaneous Series*, 30, 1-46.
- Dühnforth, M., Anderson, R.S., Ward, D. J., and Stock, G.M. (2010). Bedrock fracture control of glacial erosion processes and rates. *Geology*, 38(5), 423–426. <https://doi.org/10.1130/G30576.1>
- Eyles, N. (1983). Modern Icelandic glaciers as depositional models for ‘hummocky moraine’ in the Scottish Highlands. In E. B. Evenson, C. Schlüchter, & J. Rabassa (Eds.), *Tills and Related Deposits* (pp. 47–60). A. A. Balkema. <https://pascal-francis.inist.fr/vibad/index.php?action=getRecordDetail&idt=9576919>
- Eyles, N., Boyce, J. I., & Barendregt, R. W. (1999). Hummocky moraine: sedimentary record of stagnant Laurentide Ice Sheet lobes resting on soft beds. *Sedimentary Geology*, 123(3-4), 163-174. [https://doi.org/10.1016/S0037-0738\(98\)00129-8](https://doi.org/10.1016/S0037-0738(98)00129-8).

- Gillespie, A.R., and Clark, D.H. (2011). Glaciations of the Sierra Nevada, California, USA, In Developments in Quaternary Sciences, edited by Ehlers, J., Gibbard, P.L., and Hughes, P.D., v. 15. Quaternary Glaciations - Extent and Chronology. Elsevier. <https://doi.org/10.1016/B978-0-444-53447-7.00034-9>
- Huber, N.K. (2007). Exotic boulders at Tioga Pass, in *Geological Ramblings in Yosemite: Heyday* Books, Berkeley, California, 121 p.
- Ives, J. D. and Andrews, J. T. (1963). Studies in the physical geography of north-central Baffin Island, N.W.T. Canada Department of Mines and Technical Surveys, Geographical Branch, Geographical Bulletin, 19, 5-48. <https://doi.org/10.4095/331088>.
- Leonard, E.M., Laabs, B.J.C., Marcott, S.A., et al. (2023). Chronology and climate of the Last Glacial Maximum and the subsequent deglaciation in the northern Medicine Bow Mountains, Wyoming, USA. *Quaternary Science Advances* 12, 100109. <https://doi.org/10.1016/j.qsa.2023.100109>
- Maizels, J. K. (1983). Proglacial channel systems: Change and thresholds for change over long, intermediate and short time-scales. *Modern and ancient fluvial systems*, 251-266. <https://doi.org/10.1002/9781444303773.ch20>.
- Marcott, S.A., Clark, P.U., Shakun, J.D., Brook, E.J., Davis, P.T., and Caffee, M.W. (2019). ^{10}Be age constraints on latest Pleistocene and Holocene cirque glaciation across the western United States. *Npj Climate and Atmospheric Science* 2 (1): 1–7. <https://doi.org/10.1038/s41612-019-0062-z>
- Marren, P. M. (2005). Magnitude and frequency in proglacial rivers: a geomorphological and sedimentological perspective. *Earth-Science Reviews*, 70(3-4), 203-251. <https://doi.org/10.1016/j.earscirev.2004.12.002>.
- Marren, P. M., & Toomath, S. C. (2013). Fluvial adjustments in response to glacier retreat: Skaftafellsjökull, Iceland. *Boreas*, 42(1), 57-70. <https://doi.org/10.1111/j.1502-3885.2012.00275.x>
- McGee, D., Moreno-Chamarro, E., Marshall, J., and Galbraith, E.D. 2018. Western US lake expansions during Heinrich stadials linked to Pacific Hadley circulation. *Science Advances* 4 (11), <https://doi.org/10.1126/sciadv.aav0118>
- Phillips, F.M. (2016). Cosmogenic Nuclide data sets from the Sierra Nevada, California, for assessment of nuclide production models: I. Late Pleistocene glacial chronology. *Quaternary Geochronology* 35: 119–29. <https://doi.org/10.1016/j.quageo.2015.12.003>
- Phillips, F.M., Zreda, M., Plummer, M.A., Elmore, D., and Clark, D.H. (2009). Glacial geology and chronology of Bishop Creek and vicinity, eastern Sierra Nevada, California. *GSA Bulletin* 121 (7–8): 1013–33, <https://doi.org/10.1130/B26271.1>
- Rood, D.H., Burbank, D.W., and Finkel, R.C. (2011). Chronology of glaciations in the Sierra Nevada, California, from ^{10}Be surface exposure dating. *Quaternary Science Reviews* 30 (5): 646–61, <https://doi.org/10.1016/j.quascirev.2010.12.001>

- Schomacker, A. (2008). What controls dead-ice melting under different climate conditions? A discussion. *Earth-Science Reviews*, 90(3-4), 103-113. <https://doi.org/10.1016/j.earscirev.2008.08.003>.
- Shaw, J. and Sharpe, D.R. (1987) Drumlin formation by subglacial meltwater erosion. *Canadian Journal of Earth Sciences*, 24, 2316-2322.
- Shreve, R. L. (1985). Esker characteristics in terms of glacier physics, Katahdin esker system, Maine. *Geological Society of America Bulletin*, 96(5), 639-646. [https://doi.org/10.1130/0016-7606\(1985\)96<639:ECITOG>2.0.CO;2](https://doi.org/10.1130/0016-7606(1985)96<639:ECITOG>2.0.CO;2)
- Small, R.J. (1983). The lateral and medial moraines of valley glaciers: a contextual framework for debris transport. *Glacial Geology: An Introduction for Engineers and Earth Scientists*, 105-124.
- Wahrhaftig, C., Stock, G.M., McCracken, R.G., Sasnett, P, and Cyr, A.J. (2019). Extent of the Last Glacial Maximum (Tioga) glaciation in Yosemite National Park and vicinity, California, USGS Scientific Investigations Map 3414, <https://doi.org/10.3133/sim3414>
- Young, N.E., Briner, J.P., Leonard, E.M., Licciardi, J.M., and Lee, K. (2011). Assessing climatic and nonclimatic forcing of Pinedale glaciation and deglaciation in the western United States, *Geology* 39 (2): 171–74. <https://doi.org/10.1130/G31527.1>

Day 2, S10: Lee Vining Canyon

FIELD TRIP STOP CONTRIBUTORS: William Hammond (Nevada Geodetic Lab., Nevada Bureau of Mines and Geology, University of Nevada, Reno), Fred Phillips (New Mexico Tech), Greg Stock (National Park Service, Yosemite National Park)

LOGISTICS: Park on the large gravel turnout on the right side of Highway 120 (Tioga Road) ~1/2 mile beyond the spillway of Ellery Lake (37.940434 lat, -119.230002 lon). No walking required for this stop! The goals of this stop are to observe fractured granitic bedrock in Lee Vining Canyon and discuss possible geomorphic implications.

WALKING DISTANCE: 0.2 miles

LOCATION: Tioga Pass Stop 2 (37.940434, -119.230002)

DIRECTIONS: [Tioga Pass Stop 2 to Lee Vining Canyon](#)



Contemporary Uplift of The Sierra Nevada from GNSS Networks

Contributors: William C. Hammond, Nevada Geodetic Lab., Nevada Bureau of Mines and Geology, University of Nevada, Reno

Geodetic evidence suggests that the Sierra Nevada moves upward with respect to the Great Basin and the Earth center of mass. Data from GNSS networks in operation since 1994 reveal the distribution of uplift and subsidence signals from tectonic, earthquake cycle, magmatic, hydrological surface loading and poroelastic aquifer deformation.

Figure S10.1 shows a map of the vertical motions that have been estimated at GNSS stations using robust methods including the MIDAS non-parametric trend estimator (Blewitt et al., 2016) and GPS Imaging (Hammond et al., 2016). The patterns reveal vertical motions near 1 mm/yr in the Southern Sierra Nevada, as well as rapid subsidence in the California Central Valley/San Joaquin Valley. Uplift in central Nevada near the ruptures of the 1915 Pleasant Valley, 1954 Dixie Valley, Fairview Peak and other earthquakes are attributable to postseismic relaxation of the mantle (Gourmelen and Amelung, 2005; Hammond et al., 2009).

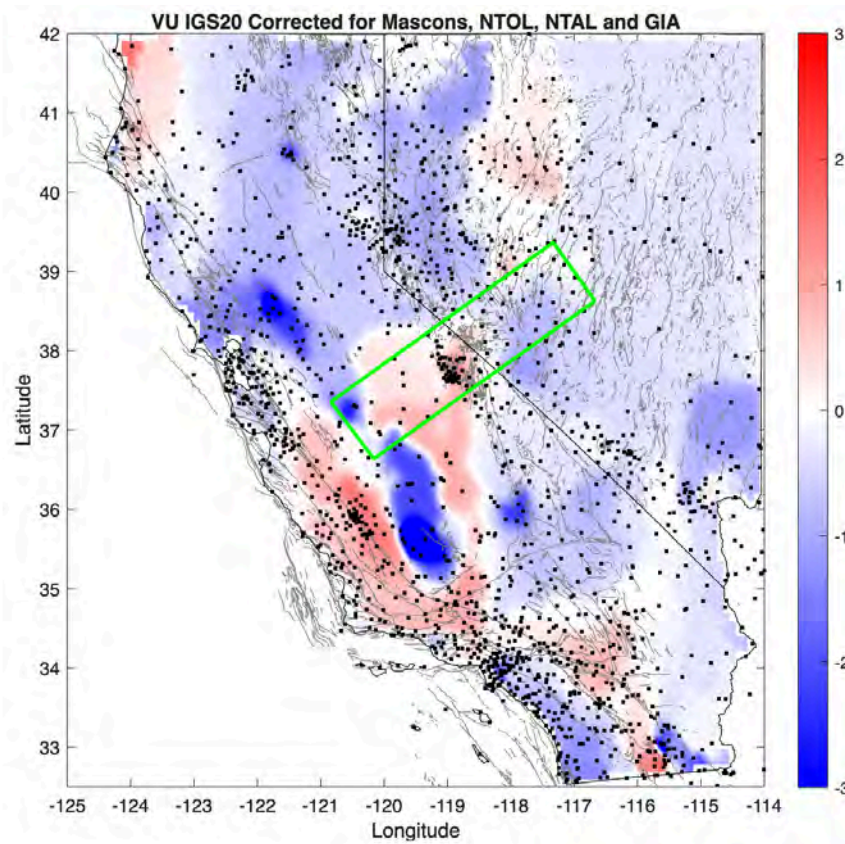


Figure S10.1. GPS Imaging of vertical GNSS velocities in California and Nevada. Sierra Nevada uplift (red=up, blue=down, color scale in mm/yr). Figure is similar to that in Hammond et al., (2016) except all recent data have been included, and a suite of surface loading corrections have been applied. Black dots indicate GNSS stations. Green rectangle indicates location of profile shown in Figure S10.2. The color scale is clipped at ± 3 mm/yr to emphasize the spatial pattern of vertical land motion.

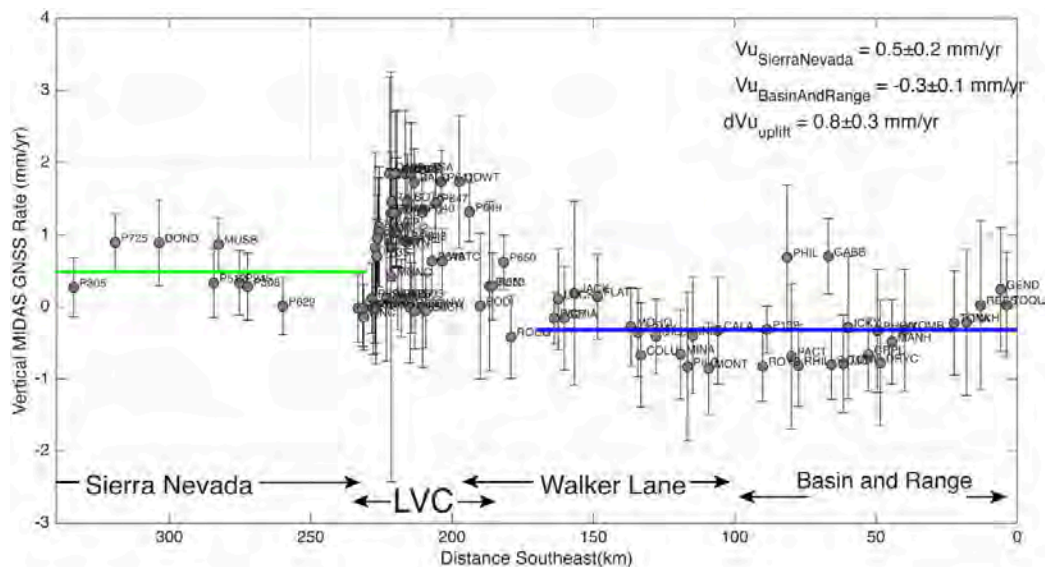


Figure S10.2. Profile of GNSS vertical velocity. Vertical bars indicate uncertainties. Annotation at bottom locates geologic domains. Blue horizontal bar indicates the mean vertical rate of stations in Basin and Range/Walker Lane. Green bar is the mean vertical rate of stations in the local Sierra Nevada. LVC is Long Valley Caldera showing high uplift rates owing to magmatic inflation. Annotation at top right indicates rates and uncertainties of site groups. Difference between Sierra Nevada and Walker Lane/Basin and Range vertical rate is 0.8 ± 0.3 mm/yr.

Long term uplift of the Sierra Nevada is thought to occur from one or more of various physical mechanisms (Henry, 2009), but the GNSS-imagined uplift could arise from more transient and recent factors. These include short-term hydrological unloading from Central Valley water withdrawal (Amos et al., 2014), drying of the range itself owing to drought cycles (Hammond et al., 2016; Argus et al., 2017). If the geodetic uplift signal is to be attributed to tectonic uplift associated with e.g., with Basin and Range / Walker Lane extension (Hammond et al., 2012), or removal of a dense, eclogitic root beneath the Sierra Nevada batholith which subsequently unloaded the lithosphere (Ducea and Saleeby, 1996; Jones et al., 2004; Zandt et al. 2004), the transient explanations must be addressed and explained explicitly.

Over the last 10 years of additional data and new stations have collected more GNSS data. Since observation began at least five California drought cycles have now come and passed. Recently the Nevada Geodetic Laboratory has incorporated surface loading model corrections for non-tidal atmospheric loading (NTAL), non-tidal ocean loading (NTOL), and GRACE-mascon-based equivalent water height loads. Yet when all these new data and corrections are applied to the processed GNSS time series and the data are imaged as before,

the pattern of uplift/subsidence (Figure 1) remains very similar to the previous published result, suggesting that the Sierra Nevada uplift is not a product of surface loading from any single drought cycle.

Looking at profiles of the vertical GNSS velocities in detail (Figure 2) gives a clearer impression of the signals in the data compared to their uncertainties and their consistency prior to the imaging. Stations considered individually do not necessarily move significantly differently than zero in the International Terrestrial Reference Frame, but taken together their combined uncertainties are smaller, and resolve significant differences in rates across the Sierra Nevada eastern margin.

References

- Amos, C. B., Audet, P., Hammond, W. C., Burgmann, R., Johanson, I. A., & Blewitt, G. (2014). Uplift and seismicity driven by groundwater depletion in central California. *Nature*, 509(7501), 483–486. <https://doi.org/10.1038/nature13275>
- Argus, D. F., Landerer, F. W., Wiese, D. N., Martens, H. R., Fu, Y., Famiglietti, J. S., Watkins, M. M. (2017). Sustained water loss in California's mountain ranges during severe drought from 2012 to 2015 inferred from GPS. *Journal of Geophysical Research: Solid Earth*, 122, 10,559–10,585. <https://doi.org/10.1002/2017JB014424>
- Blewitt, G., C. Kreemer, W.C. Hammond, J. Gazeaux, (2016). MIDAS Robust Trend Estimator for Accurate GPS Station Velocities Without Step Detection, *Journal of Geophysical Research Solid Earth*, 121, <https://doi.org/10.1002/2015JB012552>.
- Ducea M.N. Saleeby J.B. (1996). Buoyancy sources for a large, unrooted mountain range, the Sierra Nevada, California: Evidence from xenolith thermobarometry: *Journal of Geophysical Research*, v. 101, p. 8229–8244, <https://doi.org/10.1029/95JB03452>.
- Gourmelen, N., and F. Amelung (2005), Postseismic mantle relaxation in the Central Nevada Seismic Belt, *Science*, 310, 1473–1476.
- Hammond, W.C., C. Kreemer, and G. Blewitt (2009). Geodetic constraints on contemporary deformation in the northern Walker Lane: 3. Central Nevada seismic belt postseismic relaxation, p. 33–54, [https://doi.org/10.1130/2009.2447\(03\)](https://doi.org/10.1130/2009.2447(03)).
- Hammond, W. C., Blewitt, G., Li, Z., Plag, H.-P., & Kreemer, C. (2012). Contemporary uplift of the Sierra Nevada, western United States, from GPS and InSAR measurements. *Geology*, 40(7), 667–670. <https://doi.org/10.1130/G32968>.
- Hammond, W.C., G. Blewitt, C. Kreemer, (2016). GPS Imaging of vertical land motion in California and Nevada: Implications for Sierra Nevada uplift, *Journal of Geophysical Research - Solid Earth*, 121(10), p. 7681–7703, <https://doi.org/10.1002/2016JB013458>.
- Henry, C. D. (2009), Uplift of the Sierra Nevada, California. *Geology*, 37 (6): 575–576. <https://doi.org/10.1130/focus062009.1>
- Jones C.H. Farmer G.L. Unruh J. 2004, Tectonics of Pliocene removal of lithosphere of the Sierra Nevada, California: Geological Society of America Bulletin, v. 116, p. 1408–1422, doi: 10.1130/B25397.1.

Zandt, G., Gilbert, H., Owens, T.J., Ducea, M., Saleeby, J., and Jones, C., (2004), Active foundering of a continental arc root beneath the southern Sierra Nevada in California: *Nature*, v. 431, p. 41–46, doi:10.1038/nature02847.

Reconstruction of Late Cenozoic Fault Displacement on the Eastern Sierra Nevada Escarpment Based on Eastern Central Valley Sedimentary Structure

Fred M. Phillips (Department of Earth and Environmental Science, New Mexico Institute of Mining and Technology)

Abstract: Since the earliest days of geological exploration of the Sierra Nevada, earth scientists have debated whether the high elevation of the range dates back to the Late Cretaceous or, in considerable part, is due to tilting during the late Cenozoic. I have attempted to systematically address this question by reconstructing the dips of stratigraphic units in the eastern Central Valley that lap up onto the bedrock surface of the Sierra Nevada block. Dips vary systematically with latitude, with those of Eocene units in the southern San Joaquin Valley as steep as 5°, while those in the northern Central Valley dip only about 1°. Examination of dip as a function of formation age shows a nearly quiescent northern Sierra Nevada until about 5 Ma, while the southern end of the range has tilted slowly since the Eocene. The dips indicate that between 5 Ma and the present the northern Sierra underwent about 1 km of crest uplift and the southern Sierra about 3 km. With regard to the area visited during this FOP fieldtrip, the tilting reconstruction implies that the faults bounding the Mono Basin on the west have been offset by about 1 km in the past 5 Myr due to Sierra block tilting, while in the northern Owens Valley the offset has been 2 to 3 km. This difference is consistent with the difference in the height of the eastern Sierra escarpment at these localities.

This field trip traverses the span of the eastern Sierra Nevada from the northern Owens Valley in the south to the northern Mono Basin. A marked difference in the character of the Eastern Sierra Frontal Fault (ESFF) can be observed over this span. The west side of the Owens Valley (i.e., Wheeler Crest) is a relatively sheer escarpment about 3 km in height. Traveling northward, both the height and the steepness decrease where the fault bounds Long Valley. Between Long Valley and the Mono Basin the fault fronts a relatively subdued ridge less than 1 km above the terrain to the east. West of the Mono Basin, the escarpment steepens again and the height difference between the crest and Mono Lake is about 1.8 km.

This difference in escarpment height corresponds to a trend in crest height (Figure S10.3). The average crest elevation increases southward from about 3 km west of the Mono Basin to over 4 km west of the Owens Valley. This correspondence suggests that the height of the escarpment may be controlled, or at least influenced, by the amount of uplift of the crest. This hypothesis can be tested using a reconstruction of crest elevation as a function of time.

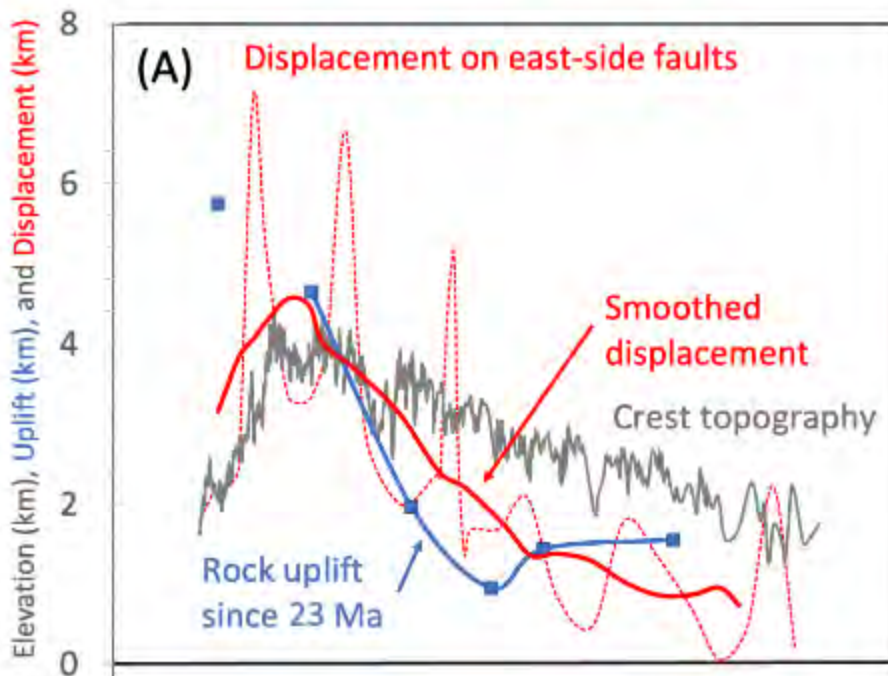


Figure S10.3. Comparison of measures of tectonic tilting in the Central Valley with vertical displacement on the Eastern Sierra fault system and with modern Sierra Nevada crest height. (A) Rock uplift since 23 Ma from this study (Figure S10.4), based on dips of Central Valley strata, compared with displacement on east-side faults. Sources for fault displacement and method of calculation are discussed in Phillips (2025). Modern Sierra crest elevation is also shown. From Phillips (2025).

The history of tilting of the Sierra Nevada block has been a matter of controversy since the earliest geological investigations of the range. G.K. Gilbert (1883) and Josiah D. Whitney (1882) argued whether canyon incision of the western flank was due to recent uplift (Gilbert) or climate change (Whitney). More recently, a large body of paleoenvironmental data has been interpreted to indicate that the crest elevation of the northern Sierra in the Eocene was about the same as today, showing that range uplift predated the Eocene, while geomorphic and thermochronological data have been taken to demonstrate substantial uplift of the range in the past several million years. See the discussion of previous studies in Phillips (2025) for a thorough review of this history.

The dips of stratigraphic units in the far eastern Central Valley that were deposited on the relatively smooth bedrock surface of the Sierra block provide a possible key to interpreting the tilting history of the block. These extend from the southern to the northern tip of the range and provide a continuous temporal record through the entire Cenozoic. The fundamental configuration of the depositional system at the eastern edge of the Central Valley has not changed much through the Cenozoic (Bartow, 1991) and thus depositional dips were likely similar to those today. However, at many localities dips are observed to increase with increasing

formation age, suggesting that the bedrock substrate had tilted with time (Bateman and Wahrhaftig, 1966). This suggestion was brilliantly pursued by Unruh (1991), who showed that dips of units in the far northern Sacramento Valley were consistent with significant tilting during the Pliocene. Jayko (2009) followed this with a broader synthesis, supporting long-term tilting of the southern Sierra Nevada.

The continuous nature of the Central Valley sedimentary record in both space and time suggested to me that a more systematic and comprehensive analysis of sedimentary unit dips might help address the longstanding disagreement over Cenozoic tectonics of the Sierra Nevada. I therefore constructed cross-sections of the formation geometries and the Sierra topography at seven locations from the far southern to the far northern Sierra. These were based on literature sources. Near the range front they largely relied on the hydrogeologic literature and in the basin interior on petroleum geology studies. These generally agreed well. In the analysis of the compiled dip data, I accounted for estimated depositional dip and for changes in dip due to sediment compaction. This tilting history was converted to crest uplift, assuming rigid behavior of the Sierra block (Figure S10.4).

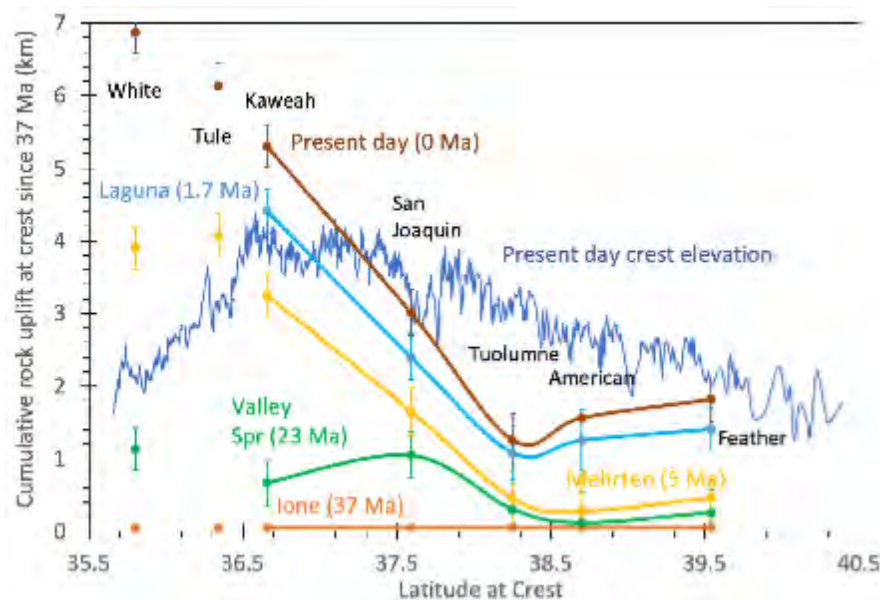


Figure S10.4 Cumulative rock uplift since 37 Ma at crest for four times, calculated assuming rigid-block rotation, as a function of latitude. The two southernmost locations are not tied to the other uplift values because the presence of normal faults between the axis of rotation and the crest allows non-rigid tilting and thus the points are overestimates of the actual uplift. From Phillips (2025).

The results reveal a distinct difference between the southern and northern Sierra Nevada. The dips indicate that the southern Sierra underwent slow tilting throughout most of the Cenozoic, followed by an episode of more rapid tilting starting about 5 million years ago and

continuing to the present. This resulted in as much as 5 km of rock uplift at the crest. In contrast, the northern Sierra was mostly stable until 5 Ma, when it underwent a brief tilting episode that lasted until about 1.7 Ma. This produced 1 to 1.5 km of uplift. The apparent dividing line between these two domains was along the course of the San Joaquin River, corresponding to the location of Mammoth Mountain on the east side of the crest. The point of minimum tilting was the Tuolumne River, corresponding to the location of Tioga Pass.

Rock uplift alone is not adequate to calculate paleoelevation; landscape erosion must also be factored in. I used a constant landscape-erosion rate of 30 m/Myr to calculate paleoelevation as a function of latitude (see discussion in Phillips (2025) for sources of data to support this erosion rate); this is shown in Figure S10.4. This figure indicates that from the Eocene to the Pliocene the crest west of the Mono Basin was the elevation high point of the Sierra Nevada, and that the elevation there has remained fairly constant. Figure S10.5 shows a little more than 1 km of rock uplift at the crest over this time, thus erosion has approximately cancelled uplift.

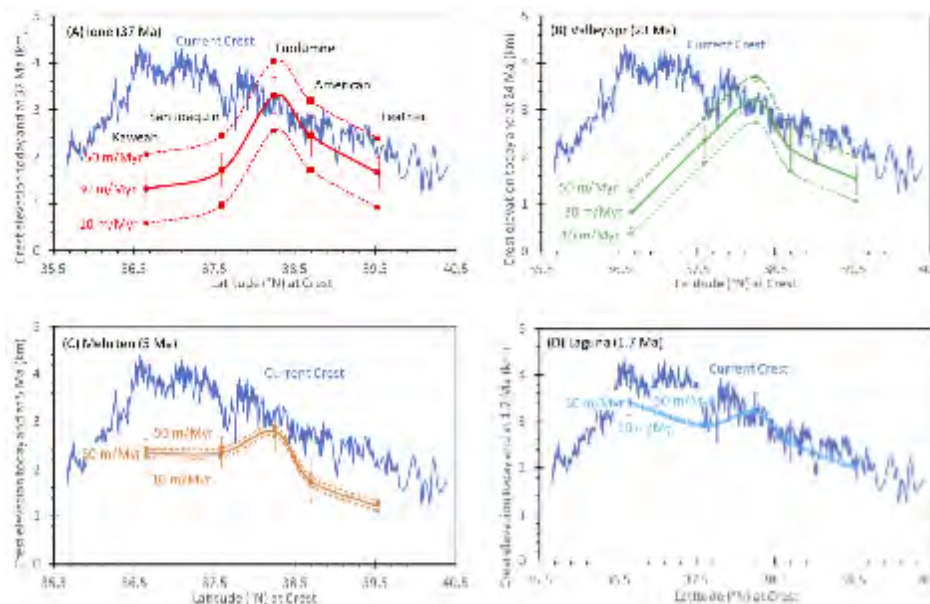


Figure S10.5. Reconstructed Sierra Nevada crest elevation for four times, including the effects of erosion. Rates in m/Ma are assumed average landscape erosion rates. For visual clarity, uncertainties are only shown for the 30 m/Ma erosion rate; those for other erosion rates are similar. From Phillips (2025).

The contrast with the northern Owens Valley (latitude $\sim 37.3^\circ$ in Figure S10.4) is great. There, the Eocene elevation was about 1.5 km, compared with a modern crest elevation of about 4 km. Total rock uplift of almost 4 km over the same period has greatly outpaced erosion,

which has totaled ~1.1 km. In the past 5 Ma the crest height has increased about 2.2 km. Thus, about two-thirds of the visible Sierra escarpment is a result of uplift since the Pliocene.

These differences have profound implications for the geomorphic nature of the Sierra landscape. This analysis indicates that the Sierra Nevada west of the Mono Basin has been in an approximate geomorphic steady state throughout the second half of the Cenozoic. In contrast, the Sierra Nevada west of the Owens Valley has been undergoing profound transient landscape evolution over this period. One outcome of this contrast has been differential preservation of sedimentary and volcanic cover on top of the batholith. From the Tuolumne River basin northward a large part of the landscape is covered by mostly Miocene intermediate volcanic rocks and older Cenozoic sedimentary rocks. South of the Tuolumne, such cover is almost entirely absent, stripped off by erosion (see Fig. 2A in Beeson and McCoy (2022)). Also, geomorphic relief is much greater south of the San Joaquin River (Fig. 2B in Beeson and McCoy(2022)).

What is responsible for this dramatic shift in behavior of the Sierra Nevada Frontal Fault at the location of the San Joaquin River? The most widely accepted explanation, and the one I accept in Phillips (2025), is that at about 5 Ma the dense arclogitic root of the Sierra batholith foundered into the deep mantle and was replaced by warm asthenospheric mantle, buoying up the Sierra block. Supporting this, Phillips et al. (2011) measured $^{40}\text{Ar}/^{39}\text{Ar}$ ages of nine small basaltic volcanos on both sides of the Sierra crest at the latitude of Bishop and obtained a mean age of 3.40 ± 0.06 Ma. Additionally, Andy Calvert and the late Wes Hildreth measured $^{40}\text{Ar}/^{39}\text{Ar}$ ages of about 50 basaltic centers in the San Joaquin drainage basin and obtained a nearly identical mean age (unpublished work). Such a large number of nearly simultaneous small eruptions over a huge area is consistent with mantle replacement as the cause. Interestingly, the northern edge of the 3.4 Ma volcanic field corresponds closely to the boundary between the relatively stable northern Sierra and the rapidly tilting southern Sierra, based on stratigraphic dip data. Ultimately, it seems probable that the transition in fault displacement and the resultant differences in geomorphology and scenery can be traced to forcings that originate far below, in the mantle.

References

- Bartow, A. J., 1991, The Cenozoic evolution of the San Joaquin Valley, California: U.S. Geological Survey Professional Paper 1501, 40 p.
- Bateman, P. C., and Wahrhaftig, C., 1966, Geology of the Sierra Nevada, *in* Bailey, E. H., ed., Geology of northern California, Volume Bulletin 190: Sacramento, California Division of Mines and Geology, p. 107–172.
- Beeson, H. W., and McCoy, S. W., 2022, Disequilibrium river networks dissecting the western slope of the Sierra Nevada, California, record significant late Cenozoic tilting and

- associated surface uplift: Geological Society of America Bulletin, v. 134, p. 2809–2853, doi:10.1130/B35463.1.
- Gilbert, G. K., 1883, Whitney's climatic changes, Part III: Science, v. 1, p. 192–195.
- Jayko, A. S., 2009, Deformation of the late Miocene to Pliocene Inyo Surface, eastern Sierra region, California, *in* Oldow, J. S., and Cashman, P. H., eds., Late Cenozoic Stucture and Evolution of the Great Basin-Sierra Nevada Transition: Boulder, Colorado, Geological Society of America Special Paper 447, p. 313–350.
- Phillips, F. M., 2025, Sedimentary stratigraphy and structure of the eastern Central Valley, California, USA: Implications for Cenozoic tectonics of the Sierra Nevada: Geosphere, v. 21, p. 514–547, doi:10.1130/GES02809.1.
- Phillips, F. M., McIntosh, W. C., and Dunbar, N. W., 2011, The chronology of late Cenozoic volcanic eruptions on relict surfaces in the south-central Sierra Nevada, California: Geological Society of America Bulletin, v. 123, p. 890–910, doi:10.1130/B30000.1.
- Unruh, J. R., 1991, The uplift of the Sierra Nevada and implications for late Cenozoic epeirogeny in the western Cordillera: Geological Society of America Bulletin, v. 103, p. 1395–1404, doi:10.1130/0016-7606(1991)103<1395:TUOTSN>2.3.CO;2.
- Whitney, J. D., 1882, The Climatic Changes of Later Geological Times: a Discussion Based on Observations Made in the Cordilleras of North America, Cambridge, Massachusetts, Cambridge University Press, 394 p.

Geomorphic implications of fracture density differences in granitic rocks across the Sierra Nevada crest

Contributor: Greg Stock (National Park Service, Yosemite National Park)

The descent through Lee Vining Canyon on Highway 120 reveals a remarkable array of geological features, including intensely fractured and altered granitic bedrock, a deep glacially sculpted canyon, expansive talus and debris-flow deposits, and massive lateral moraines. Are all of these features all related?

The bedrock here is composed of Jurassic age (~218 Ma) granitic rocks (Barth et al., 2011). Note the fracture spacing, which is on the order of 1-2 m (Figure S10.6). This contrasts strongly with late Cretaceous (~92-86 Ma) granitic bedrock of the Tuolumne Intrusive Suite (Bateman, 1992) ~20 km west in the Tuolumne Meadows region, where glacially sculpted bedrock exhibits fracture spacing on the order of tens of meters (Figure S10.6). Fracturing reduces rock strength, but also (and perhaps more importantly) breaks bedrock into discrete blocks that are easily extracted and transported by geomorphic processes (Molnar et al., 2007). What controls the fracture density in Lee Vining Canyon has not been studied in detail, but important factors likely include proximity to faults (e.g., Molnar et al., 2007), shear and alteration zones, and metamorphic contacts. All of these are present in Lee Vining Canyon (Kistler, 1966) but essentially absent in the core of the Tuolumne Intrusive Suite. The granitic bedrock in Lee Vining

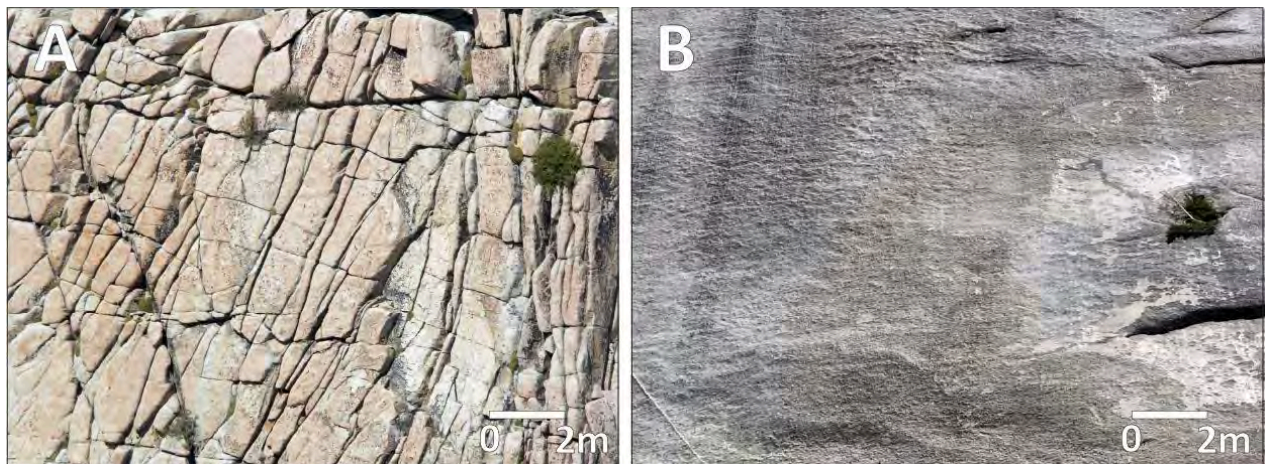


Figure S10.6. Comparison of bedrock fracture density on opposite sides of the central Sierra crest. (A) Granitic bedrock on the north wall of Lee Vining Canyon showing dense fracturing, with fracture spacing on the order of 1-2 m. (B) North face of Pywiack Dome near Tenaya Lake, 22 km west of (A), showing massive granitic bedrock generally lacking fractures. Note similar scale for both photos.

Canyon is positioned in a “rock crusher” between the late Cretaceous (and possibly reactivated) Sierra Crest Shear Zone to the west (Tikoff and de Saint Blanquat, 2010) and the normal range-front Mono Lake Fault (1.0-5.0 mm/yr slip rate) to the east (USGS and CGS, 2026). Rock damage in the footwall of the intermediate normal Silver Lake Fault (0.2-1.0 mm/yr slip rate), located 1.5 km east of this stop, and another unnamed fault 5 km farther east (USGS and CGS, 2026), are of particular interest with respect to the dense rock fracturing in this area (e.g., Berg and Skar, 2005).

Dense fracturing has contributed to extensive rockfall talus and debris flow deposits in Lee Vining Canyon. On the north wall of the canyon, Highway 120 crosses a large talus slope, known as “Blue Slide” (Figure S10.7), an accumulation of rockfall debris from the densely fractured mountain slope above. The roadcut has locally steepened the talus slope above the highway, producing frequent rockfalls onto the road.

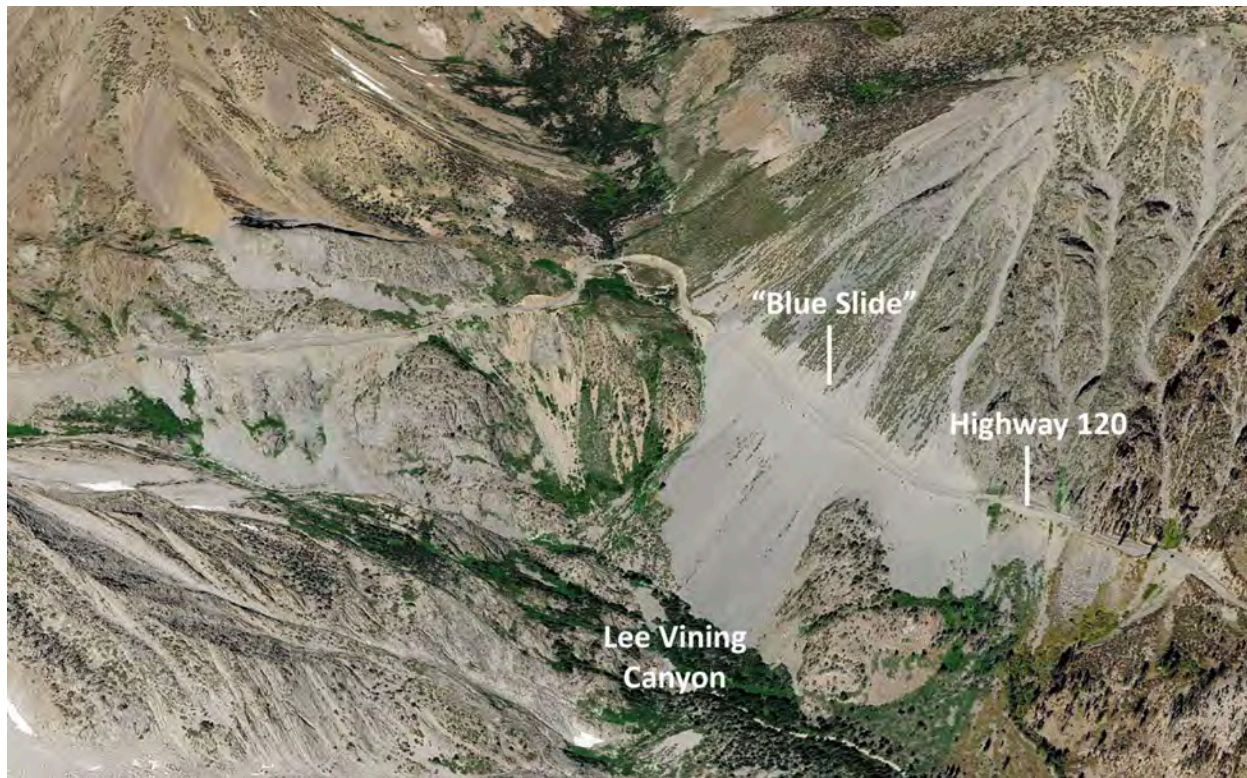


Figure S10.7. Oblique view to the northwest of the “Blue Slide” in Lee Vining Canyon, a large talus slope cut by Highway 120, resulting in frequent rockfalls onto the road.

Dense fracturing in Lee Vining Canyon may also help to explain the unusually large lateral moraines bounding the lower canyon (Figure S10.8). Although glaciers on the western slope of the Sierra were ~5-7 times longer than those on the eastern slope (Kessler et al., 2006), east-side lateral and terminal moraines are typically much larger than their west-side

counterparts. As elaborated in Glazner and Stock (2010), differences in moraine size may be influenced by a number of factors, including variations in topography, precipitation, and erosional processes across the Sierra crest. However, rock fractures have been shown to increase the efficiency of glacial quarrying (e.g., Theiss et al., 2026), and ^{10}Be concentrations in deglaciated bedrock in the Tuolumne watershed demonstrate that fracture spacing exerts strong control on glacial erosion rates, which are quite low in unfractured rock (Dühnforth et al., 2010). Thus, the dense fracturing of granitic bedrock here could explain the especially large moraines in lower Lee Vining Canyon—and perhaps other eastern Sierra canyons.



Figure S10.8. Aerial view eastward showing nested moraines in lower Lee Vining Canyon. Their large size compared to the relatively small catchment, and to moraines associated with similarly sized west-side catchments, may be linked to the densely fractured bedrock in Lee Vining Canyon. Photo by Allen Glazner.

References

- Barth, A.P, Walker, J.D., Wooden, J.L., Riggs, N.R., and Schweickert, R.A. (2011) Birth of the Sierra Nevada magmatic arc: Early Mesozoic plutonism and volcanism in the east-central Sierra Nevada of California: *Geosphere*, v. 7 (4), p. 877-897, <https://doi.org/10.1130/GES00661.1>
- Bateman, P.C. (1992) Plutonism in the central part of the Sierra Nevada batholith, California: U.S. Geological Survey Professional Paper 1483, 186 p.
- Berg, S.S., and Skar, T. (2005) Controls on damage zone asymmetry of a normal fault zone: outcrop analyses of a segment of the Moab fault, SE Utah. *Journal of Structural Geology* 27, 1803-1822, <https://doi.org/10.1016/j.jsg.2005.04.012>
- Dühnforth, M., Anderson, R.S., Ward, D.J., and Stock, G.M. (2010). Bedrock fracture control of glacial erosion processes and rates: *Geology*, v. 38 (5), p. 423–426. <https://doi.org/10.1130/G30576.1>
- Glazner, A.F., and Stock, G.M. (2010) *Geology Underfoot in Yosemite National Park*. Mountain Press, Missoula, Montana, 312 p.

- Kessler, M.A., Anderson, R.S., and Stock, G.M. (2006) Modeling topographic and climatic control of east-west asymmetry in Sierra Nevada glacier length during the Last Glacial Maximum: *Journal of Geophysical Research—Earth Surface*, v. 111, F2, <https://doi.org/10.1029/2005JF000365>
- Kistler, R.W. (1966) Geologic map of the Mono Craters quadrangle, Mono and Tuolumne Counties, California. U.S. Geological Survey GQ-462, https://ngmdb.usgs.gov/Prodesc/proddesc_349.htm
- Molnar, P., Anderson, R.S., and Anderson, S.P. (2007) Tectonics, fracturing of rock, and erosion: *Journal of Geophysical Research—Earth Surface*, 112 F03014, <https://doi.org/10.1029/2005JF000433>
- Theiss, C.R., Cuffey, K.M., Xia, C., and Zhao, Q. (2026) How glaciers exploit prior fractures to quarry beds of partially intact rock: *Journal of Glaciology*, v. 72, e29, 1-13, <https://doi.org/10.1017/jog.2026.10141>
- Tikoff, B., and de Saint Blanquat, M. (2010) Transpressional shearing and strike-slip partitioning in the Late Cretaceous Sierra Nevada magmatic arc, California. *Tectonics* 16(3), 442-459, <https://doi.org/10.1029/97TC00720>
- U.S. Geological Survey and California Geological Survey, Quaternary fault and fold database for the United States, accessed August 18, 2026, at: <https://www.usgs.gov/natural-hazards/earthquake-hazards/faults>

FOP 2026 DAY 3



Day 3, S11: Wilson Creek Canyon

FIELD TRIP STOP CONTRIBUTORS: Sidney Hemming (Lamont-Doherty Earth Observatory, Columbia University), Jorge Vazquez (US Geological Survey), Scott Stine, Emeritus California State University, East Bay

LOGISTICS: Park on the gravel road and then we will hike together to the Wilson Creek Canyon. Terrain will be uneven.

WALKING DISTANCE: 1 mile

LOCATION: Wilson Creek Canyon (38.02869, -119.13120)

DIRECTIONS: [Brown's Owens River Campground to Wilson Creek Canyon](#)



Wilson Creek Canyon Stratigraphic and Geochemical Evidence for Lake Level Variations through the Last Glacial Cycle

Contributor: Sidney Hemming (Lamont-Doherty Earth Observatory, Columbia University)

The presenter, Sidney Hemming, is representing the work of three former Ph.D. students who are unable to attend this FoP. Susan Zimmerman conducted Ph.D. research on the Wilson Creek Formation (Zimmerman, 2006), and published papers on the relative paleomagnetic intensity (Zimmerman et al., 2006), ice rafted detritus (Zimmerman et al., 2011a) and chemostratigraphy (Zimmerman et al., 2011b). The 2011 papers provide an integrated examination of the stratigraphic evidence for the lake level history presented in this section. Rahul Sahajpal conducted Ph.D. research on geochemical proxies of lake level in the Mono Basin (Sahajpal, 2013) and published papers on the Li concentration in leachates from Wilson Creek Formation samples (Sahajpal et al., 2011) and thermodynamic calculations of water chemistry with evaporation and dilution (Sahajpal et al., 2021). Guleed Ali conducted Ph.D. research using high precision U/Th dating of clean carbonate occurrences on the stratigraphic and geomorphic evidence for lake level variation through Wilson Creek Formation time (Ali, 2017) and published papers on the rise of lake level at the beginning of Wilson Creek Formation deposition (Ali et al., 2022) and on the age of the famous magnetic excursion in the Mono Basin (Ali et al., 2026). He is also preparing further publications that stem from the Ph.D. work, with significant additional advances (Ali et al., in preparation-a; Ali et al., in preparation-b).

The Wilson Creek Formation, beautifully and famously exposed in Wilson Creek Canyon, was named and described by Ken Lajoie in his iconic 1968 thesis. Lajoie (1968) documented exposures of the formation all around the Mono Basin, and detailed the facies distributions that allowed putting the sedimentary packages into a lake level context. A figure created by Lajoie for Susan Zimmerman (Zimmerman et al., 2011b) is presented here as Figure 1.

Three deltaic/terrace sand and gravel sequences interfinger with three lacustrine silt sequences. Importantly, a large number of ash layers allows correlation of the stratigraphy around the basin. Lajoie (1968) documented 19 ash layers in packages A-E, both counted from top to bottom. Guleed Ali (personal communication) has found a number of the ashes within riverine and littoral sequences that tie them directly to shoreline position, providing a lake level estimate at those specific times.

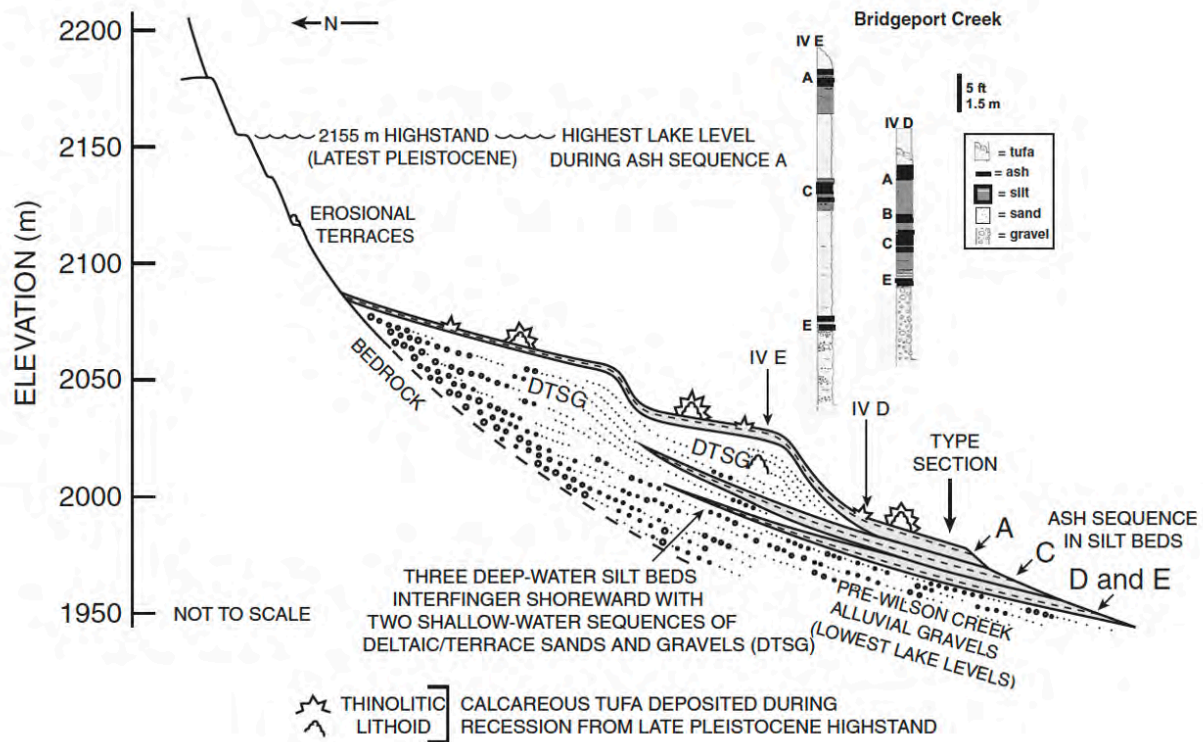


Figure S11.1. Schematic cross section of the Wilson Creek Formation in the Mono Lake basin based on Lajoie (1968). At elevations below ~2000 m, the formation is composed of lacustrine silts and preserves the complete ash sequence everywhere; between ~2000 and ~2050 m, deep-lake silts interfinger with deltaic and nearshore sands; only ash package A occurs in deep-lake silts between ~2050 and 2085 m. Thinolitic and lithoid tufa occur on and within the Wilson Creek Formation as mounds, fans, and plates. Inset: Bridgeport Creek exposures display the typical stratigraphy: IV E above the break of the lower terrace shows three sequences of silts separated by deltaic sands, while IV D below the slope break preserves the complete series of ashes in deep-lake silt. Figure courtesy of K.R. Lajoie. Figure reproduced from Zimmerman et al. (2011b).

As viewed in lower Wilson Creek Canyon type section, the Wilson Creek Formation is almost entirely composed of lacustrine silts and volcanic ashes (Fig. 1, Fig. 2). It is possible, even likely, that there were multiple drops in lake level below this elevation, but they are cryptic. However, between Ash packages B and A, the lake level dropped significantly below the level of this outcrop (detailed in Ali, 2017, and Ali et al., paper in preparation-a). A cut and fill event can be seen nearby in the Mill Creek exposure but it is too steep for a large group to visit. Guleed Ali (personal communication) has traced this low stand to 1990–1980 m, the lowest levels recorded in the WCF and as much as 75 m below the previously recognized low stand, and has constrained the age to between 26.6 and 23.0 ka (Ali et al., paper in preparation-a). He has also

documented slumped sediment within the sequence that contains the package B ashes, confirming that the drop in lake level postdated ash package B. And ash package A, in turn, overlies the channel fill within silt layers that rest conformably on the beds below. Package A ashes can be found in lacustrine silt as high as 2143 m (Guleed Ali, personal communication). Guleed Ali has bracketed the timing of the dramatic rise at the end of Wilson Creek Formation time to between 16.12 ± 0.05 and 15.92 ± 0.04 ka—an interval spanning 180 ± 90 years (Ali et al., paper in preparation-b). At Wilson Creek Canyon the expression of the low stand is seen as a discontinuous sand layer overlain by deformed silts containing some sand (Fig. 2).

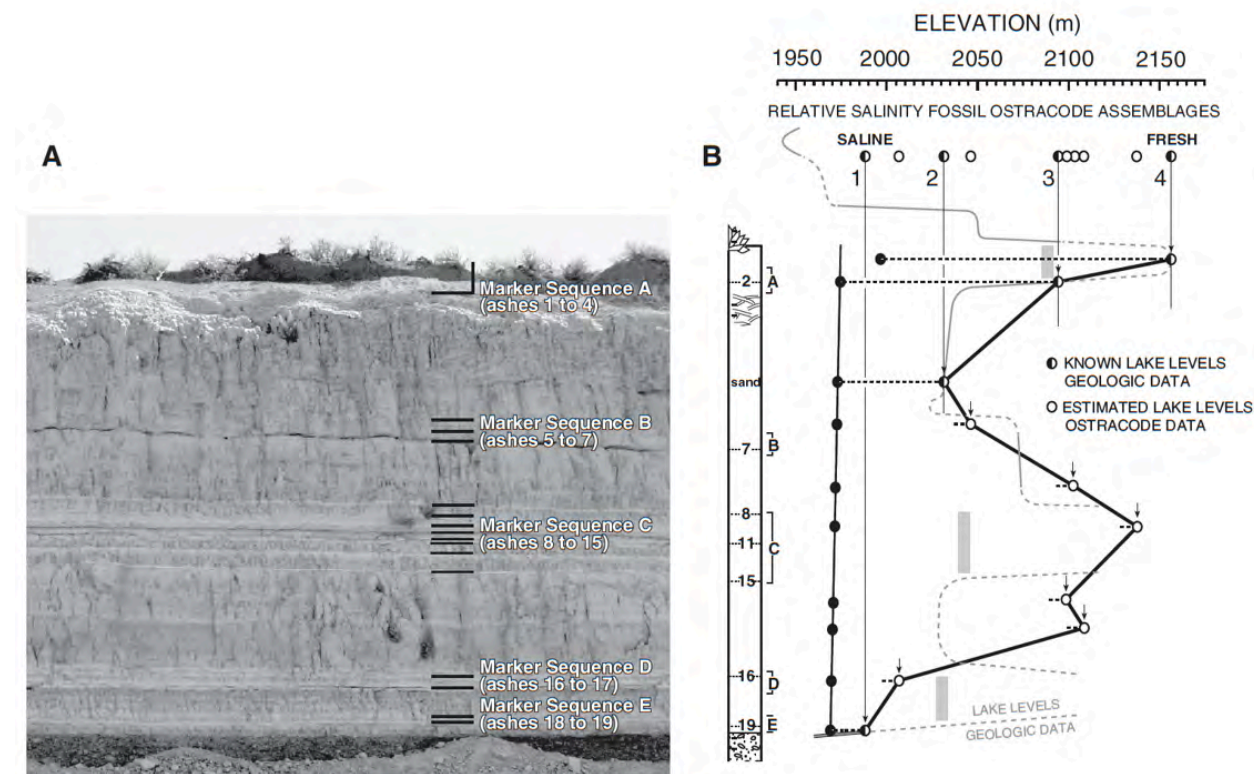


Figure S11.2. (A) Photograph of the type section of the Wilson Creek Formation, in the lower Wilson Creek canyon. Eighteen rhyolitic ashes and one basaltic ash were grouped by Lajoie (1968) into five Marker Sequences, useful for intrabasin correlation. Above Marker Sequence B (Ash 5), there is a discontinuous sand layer (yellow arrow) and beds are inclined and include thin sand lenses (highlighted with bracket labeled “non-horizontal”). (B) Stratigraphic column (modified from Lajoie, 1968), showing major ash layers and bundling into Marker Sequences A–E. Lake-level curve (gray line) is based on stratigraphic evidence, with ostracode-based relative salinity curve (black line), modified by K.R. Lajoie from Forester (1987). Black dots on vertical line indicate stratigraphic position of each ostracode sample; half-filled dots indicate known lake levels based on independent geologic data; open dots indicate relative salinity of ostracode samples without a priori lake-level information. Numbers below open dots indicate geologic control on lake level: 1—lake level just above base of type section, ~15 m depth;

2—lake at level of lower terrace (~2035 m elevation); 3—lake level at time of eruption of Ash 2, constrained by small delta of ash above Mono Post Office location on the Sierran front; 4—lake level at deposition of Ash 1, highest level of Wilson Creek time, based on deep-lake silts deposited at 2085 m elevation. Figure modified from Zimmerman et al. (2011b).

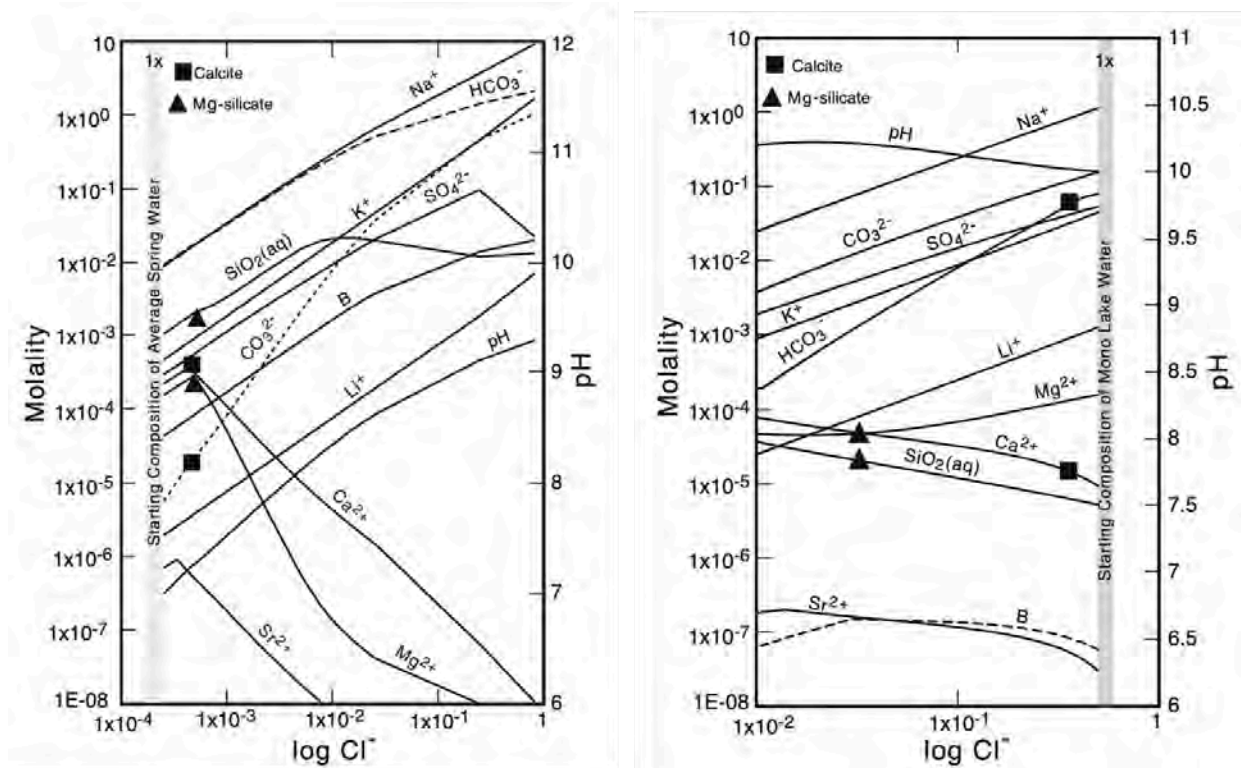


Figure S11.3. A) Calculated results of various chemical species plotted against Cl^- concentration ($\log 0.0001 = 1\times$ and $\log 1 = 10,000\times$) with evaporation of average spring waters feeding Mono Lake at 25°C and $\text{CO}_2 = 350$ ppm. These results from our simulation of evaporation of spring water closely match the theoretical results of evaporation of spring water proposed by Garrels and Mackenzie (1967). The inflection point for any element indicates the conditions under which it starts to precipitate. The symbols indicate the conditions under which minerals like calcium carbonate and Mg silicate precipitate. This is the Garrels and Mackenzie (1967) model. B) Calculated results of various chemical species plotted against Cl^- concentration from evaporation and mixing-based model predicting the evolution of Mono Lake from the present-day lake to its spill level at 25°C and $\text{CO}_2 = 350$ ppm. Both figures reproduced from Sahajpal et al. (2021)

The ability to connect geodetic lake level information and ages (Fig. 2; also Guleed Ali, personal communication on ashes in shoreline deposits) with proxies (Fig. 2, also Zimmerman et al., 2011b; Sahajpal et al., 2011) provides a powerful approach for ground-truthing stratigraphic and geochemical proxies of lake level. Mono Lake has extreme geochemistry due to its closed

basin hydrology and because of a slightly higher carbonate to calcium ratio than CaCO_3 . Garrels and Mackenzie (1967) famously modeled the composition of Mono Lake water by a simple theoretical evaporation experiment with thermodynamic calculations (Fig. 3a). Sahajpal et al. (2021) explored the effects of variations in the fresh water contribution, and extended the thermodynamic simulations to include dilution from a highly evaporated (modern) composition (Fig. 3b). These calculations help explain the variations in carbonate and Mg-silicate phases through the formation, and set the stage for more refined geochemical proxy development.

References

- Ali, Guleed, 2018, Last glacial and deglacial fluctuations of Mono Lake, California [Ph.D. thesis] Columbia University.
- Ali, G. A. H., Lin, K., Hemming, S. R., Cox, S. E., Ruprecht, P., Zimmerman, S. R. H., Stine, S., and Wang, X., 2022, Emergence of wet conditions in the Mono Basin of the Western USA coincident with inception of the Last Glaciation, *GSA Bulletin* 134, 2267–2279; <https://doi.org/10.1130/B36084.1>
- Ali, G., Lin, K., Hemming, S., and Wang, X., 2026, Carbonate U/Th ages confirm the Mono Basin excursion as the Laschamps geomagnetic excursion, *GSA Bulletin* 138, 2910–2918; <https://doi.org/10.1130/B38358.1>
- Ali, G. and others, in prep- a, Great Basin lakes during the Last Glacial Maximum
- Ali, G. and others, in prep- b, Deglacial lake level fluctuations of Mono Lake, California
- Forester, R.M., 1987, Late Quaternary paleoclimate records from lacustrine ostracodes, in Ruddiman, W.F., and Wright, H.E., Jr., eds., *North American and Adjacent Oceans during the Last Deglaciation: Boulder, Colorado, The Geology of North America*, Geological Society of America, p. 261–276.
- Garrels, R.M., and Mackenzie, F.T., 1967, Origin of the chemical compositions of some springs and lakes, in Gould, R.F., ed., *Equilibrium Concepts in Natural Water Systems*: Washington, D.C., American Chemical Society Publication, p. 222–242, <https://doi.org/10.1021/ba-1967-0067.ch010>.
- Lajoie, K.R., 1968, Quaternary Stratigraphy and Geologic History of Mono Basin, Eastern California [Ph.D. thesis]: Berkeley, California, University of California.
- Sahajpal, Rahul, 2013, Developing geochemical proxies for a high resolution hydroclimate record in Mono Lake basin [Ph.D. thesis]: City University of New York
- Sahajpal, R., Zimmerman, S. R. H., Datta, S., Hemming, N. G., and Hemming, S. R., 2011, Assessing Li and other leachable geochemical proxies for paleo-salinity in lake sediments from the Mono Basin, CA (USA), *Geochimica et Cosmochimica Acta* 75, 7855–7863, [10.1016/j.gca.2011.10.001](https://doi.org/10.1016/j.gca.2011.10.001)
- Sahajpal, R., Hemming, N. G., Rouff, A. A., Hemming, S. R., Zimmerman, S., and Sahajpal, R., 2021, Thermodynamic evaporation and freshwater mixing models to test salinity proxies

- for late Pleistocene lake levels, Mono Lake, California, in Editor(s) Scott W. Starratt and Michael R. Rosen, *From Saline to Freshwater: The Diversity of Western Lakes in Space and Time*, GSA Special Paper 536, <https://doi.org/10.1130/SPE536>
- Zimmerman, Susan, 2006, Chronology and paleoclimate records of the late Pleistocene Wilson Creek Formation at Mono Lake, California [Ph.D. thesis] Columbia University.
- Zimmerman, S.H., Hemming, S., Kent, D.V., and Searle, S., 2006, Revised chronology for late Pleistocene Mono Lake sediments based on paleointensity correlation to the global reference curve: *Earth and Planetary Science Letters*, v. 252, p. 94–106, <https://doi.org/10.1016/j.epsl.2006.09.030>
- Zimmerman, S.R.H., Pearl, C., Hemming, S.R., Tamulonis, K., Hemming, N.G., and Searle, S.Y., 2011a, Freshwater control of ice rafted debris in the last glacial period at Mono Lake, California, USA: *Quaternary Research*, v. 76, p. 264–271, <https://doi.org/10.1016/j.yqres.2011.06.003>
- Zimmerman, S.R.H., Hemming, S.R., Hemming, N.G., Tomascak, P.B., and Pearl, C., 2011b, High-resolution chemostratigraphic record of late Pleistocene lake-level variability, Mono Lake, California: *Geological Society of America Bulletin*, v. 123, p. 2320–2334, <https://doi.org/10.1130/B30377.1>

Day 3 S12: Wilson Creek Formation at Grand Gash

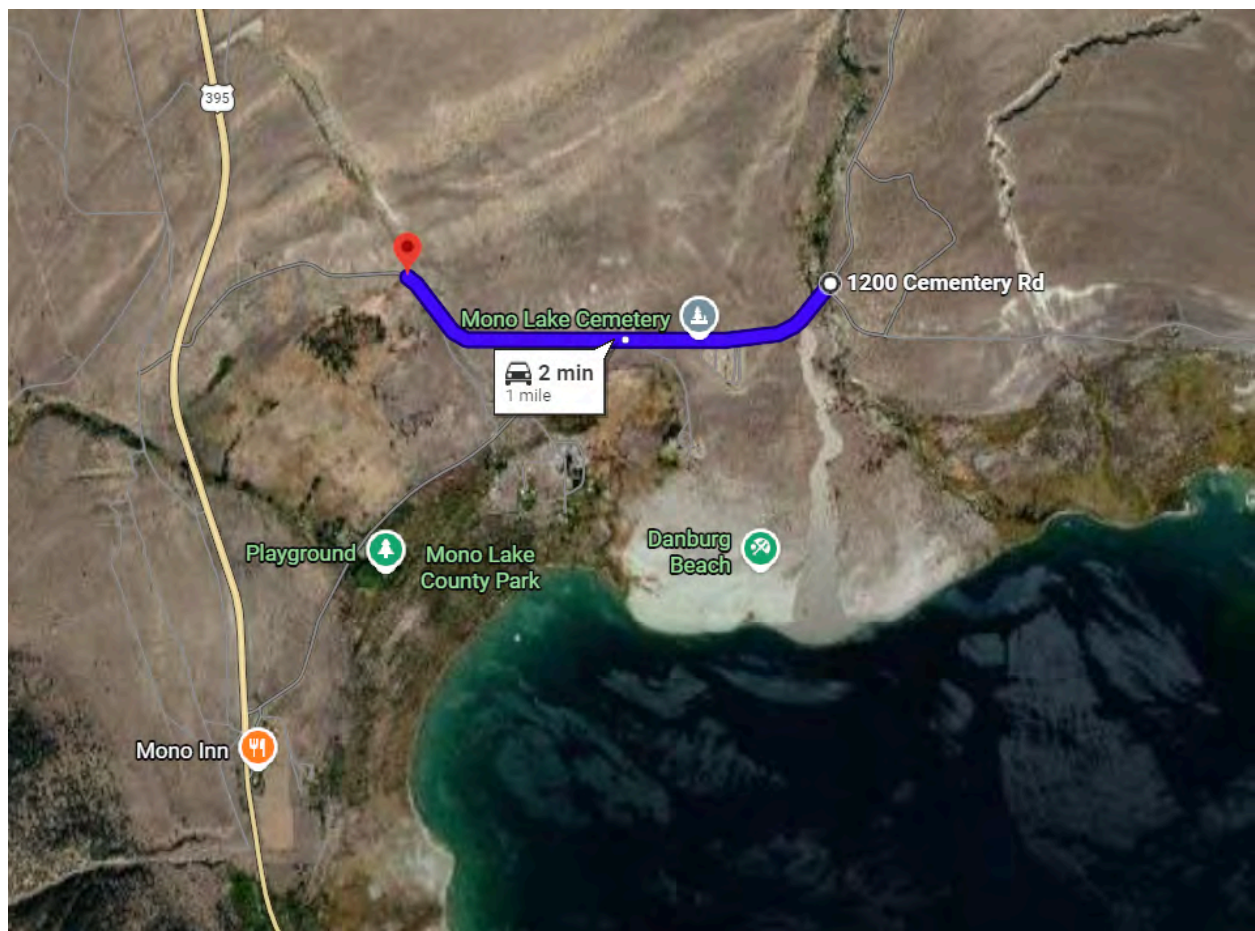
FIELD TRIP STOP CONTRIBUTORS: Marith Reheis (Retired, Sidney Hemming (Lamont-Doherty Earth Observatory, Columbia University), Jorge Vazquez (US Geological Survey), Scott Stine, Emeritus California State University, East Bay

LOGISTICS: Park on the large gravel road and hike to the Grand Gash. No walking required for this stop! The goals of this stop are to observe fractured granitic bedrock in Lee Vining Canyon and discuss possible geomorphic implications.

WALKING DISTANCE: 0.3 miles

LOCATION: Wilson Creek Formation at Grand Gash (38.023631, -119.148737)

DIRECTIONS: [Wilson Creek Canyon to Grand Gash](#)



Pleistocene history of Lake Russell

Contributor: Marith Reheis

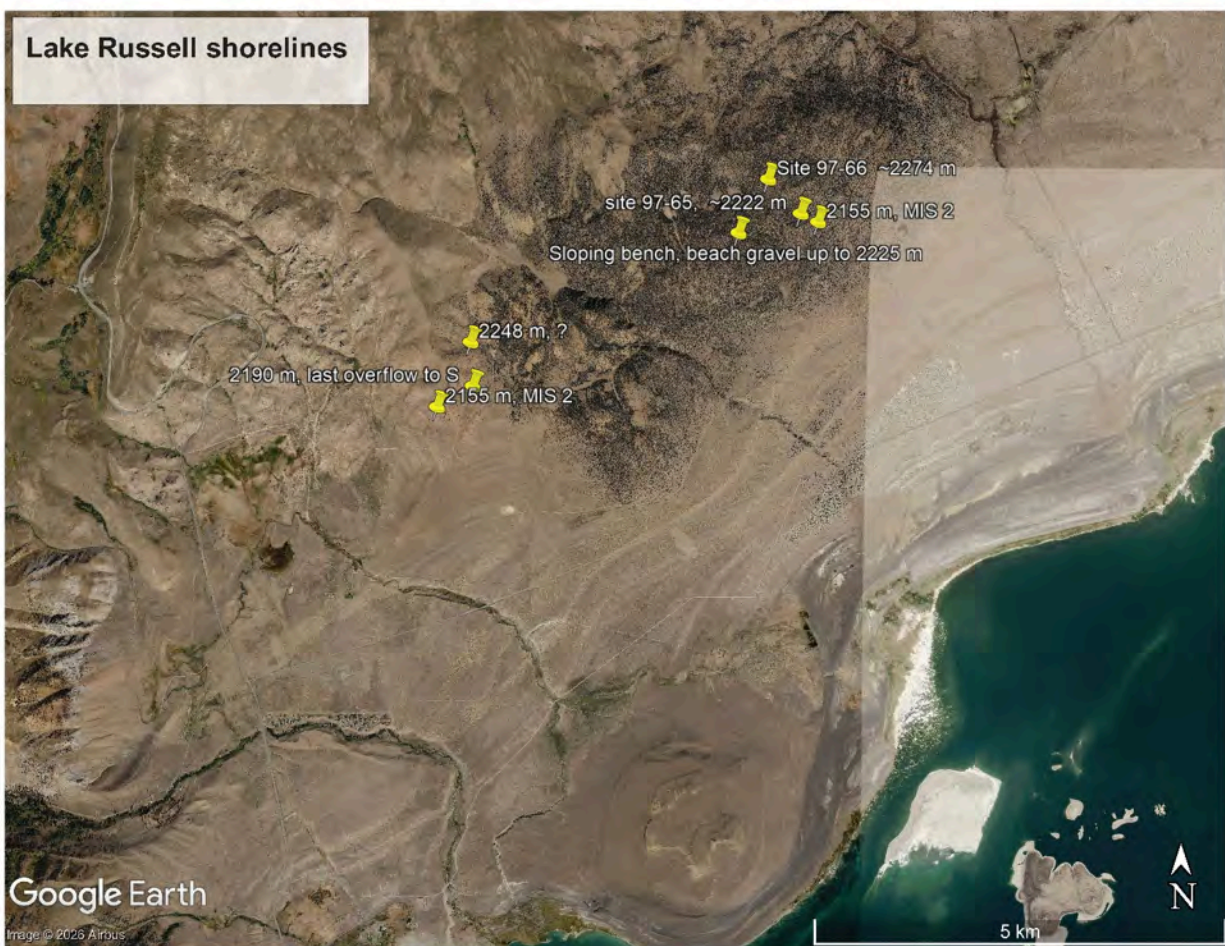
A lake has at least episodically existed in the Mono Basin since the late Pliocene (Gilbert et al., 1968), and likely continuously during the Pleistocene and Holocene. Pluvial Lake Russell was named in honor of Israel Russell, who first studied this lake (Russell, 1889). This stop will briefly review the limited evidence for periods during the Pleistocene when shoreline elevations were episodically far higher than the last highstand (2155 masl) during MIS 2. During these periods, Lake Russell sometimes overflowed to the north into the Lahontan drainage basin, and later to the south into the Owens drainage basin. Following are the abstract and Figure 2 from Reheis et al. (2002) that describe these relations; please read the publication for details.

“Mono Basin, on the eastern flank of the central Sierra Nevada, is the highest of the large hydrographically closed basins in the Basin and Range province. We use geomorphic features, shoreline deposits, and basalt-filled paleochannels to reconstruct an early to middle Pleistocene record of shorelines and changing spillways of Lake Russell in Mono Basin. During this period of time, Lake Russell repeatedly attained altitudes between 2205 and 2280 m—levels far above the present surface of Mono Lake (~1950 m) and above its last overflow level (2188 m). The spill point of Lake Russell shifted through time owing to late Tertiary and Quaternary faulting and volcanism. During the early Pleistocene, the lake periodically discharged through the Mount Hicks spillway on the northeastern rim of Mono Basin and flowed northward into the Walker Lake drainage basin via the East Walker River. Paleochannels recording such discharge were incised prior to 1.6 Ma, possibly between 1.6 and 1.3 Ma, and again after 1.3 Ma (ages of basaltic flows that plugged the paleochannels). Faulting in the Adobe Hills on the southeastern margin of the basin eventually lowered the rim in this area to below the altitude of the Mount Hicks spillway. Twice after 0.76 Ma, and possibly as late as after 0.1 Ma, Lake Russell discharged southward through the Adobe Hills spillway into the Owens–Death Valley system of lakes. This study supports a pre- Pleistocene aquatic connection through Mono Basin between the hydrologically distinct Lahontan and Owens–Death Valley systems, as long postulated by biologists, and also confirms a probable link during the Pleistocene for species adapted to travel upstream in fast-flowing water. Keywords: biogeography, drainage changes, fish, Great Basin, Mono Basin.”

- Reheis, M.C., Stine, S., and Sarna-Wojcicki, A.M., 2002, Drainage reversals in Mono Basin during the late Pliocene and Pleistocene: Geological Society of America Bulletin, v. 114, p. 991–1006.
[doi.org/10.1130/0016-7606\(2002\)114%3C0991:DRIMBD%3E2.0.CO;2](https://doi.org/10.1130/0016-7606(2002)114%3C0991:DRIMBD%3E2.0.CO;2)
- Russell, I.C., 1889, Quaternary history of Mono Valley, California: U.S. Geological Survey Annual Report, v. 8, p. 261–394.

Stop 13 optional, Day 3—*Hike to beach deposits of Early to Middle Pleistocene Lake Russell*

Starting from the intersection of a gravel road with CA 167 (38.0570 N, 119.1166 W), drive N on the gravel road for 0.5 mi, then E and NE on a gravel road that skirts the front of the hills. Drive 3.2 miles on this road, crossing Rancheria Gulch, veer left at a fork (if passable), and park at a faint two-track that climbs to the NE (38.0883 N, 119.0791 W). Hike up this road aiming for the top of a hill (labeled 7291T on 7.5' topographic map) about 0.4 mile from the parking spot.



On Figure 2 of Reheis et al. (2002), reproduced above, this point (Station 97-65, field notes) and another location higher up the slope (Station 97-66) are shown as elevations of polished, well rounded, lacustrine pebbles atop bedrock at elevations of 2225 and 2274 masl. These locations are significantly higher than both the MIS 2 highstand of Lake Russell at 2155 masl and the last southward overflow of the lake during MIS 6 (or 4) at 2188 masl. These and other high-elevation sites are interpreted to represent a lake that episodically discharged northward via the Mt. Hicks spillway to enter the Lahontan drainage basin during the Early to early Middle Pleistocene. Original field notes state the following: “Top of hill 7291T has beach gravel right on top; slopes below have locally abundant concentrations of beach gravel, though they are almost all fractured and not entire clasts like the [MIS 2] beach gravel....Top of hill also has a ‘medicine

ring' overlooking the lake." The photos below depict the view from this site and beach clasts on a rock within that ring. Beach clasts can be traced uphill from this site to as high as ~2274 masl. They also blanket a gently sloping bench that lies SW of hill 7291T (site 97-65) to an elevation of ~2225 masl.

