

Drainage reversals in Mono Basin during the late Pliocene and Pleistocene

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ABSTRACT

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.

INTRODUCTION

The Mono Basin abuts the eastern flank of the central Sierra Nevada (Fig. 1). This young structural depression (Gilbert et al., 1968; Pakiser, 1976; Huber, 1981) is the highest of the large hydrographically closed basins in the Basin and Range province. Mono Lake is the saline, alkaline remnant of Pleistocene Lake Russell, which occupied Mono Basin (Fig. 2). Throughout much of the Pleistocene the lake stood at altitudes far above its present surface (1958 m in A.D. 1950) and its late Pleistocene levels; at times the lake spilled into adjacent watersheds through gaps in the basin rim. We find that the spill point of Lake Russell shifted through time, and we infer that the shifts were caused by late Tertiary and Quaternary faulting and volcanism (Gilbert et al., 1968; Bailey et al., 1976; Metz and Mahood, 1985; Bursik and Sieh, 1989). During the early Pleistocene, the lake periodically spilled northward into the Lahontan drainage basin; afterward, it spilled southward into the Owens–Death Valley system of lakes.

A high-standing basin with shifting spill points has the capacity to function as a biogeographic “switchyard” through which obligately aquatic species (those that cannot sur-

vive out of water for even short periods) can expand their range across otherwise insurmountable drainage divides. Biologists, pointing to several such species that today occur in the hydrologically distinct Lahontan and Owens–Death Valley systems (e.g., Hubbs and Miller, 1948; Minckley et al., 1986; Hershler and Sada, 2002), have long speculated that Mono Basin may have linked these two systems at some time in the past (e.g., Hubbs and Miller, 1948; Taylor, 1985). This study confirms that linkage and places the transfer from north to south at sometime after 1.3 Ma and likely after 0.76 Ma.

Hydrographic Setting

The rim of Mono Basin is much lower on the east than on the north, west, and south. This eastern rim is breached in several places by low divides, two of which show evidence of past scouring by lake overflow (Fig. 2). The lowest divide at present lies in greatly faulted terrain of the Adobe Hills at an altitude of 2189–2190 m. This feature, here informally named the “Adobe Hills spillway,” is the gap through which the lake would spill today if it were to fill the basin. A south-draining channel heading at this gap carried the last overflow of Lake Russell into Adobe Valley and thence into the Owens–Death Valley system (Figs. 1 and 2; Putnam, 1949).

Farther north along the east rim of the basin, at the north edge of a small subbasin called Alkali Valley, lies the informally named “Mount Hicks spillway.” At an altitude of 2237 m, it stands ~47 m above the Adobe Hills spillway. A channel that heads at this

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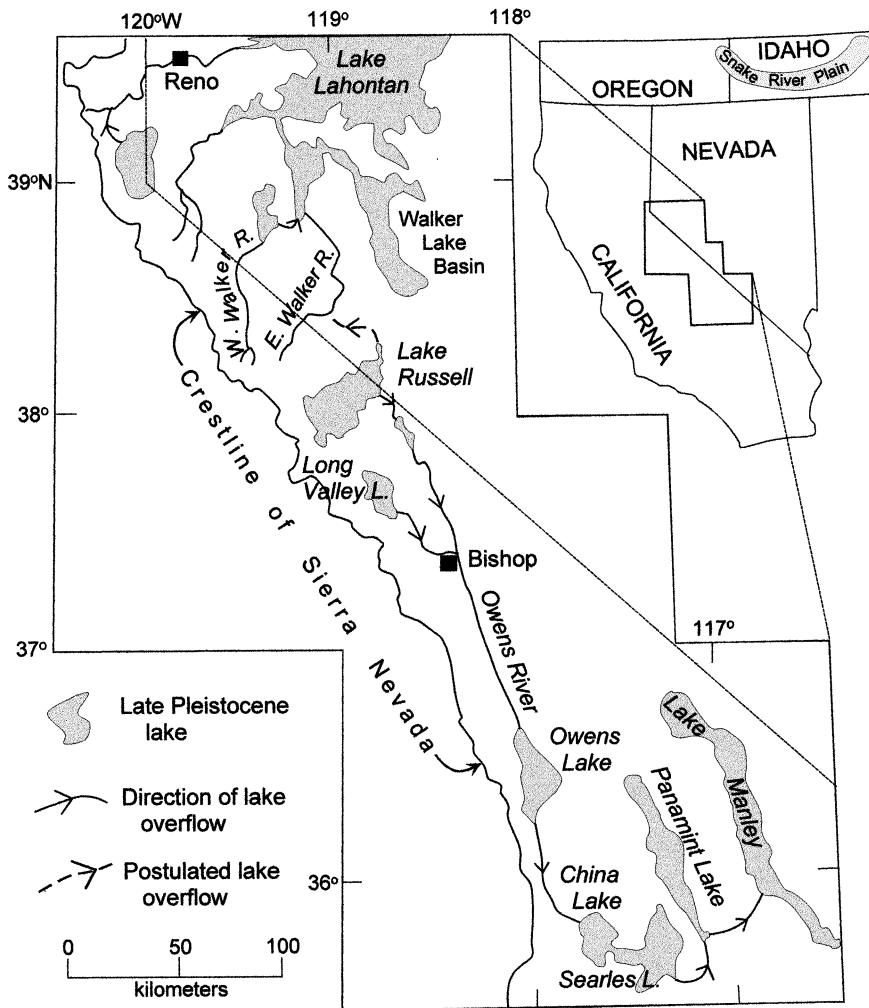


Figure 1. Regional map showing extent of late Pleistocene lakes, particularly Lake Russell in Mono Basin, and overflow connections along the eastern Sierra Nevada.

gap drains north toward Mud Spring Canyon (Fig. 2), a tributary of the East Walker River.

Geologic Setting

Mono Basin (Fig. 2) is a southwest-tilted half graben that probably began forming at ca. 4–3 Ma (Gilbert et al., 1968; Huber, 1981). Most of the subsidence is due to displacement along the Sierra Nevada frontal fault (Bursik and Sieh, 1989). The basin is structurally and topographically closed by a west-trending monocline on the north, coincident with the Bodie Hills, and by the north-dipping flank of the Long Valley caldera on the south. The granitic and metamorphic rocks exposed along the Sierra Nevada on the west likely also form the basement complex beneath Mono Basin. These rocks are overlain by a thick accumulation of Miocene to Holocene volcanic rocks

and interbedded alluvial and lacustrine sediments (Gilbert et al., 1968).

Of particular interest to our study are the Pliocene and Pleistocene volcanic flows that form the eastern margin of the basin. The area of the Adobe Hills spillway (Fig. 2) is underlain by voluminous flows of olivine basalt locally interbedded with lacustrine and fluvial deposits. These diatomaceous sands and silts contain abundant fossils of mollusks and fish. On the basis of several K-Ar ages, the basalt flows were emplaced between ca. 4 and 3 Ma (Gilbert et al., 1968; Robinson and Kistler, 1986). The volcanic flows are extensively cut by north-striking normal faults and intersecting northeast-striking, presumably left-lateral faults (Fig. 3). This configuration has produced a zone of extension and net down-to-the-south displacement described by Gilbert et al. (1968) as a structural “knee” or pull-apart

zone. The concentration of microseismicity in this zone (dePolo et al., 1993) shows that extension is ongoing.

Andesitic flows of late Pliocene and Pleistocene age cover much of the northeastern margin of the Mono Basin (Fig. 2) in the vicinity of the Mount Hicks spillway (Gilbert et al., 1968; Robinson and Kistler, 1986; Lange et al., 1993). These flows are cut locally by north- and northeast-striking faults that are less abundant, shorter, and apparently of less displacement than those in the area of the Adobe Hills spillway (Stewart, 1982).

Previous Work

Lake Russell/Mono Lake has a well-known late Quaternary history established by Russell (1889), Putnam (1949), Lajoie (1968), and Stine (1987, 1990). About 13 000 yr ago the lake attained its highest level of late-glacial and postglacial time (Lajoie et al., 1982; Benson et al., 1990). A conspicuous strandline encircling the basin marks this highstand at an altitude of 2155 m (Fig. 2), though at places along the Sierran front it has been downfaulted by as much as 27 m. This shoreline lies 35 m below the Adobe Hills spillway, indicating that the lake has not overflowed since before 13 000 yr B.P.

Discontinuous but conspicuous remnants of a higher shoreline terrace lying at and below an altitude of 2188 m are found at many places around the basin (Figs. 2 and 4A). Such terrace remnants extend into the Adobe Hills spillway; the slightly higher apparent altitude of the spillway (~2189–2190 m) is due either to inaccurate contours in the topographic map or to displacement by faults near the spillway (Fig. 3). Putnam (1949), who first noted this terrace and its associated spillway, thought that it dated from the time of Tahoe glaciation, because he found the terrace on the Tahoe moraines of June Lake (at Aeolian Buttes, Fig. 2). He recognized that Lake Russell overflowed through a series of shallow basins (Fig. 3) into Pleistocene Lake Adobe in Adobe Valley. Lake Adobe, in turn, overflowed into Benton Valley (Fig. 2), a northeastern arm of Owens Valley, by way of a narrow slot cut into granitic rocks of the Benton Range (Fig. 4C; Hubbs and Miller, 1948). The late Pleistocene highstand of Lake Adobe lay at an altitude of ~1975 m (Fig. 2; Snyder et al., 1964; Batchelder, 1970).

Previous information on fluctuations of Lake Russell that predate the conspicuous shorelines at 2155 m and 2188 m is scant. A driller's log from a wildcat well drilled on Paoha Island in 1908 (Scholl et al., 1967) re-

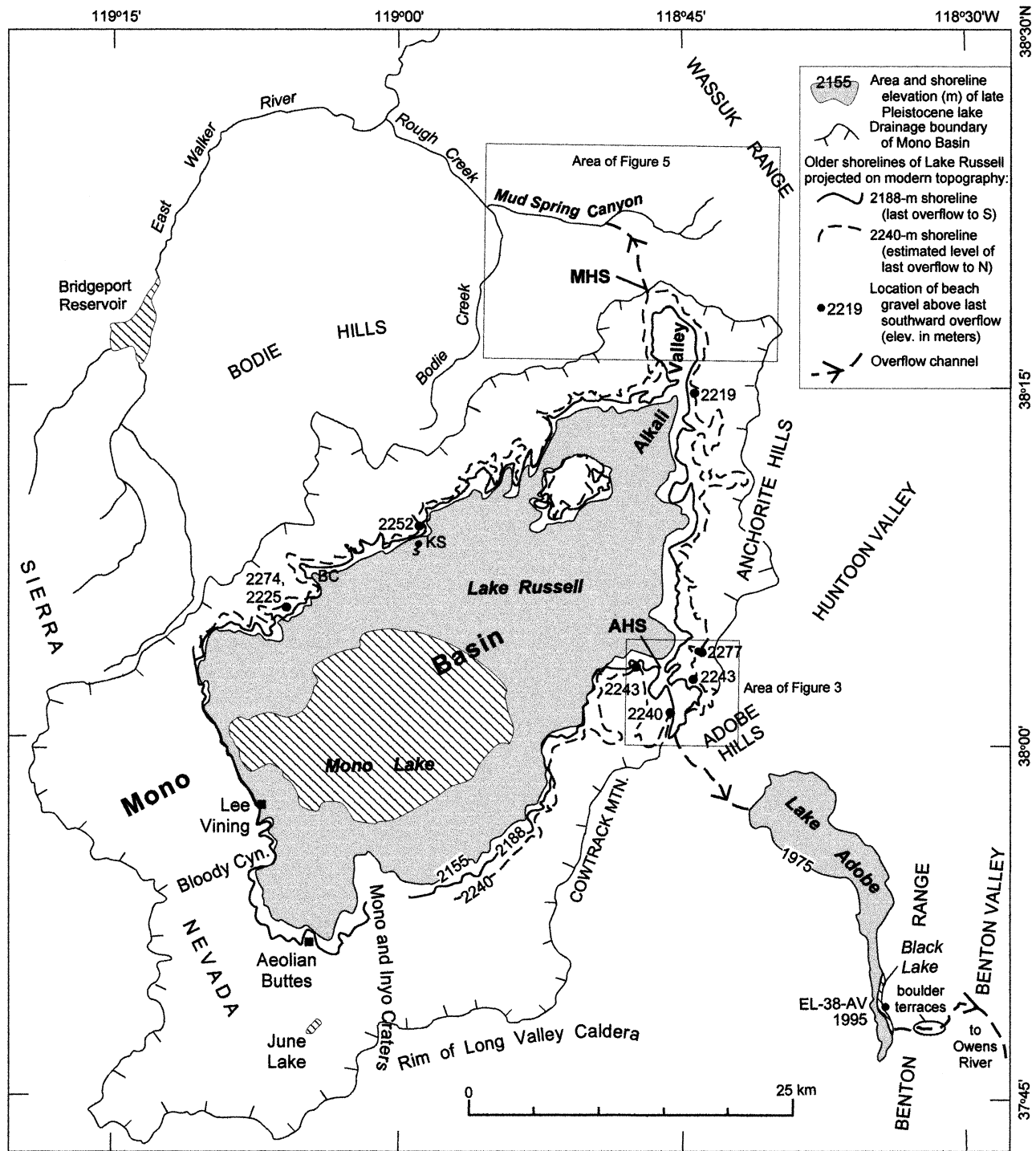


Figure 2. Map showing geography of Mono Basin and surrounding area, shorelines of Lake Russell and Lake Adobe, and locations of lacustrine deposits higher than the last overflow altitudes of these lakes. Diagonal lines are modern lakes. Light dashed lines show areas of Figures 3 and 5. KS—Kirkwood Spring, BC—mouth of Bridgeport Canyon, AHS—Adobe Hills spillway, MHS—Mount Hicks spillway.

records ~600 m of apparently lacustrine sediment that contains a thick layer of light pink rock and pink sand at depths between 411 and 451 m. Lajoie (1968) interpreted this pink ma-

terial to be the 760 ka Bishop Tuff (age from Sarna-Wojcicki et al., 1991). When he applied the sedimentation rate of material overlying the presumed Bishop Tuff (~0.6 m/k.y.) to the

sediments underlying it, Lajoie concluded that a lake may have occupied Mono Basin continuously for at least the past million years.

Russell (1889) hinted, and Hubbs and Mill-

er (1948) proposed, that there had been an earlier connection between the Lahontan and Mono Basins via the East Walker River and Alkali Valley by way of the gap here called the Mount Hicks spillway. Taylor (1985) and Hershler and Sada (2002) inferred a similar connection from the distribution of freshwater mollusks. Miller and Smith (1981) used fossils and data on modern distributions of *Chasmistes* sp. (a sucker) in the western Great Basin to argue that populations of this fish ranged south from the Lahontan system into Mono Basin during the Pliocene (no fish currently live in Mono Lake); they also reported *Chasmistes* fossils in lower Pleistocene deposits near China Lake (Fig. 2).

Dating Methods

Stratigraphic sections within the study area locally contain tephra layers; the U.S. Geological Survey Tephrochronology Laboratory has identified these through chemical correlation with tephra samples of known age. Glass shards from tephra layers were analyzed by electron microprobe for major-oxide composition by using methods of Sarna-Wojcicki et al. (1984) and Sarna-Wojcicki (2000). Sample compositions were compared with compositions in a database of previously analyzed tephra layers by using statistical programs (Table 1). Values for glass compositions in Table 1 are normalized to 100% to correct for the variable amounts of hydration of the volcanic glass (original analytical data are in Appendix Table A1). In addition to hydration, some postdepositional alkali exchange has occurred in many of the samples. Correlations were based on similarity in chemical composition as well as petrographic characteristics (such as shard morphology and phenocryst mineral assemblage) and stratigraphic data.

Radiometric dating of the Quaternary basalt flows includes whole-rock K-Ar ages for this study (Table 2) and from previous work (summarized by Robinson and Kistler, 1986, and Lange et al., 1993), as well as ⁴⁰Ar/³⁹Ar ages (Table 3) provided by the New Mexico Geochronology Research Laboratory. Laboratory analytical procedures and methods of age calculation are described in GSA Data Repository Table DR1.¹ Following recommendations in the analytical report (W.C. McIntosh, 2000, written commun.), we use the inverse isochron ages here.

¹GSA Data Repository item 2002102, analytical data and descriptions of stratigraphic units, is available on the Web at <http://www.geosociety.org/pubs/ft2002.htm>. Requests may also be sent to editing@geosociety.org.

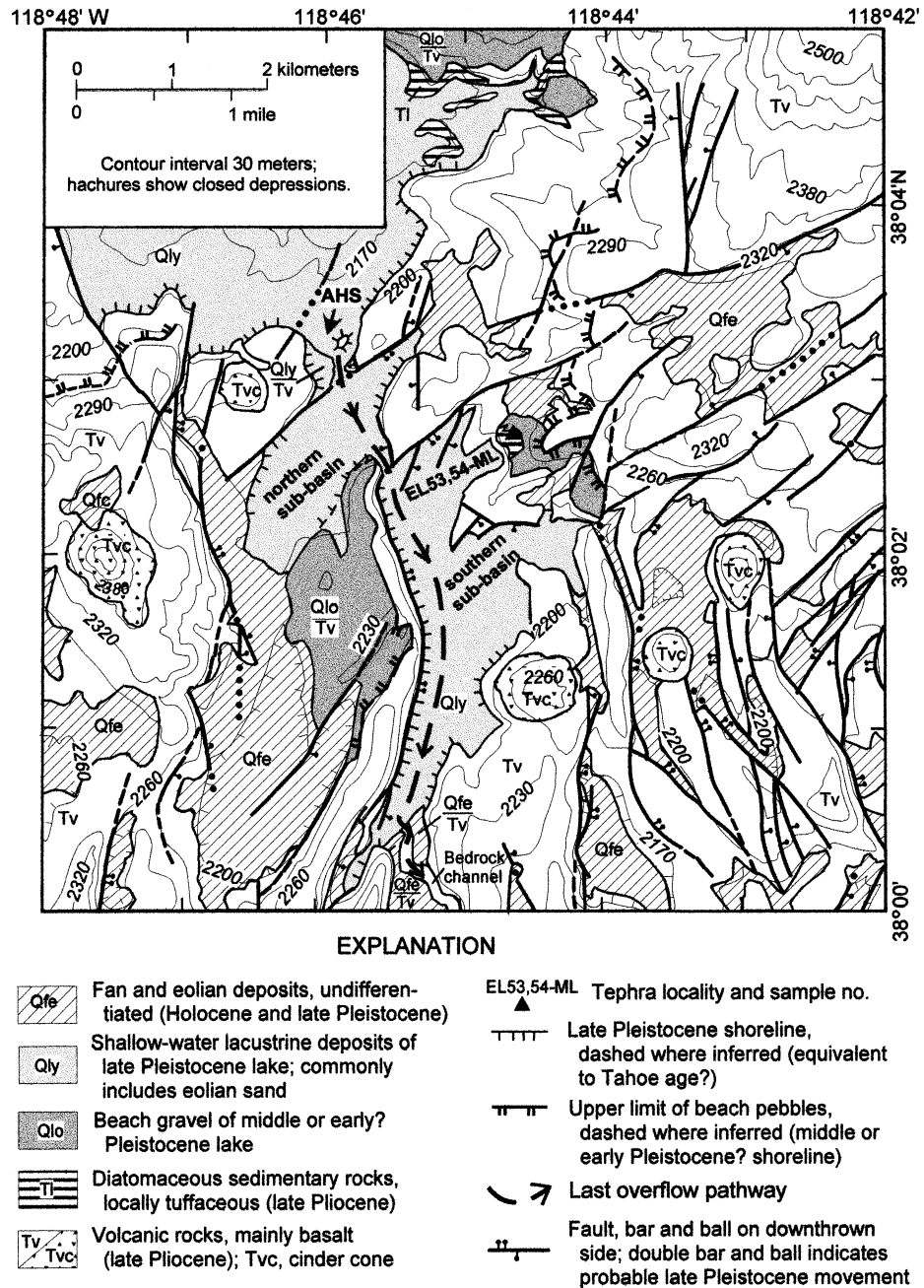


Figure 3. Geologic and topographic map of the area of Adobe Hills spillway (AHS) and drainage channel. Dual map units (e.g., Qly/Tv) indicate thin surficial deposit overlying older unit.

EVIDENCE FOR YOUNGER SOUTHWARD OVERFLOW INTO ADOBE VALLEY

Mapping conducted in the area of the Adobe Hills spillway (Fig. 2) shows that Lake Russell last overflowed through a series of small, shallow, fault-controlled subbasins (Figs. 3 and 4A). These subbasins developed as pull-apart structures at the intersections of

north- and northeast-striking faults (Gilbert et al., 1968). The apparent displacement of Pliocene basalt along most of the faults is down to the east and southeast, effectively lowering the rim of the Mono Basin near the Adobe Hills spillway. Owing to areally extensive and locally thick blankets of Holocene ejecta from the Mono and Inyo Craters to the southwest (Fig. 2), scarps associated with these faults are often obscured. The fresh appearance of some

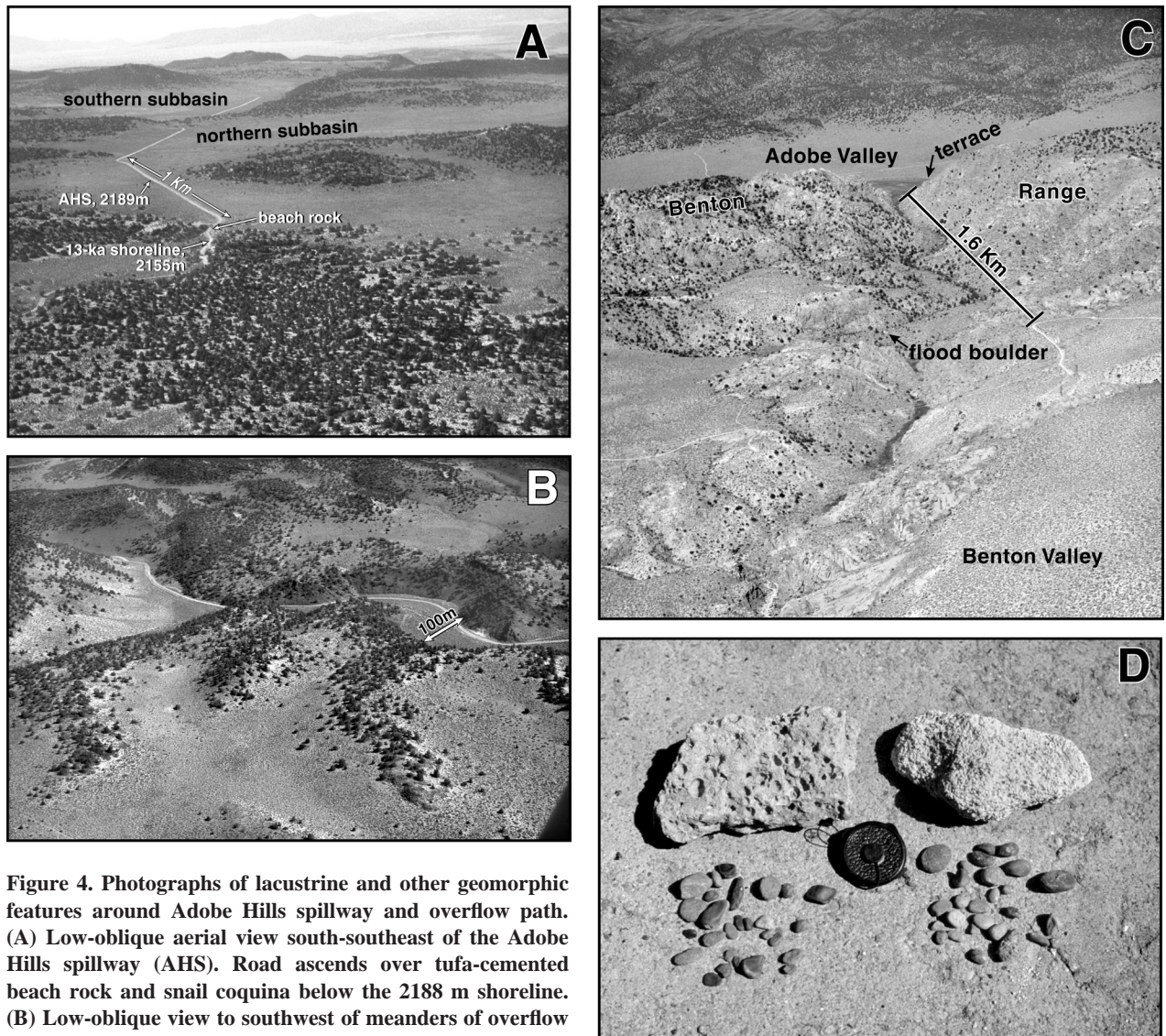


Figure 4. Photographs of lacustrine and other geomorphic features around Adobe Hills spillway and overflow path. (A) Low-oblique aerial view south-southeast of the Adobe Hills spillway (AHS). Road ascends over tufa-cemented beach rock and snail coquina below the 2188 m shoreline. (B) Low-oblique view to southwest of meanders of overflow channel incised into basalt, taken near position marked “bedrock channel” at bottom of Figure 3. (C) Low-oblique aerial view west up the channel cut through the Benton Range; Adobe Valley in the distance. Large granite boulder (at tip of arrow) lies on lower flood terrace. (D) Comparison of beach pebbles and snail coquina (left side) associated with 2188 m shoreline with identical materials (right side) from an altitude of ~2243 m east of Adobe Hills spillway; 50 mm lens cap for scale. Ovoid pits in clasts at left are imprints of snail shells, commonly lined with shell material. White spots on clast at right are smaller snail shells.

of the scarps and the concentration of micro-seismicity in the area of the Adobe Hills spillway (dePolo et al., 1993) indicate that some of these faults remain active.

Character of the Adobe Hills Spillway and Overflow Channel

The Adobe Hills spillway consists of a 300-m-wide gap, <8 m deep at its deepest point, cut into a sill composed of Pliocene basalt (AHS in Figs. 3 and 4A). A broad bench (shown as Qly/Tv in Fig. 3) that slopes gently upward from 2195 to 2200 m in altitude forms

most of the southwest side of the spillway; this bench is abundantly littered with very well rounded pebbles interpreted as beach pebbles.² Inset below this bench is the lowest part of the spillway at ~2190 m altitude. Im-

²Here and throughout the paper, “beach pebbles” or “beach gravel” refers to clasts that are very well rounded and usually polished and commonly have flattened discoidal or spindle shapes that are produced by wave erosion at shorelines of lakes or oceans. The beach pebbles in the Adobe Hills area and in many other parts of the Mono Basin are also distinguished by an abundance of polished black, red, and clear obsidian.

mediately southeast of the 75-m-wide spillway are two small basins, each defined by a bedrock sill, through which the waters of Lake Russell spilled during episodes of overflow. Hydrographically closed prior to the overflow events, these basins became connected when overflow breached their confining sills. The southern lip of the southern basin in this overflow chain is breached by a 100-m-wide channel incised 12–15 m deep into basalt (Figs. 3 and 4B); this meandering channel conducted spillage southward into Adobe Valley, where it fed Lake Adobe (Fig. 2). The upper end of this channel is currently at 2173 m, well be-

TABLE 1. LIST OF COMPOSITIONALLY CLOSEST MATCHES TO TEPHRA LAYERS

No.	Field sample no.	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MgO	MnO	CaO	TiO ₂	Na ₂ O	K ₂ O	Sim. coeff.	Comments	
1	EL-38-AV	76.7	13.2	0.76	0.03	0.05	0.44	0.07	3.81	4.93	1.0000	Tephra layer exposed on hillslope below, and presumably stratigraphically beneath, beach gravel of Lake Adobe on the west side of Adobe Valley (Fig. 2). Tephra lies within diatomaceous, freshwater marsh deposits. Closest matches are to Mono Glass Mountain ash beds, ca. 0.9 to ca. 1.2 Ma (2, 3, 9, 15), and to the Bishop Tuff and ash bed, ca. 0.76 Ma (4, 5, 7, 16, and 19).	
2	TECO-30B-2	77.9	13.1	0.74	0.03	0.04	0.44	0.07	3.75	4.98	0.9890		
3	OL92-1030	76.9	12.9	0.74	0.04	0.03	0.44	0.07	3.90	4.93	0.9872		
4	TECO-28B	77.0	12.9	0.75	0.04	0.03	0.43	0.06	3.80	5.00	0.9863		
5	BT-14	77.2	12.8	0.75	0.03	0.04	0.43	0.06	3.77	4.92	0.9851		
7	TECO-10	76.8	13.1	0.78	0.04	0.03	0.42	0.08	3.78	5.01	0.9822		
9	TECO-29	76.7	13.1	0.81	0.04	0.04	0.43	0.07	3.85	4.96	0.9818		
15	BT-15	77.2	12.8	0.77	0.04	0.03	0.46	0.06	3.85	4.85	0.9793		
16	YJC-1-87	77.3	12.9	0.73	0.04	0.05	0.44	0.07	3.74	4.75	0.9790		
19	BT-11D2	77.2	13.1	0.76	0.02	0.04	0.42	0.05	3.73	4.66	0.9775		
1	EL-102-MS	74.4	13.8	2.27	0.12	0.05	0.94	0.17	4.34	3.86	1.0000		Tephra layer exposed in small prospect cut in fluvial deposits of unit Ts on north side of Mud Spring Canyon (Fig. 5). Closest matches with the Lawlor Tuff, ca. 4.1 Ma, and next best with the Huichica Tuff, ca. 4.0 Ma (Sarna-Wojcicki et al., 1991), both erupted from the Sonoma volcanic field in the Coast Ranges of California. The Lawlor Tuff is found in the San Joaquin Valley, the Los Angeles Basin, the Mojave Desert, and the Alturas area of California. There, it underlies the "Alturas ash," which correlates with the Kilgore Tuff, dated ca. 4.5 Ma by Morgan and McIntosh (2002).
2	758-218	73.7	14.4	2.24	0.10	0.04	0.90	0.15	4.30	4.12	0.9703		
3	KT-6	74.3	13.6	2.24	0.09	0.05	0.89	0.18	4.61	4.05	0.9683		
4	ALT-11	73.5	14.5	2.17	0.10	0.03	0.95	0.16	4.33	4.16	0.9681		
8	JCD-(1)	73.6	14.3	2.25	0.08	0.03	1.06	0.17	4.39	4.04	0.9628		
9	MP-8	74.0	14.1	2.19	0.09	0.03	0.86	0.16	4.33	4.19	0.9624		
10	KT-7(2)	73.6	14.3	2.16	0.09	0.04	0.91	0.18	4.69	3.98	0.9618		
11	WC-B3B	73.8	13.9	2.17	0.09	0.07	0.93	0.17	4.30	4.55	0.9614		
12	KT-7(1)	74.0	13.5	2.22	0.10	0.04	0.94	0.17	4.77	4.25	0.9611		
14	H-1	73.6	14.6	2.29	0.09	0.06	1.05	0.19	4.44	3.70	0.9594		
1	EL-103-MS	77.1	12.8	0.61	0.05	0.05	0.50	0.09	2.72	6.00	1.0000	Tephra layer in pumiceous sand and gravel of unit Tg, buried by flow Qb5 on south side of Mud Spring Canyon. Closest matches with late Miocene tephra at ca. 11.5 Ma (2), the age of which is interpolated from position between beds correlated to beds of known age (Stewart et al., 1999). Also similar to a closely related group of tephra from Willow Wash in southern Fish Lake Valley (3, 6, 7, 9, 18) and to a nearby tephra, EL-35A-MS, in similar sediments (20; see Fig. 7). The proximity of these beds and their stratigraphic relationships support a preferred correlation to the Fish Lake Valley tephra group, whose age range is >2.58 Ma, <2.81 Ma (Reheis et al., 1991, 2002).	
2	2379-57B	77.0	13.0	0.62	0.08	0.07	0.54	0.10	2.68	5.98	0.9799		
3	FLV-25-WW	77.6	12.4	0.61	0.07	0.07	0.56	0.10	2.89	5.66	0.9567		
5	SP-98-1	77.2	12.7	0.58	0.05	0.05	0.42	0.10	2.83	6.06	0.9556		
6	FLV-33-WW	76.9	12.9	0.64	0.07	0.07	0.62	0.12	2.69	5.96	0.9552		
7	FLV-121-WW	76.6	12.9	0.65	0.04	0.06	0.57	0.07	2.85	6.20	0.9539		
9	FLV-32-WW	77.2	12.9	0.61	0.07	0.08	0.62	0.12	2.80	5.65	0.9527		
18	FLV-34-WW	76.8	12.9	0.62	0.08	0.06	0.62	0.10	2.52	6.22	0.9453		
20	EL-35A-MS	77.4	12.6	0.63	0.05	0.05	0.53	0.06	2.25	6.37	0.9429		
1	EL-104-MS	76.8	13.0	1.14	0.03	0.06	0.54	0.05	3.74	4.66	1.0000		Tephra layer within colluvium or landslide deposits burying the flank of flow Qb5 on south side of Mud Spring Canyon. Closest matches are to a Mono Craters tephra layer (2) from a core in Walker Lake, with an interpolated age from ¹⁴ C dates of 1190 yr B.P. (Sarna-Wojcicki et al., 1988). Other matches are with tephra samples from Walker Lake cores that range from early to late Holocene (3, 5, 11), and to the surface pumice from Panum Crater (4). Also similar to a tephra from Fish Lake Valley (30) dated by associated charcoal as 1,620 ± 60 ¹⁴ C yr B.P. Thus, tephra EL-104-MS is probably from the Mono Craters and is most likely late Holocene in age.
2	WL-8-2A (28.5–33.0 cm)	77.0	12.8	1.12	0.03	0.05	0.54	0.06	3.76	4.66	0.9928		
3	WL-4-4 (10.83 m)	76.9	12.9	1.10	0.04	0.06	0.54	0.06	3.77	4.64	0.9912		
4	KRL82282	76.7	12.9	1.13	0.02	0.06	0.54	0.06	3.87	4.68	0.9909		
5	WL-4-4B (12.00 m)	76.9	12.8	1.13	0.03	0.05	0.55	0.05	3.79	4.64	0.9898		
8	WL 4-2 (3.31 m)	76.7	12.9	1.14	0.04	0.04	0.55	0.06	3.92	4.67	0.9876		
11	WL 8-1B (92–94 cm)	77.0	12.7	1.11	0.02	0.06	0.54	0.07	3.84	4.65	0.9869		
30	FLV-164-LC	77.2	12.6	1.10	0.03	0.06	0.55	0.06	3.76	4.61	0.9831		
1	EL-105-MS	77.1	13.0	0.63	0.06	0.06	0.62	0.09	3.05	5.41	1.0000	Tephra layer from north of Alkali Valley. Poorly exposed; pumiceous deposits appear to overlie Miocene andesite and underlie a terrace formed by discharge from Lake Russell through the Mount Hicks spillway (unit Qg, Fig. 5). Closest matches are to a related group of tephra from Willow Wash in southern Fish Lake Valley (2, 3, 6, 12) whose age range is >2.58 Ma, <2.81 Ma (Reheis et al., 1991; Sarna-Wojcicki et al., 1999, written commun.), as well as to older Miocene tephra from Horsethief Canyon in northern Eureka Valley (4, 5) that are between 5.9 and 5.3 Ma (Reheis and Sawyer, 1997).	
2	FLV-24-WW	76.7	13.0	0.67	0.07	0.07	0.63	0.08	3.18	5.58	0.9741		
3	FLV-32-WW	77.2	12.9	0.61	0.07	0.08	0.62	0.12	2.80	5.65	0.9725		
4	FLV-85-HT	76.7	13.1	0.72	0.06	0.05	0.62	0.08	3.11	5.56	0.9690		
5	FLV-86-HT	76.6	13.1	0.72	0.06	0.05	0.60	0.07	3.06	5.71	0.9619		
6	FLV-33-WW	76.9	12.9	0.64	0.07	0.07	0.62	0.12	2.69	5.96	0.9614		
12	FLV-25-WW	77.6	12.4	0.61	0.07	0.07	0.56	0.10	2.89	5.66	0.9546		

Note: Analyzed in this study (localities in Figs. 2 and 5), based on electron-microprobe analysis of volcanic glass shards. Shards were analyzed by using the JEOL8900*instrument. Total (R) = total recalculated to a 100% fluid-free basis. Analyses were performed by C.E. Meyer, U.S. Geological Survey, Menlo Park, California, during 1996–1999. Identifications of closest matches are based on calculations of similarity coefficients (Sim. coeff.) by using oxides of the following elements: Na, Al, Si, K, Ca, and Fe. Closely matching samples that do not provide age constraints or stratigraphic context were omitted from the table. Initial oxide concentrations and totals for the samples analyzed in this study are given in Appendix Table A-1.

*Use of trade names is for informational purposes only and does not imply endorsement of product by the U.S. Geological Survey.

TABLE 2. ANALYTICAL DATA FOR K-Ar AGE DETERMINATION

Sample number	Map unit*	Material dated	K ₂ O (wt%)	⁴⁰ Ar (× 10 ⁻¹¹ mol/g)	⁴⁰ Ar*/Σ ⁴⁰ Ar (%)	Age (Ma ± σ)
EL-KA-2	Qa	Whole rock	3.49	2.3424	23.3	1.44 ± 0.06

Note: E. McKee, U.S. Geological Survey, 1995, written commun.; analysis by U.S. Geological Survey Menlo Park laboratory.

*See Figure 5 for map unit and sample location.

low the 2189–2190 m level at the spillway. Whether this incision was accomplished by spillage from just one or several overflow events is unknown. At the lower end of the canyon in Adobe Valley, young fan deposits have buried any possible evidence of deltaic deposits.

TABLE 3. $^{40}\text{Ar}/^{39}\text{Ar}$ ANALYTICAL DATA FOR GROUNDMASS CONCENTRATES OF BASALT SAMPLES FROM MONO BASIN

ID	Temp (°C)	$^{40}\text{Ar}/^{39}\text{Ar}$	$^{37}\text{Ar}/^{39}\text{Ar}$	^{36}Ar $^{39}\text{Ar}(\times 10^{-3})$	$^{39}\text{Ar}_k$ $(\times 10^{-16} \text{ mol})$	K/Ca	$^{40}\text{Ar}^*$ (%)	^{39}Ar (%)	Age (Ma)	$\pm 1\sigma$ (Ma)	
EL-KA-6, 136.99 mg groundmass concentrate, $J = 0.000456 \pm 0.10\%$, laboratory number = 50643-01											
A	◇	625	5731.	0.2448	19385	1.04	2.1	0.1	0.1	2	1794
B	□	700	56.4	0.5729	193	85.2	0.89	-1	6.1	-0.49	0.32
C	◇	750	28.89	0.425	98.21	24.8	1.2	-0.3	7.8	-0.08	0.65
D	◇	800	19.79	0.6448	67.2	215.6	0.79	-0.1	23	-0.01	0.1
E	◇	875	19.05	0.5796	64.94	272.1	0.88	-0.5	42.1	-0.07	0.08
F	◇	975	24.14	0.6648	81.65	296.9	0.77	0.3	63	0.06	0.1
G	◇	1075	45.6	0.8384	154.1	122	0.61	0.3	71.5	0.11	0.22
H	□	1250	65.	1.123	220	258	0.45	0.2	89.7	0.08	0.22
I	□	1650	54.25	2.29	180	146.9	0.22	2.3	100	1.04	0.24
Total gas age			$n = 9$		1422.6	2.01			0.1	3.0*	
Plateau			MSWD = 0.6 $n = 7$		St. B-H	1274.6	0.75		89.6	-0.02	0.12*
Isochron			MSWD = 0.7 $n = 7$		St. A-G				$^{40}\text{Ar}/^{36}\text{Ar} = 295 \pm 5^*$	0.04	0.04*
EL-KA-7, 148.34 mg groundmass concentrate, $J = 0.000456 \pm 0.10\%$, laboratory number = 50621-01											
A	◇	625	2359	1.096	7911	5.47	0.47	0.9	0.2	17	135
B	□	700	78.03	1.09	254.2	55.7	0.47	3.9	2	2.48	0.56
C	◇	750	38.65	1.014	118.2	8.19	0.5	9.8	2.2	3.1	2
D	◇	800	7.217	0.8551	17.86	297.2	0.6	27.8	11.7	1.65	0.05
E	◇	875	3.614	0.6117	5.617	452.6	0.83	55.5	26.2	1.65	0.03
F	◇	975	2.896	0.4215	3.296	903.8	1.2	67.6	55.2	1.61	0.01
G	◇	1075	3.225	0.3218	4.456	694.7	1.6	60	77.4	1.59	0.02
H	◇	1250	10.27	0.6668	28.23	596.2	0.77	19.3	96.5	1.63	0.04
I	◇	1650	37.61	2.599	120.4	107.8	0.2	6	100	1.86	0.2
Total gas age			$n = 9$		3121.7	0.5			1.7	0.5*	
Plateau			MSWD = 1.2 $n = 8$		St. B-I	3116.3	0.5		99.8	1.62	0.02*
Isochron			MSWD = 0.7 $n = 9$		St. A-I				$^{40}\text{Ar}/^{36}\text{Ar} = 298 \pm 2^*$	1.6	0.02*
EL-KA-9, 148.06 mg groundmass concentrate, $J = 0.000456 \pm 0.10\%$, laboratory number = 50644-01											
A	◇	625	412.4	0.2955	1383	16.9	1.7	0.9	0.6	3.2	7.9
B	◇	700	18.33	0.4618	58.55	139.6	1.1	5.8	5.7	0.88	0.13
C	◇	750	12.11	0.4243	37.01	39.7	1.2	10	7.1	0.99	0.32
D	◇	800	7.441	0.6793	20.69	350.3	0.75	18.6	19.8	1.14	0.04
E	◇	875	5.312	0.585	12.63	402.5	0.87	30.7	34.5	1.34	0.03
F	◇	975	3.911	0.4516	8.028	771.2	1.1	40.3	62.5	1.3	0.02
G	◇	1075	4.667	0.5438	10.78	456.9	0.94	32.7	79	1.26	0.03
H	◇	1250	11.95	1.221	35.54	490.1	0.42	13	96.8	1.27	0.05
I	□	1650	14.58	2.534	43.34	87.2	0.2	13.6	100	1.64	0.16
Total gas age			$n = 9$		2754.5	1.59			1.3	0.2*	
Plateau			MSWD = 2.1 $n = 5$		St. E-I	2207.9	0.49		80.2	1.29	0.04*
Isochron			MSWD = 3.7 [§] $n = 8$		St. A-H				$^{40}\text{Ar}/^{36}\text{Ar} = 291 \pm 6^*$	1.32	0.08*

Note: Ages provided by New Mexico Geochronology Research Laboratory. Isotopic ratios corrected for blank, radioactive decay, and mass discrimination; not corrected for interfering reactions. Individual analyses show analytical error only; plateau and total gas age errors include error in J and irradiation parameters. Analyses in italics are excluded from final age calculations. Discrimination = 1.00468 ± 0.00093 a.m.u.; n = number of heating steps, □ = steps included in plateau calculation, ◇ = steps included in isochron, K/Ca = molar ratio calculated from reactor-produced $^{39}\text{Ar}_k$ and $^{37}\text{Ar}_{ca}$. MSWD = mean square of weighted deviations.

*2 σ error.

[†]Steps included in plateau calculation.

[§]MSWD outside of 95% confidence interval.

Abundant beach pebbles lie as much as 10 m above the lip of the Adobe Hills spillway (the highest level of these is shown by the light-hachured line in Fig. 3). The highest altitude of abundant beach pebbles around the northern small basin and within the Adobe Hills spillway is ~2200 m (7220 ft), whereas the highest altitude in the southern small basin is 2191 m (7190 ft). This difference in altitude could have been produced by one or some combination of the following: (1) south-down displacement of the southern basin by ~9 m along the faults separating it from the northern small basin and (2) progressive filling and spilling of successively lower small basins during overflow through the Adobe Hills spill-

way. Displacements of the basalt flows by some faults suggest long-term slip rates of at least 1 cm/k.y., which could easily account for part or all of the difference in altitude.

Littoral deposits that lie immediately lake-ward (north) of the Adobe Hills spillway (Fig. 4A) are composed locally of tufa-cemented beach rock, in places made up of a snail coquina (Fig. 4D) dominated by *Carinifex newberryi* [Lea] and *Vorticifex gesteri* [Hanna] (identification by Peter Rodda, California Academy of Sciences). A radiocarbon age on this tufa is $29000 \pm 1000/_{-800}^{+1000} \text{ }^{14}\text{C}$ yr B.P. (catalog no. LDGO 1722e); we consider this to be a minimum age for overflow through the spillway because of the potential for postdeposi-

tional contamination caused by porosity of the material. Uranium-series dating of the beach rock and the snails proved unsuccessful because of high thorium content (W. Broecker, Lamont-Doherty Institute, 1990, personal commun. to Stine). ^{36}Cl exposure dating on wave-rounded basalt and tufa below the 2188 m shoreline from ~5 km west of the spillway (F. Phillips, New Mexico Tech., 1999, oral commun.) yielded preliminary minimum ages of 50–80 ka. These ages are consistent with the inference that the 2188 m highstand must be Tahoe or younger in age because shorelines at this altitude are etched on Tahoe moraines near June Lake (Fig. 2). However, relative weathering data do not provide a firm correlation of the Tahoe moraines of June Lake to younger or older Tahoe advances (Bursik and Gillespie, 1993). To the north at Bloody Canyon, younger Tahoe I moraines were dated as 66 ka to older than 56 ka by ^{36}Cl (Phillips et al., 1990); the morphologically older Tahoe II and Mono Basin moraines were dated as older than 100 ka to as old as ca. 200 ka. Farther south along the Sierra Nevada, dated basalt flows interbedded with moraines provide limiting ages of older than 53 ka and younger than 119 ka for younger Tahoe and older than 131 ka for older Tahoe (Gillespie, 1982, 1984).

Lake Adobe and Its Overflow

The southward spillage from Lake Russell flowed into Adobe Valley. This small basin, which today holds Black Lake, was during Pleistocene time the site of Lake Adobe. Black Lake is not dependent on spills from Mono Basin to expand during wetter climatic periods because it is fed by runoff from Cow-track Mountain and the north flank of the Long Valley caldera. Its last highstand at ~1975 m (Fig. 2; Snyder et al., 1964; Batchelder, 1970), just 10 m higher than the present lake, was probably attained only through local runoff. We identified an eroded lake terrace with abundant beach pebbles and sand at an altitude of ~1995 m on the east side of Adobe Valley. These deposits are >2 m thick and rest unconformably on a cemented layer, probably a duripan horizon of a paleosol, that caps diatomaceous sand and silt containing a tephra layer (EL-38-AV, Fig. 2 and Table 1). The diatoms represent shallow, fresh-water marsh environments (J.P. Bradbury, U.S. Geological Survey, 1996, written commun.). The tephra is correlated with the Bishop Tuff (760 ka) or the ca. 1 Ma Glass Mountain tephra layers (Reheis et al., 2002). On the basis of the unconformity, we think that the diatomaceous

sediments are significantly older than the overlying beach deposits; these deposits at 1995 m altitude indicate that Lake Adobe once stood ~20 m above its youngest shoreline.

A perched, stream-cut terrace within the bedrock notch through which Adobe Valley drains into Benton Valley (Fig. 4C) lies at the same height (30 m) as the older beach gravel above Black Lake. Within and at the mouth of a precipitous granite gorge below the notch, remnants of two terraces ~43 m and 30 m above the modern drainage are littered with giant granite boulders, as much as 5 m in diameter, that clearly were eroded from within the gorge. The older terrace deposit overlies rhyolite mapped as the ca. 2 Ma tuff of Taylor Canyon (Krauskopf and Bateman, 1977); both terraces also are inset within nearby outcrops of the Bishop Tuff. We interpret these bouldery deposits to represent two episodes of high-volume, possibly catastrophic, discharge from Adobe Valley that postdated the eruption of the Bishop Tuff. These episodes may have been caused by discharge from Lake Russell that filled Lake Adobe and caused it to spill over a formerly much higher sill. These discharge episodes likely produced rapid incision of the bedrock gorge, as indicated by the giant flood boulders (Fig. 4C). Such boulders could be easily transported because the gradient is so steep; the present valley floors are only 3.5 km apart but 245 m different in altitude, and the upper part of the gorge is a virtual waterfall with a 17% slope. Active faulting along the valley margins may also have contributed to incision.

EVIDENCE FOR SHORELINES HIGHER THAN 2188 m

Recent discoveries (Fig. 2) include (1) lacustrine deposits in the southeastern corner of Mono Basin representing at least one lake stand that greatly exceeded the 2188 m shoreline, and (2) beach gravels at three sites on the north side of the basin at altitudes between 2215 m and 2268 m. The modern altitudes may be different from the original altitudes owing to postdepositional faulting in this region of active tectonics. However, they should generally represent minimum values of the original altitudes because the extensional regime in this area during the Pleistocene should produce mainly downward displacement. As we lack age data, correlations among shoreline sites affected to different degrees by faulting is clearly speculative.

Description of High Shoreline Deposits

Lacustrine deposits near the Adobe Hills spillway (Fig. 3) overlie basalt and crop out at altitudes between 2205 and 2220 m. The deposits comprise an upwardly coarsening sequence of tuffaceous silt, sand, and beach gravel containing two tephra layers. Correlations based on chemical analyses indicate that these tephra (EL-53-ML and 54-ML) are latest Pliocene in age (2.5–2.0 Ma; Reheis et al., 2002). These lacustrine sediments are mapped as part of unit T1 (Fig. 3), but they presumably overlie the older ca. 3 Ma lacustrine deposits, which consist mostly of diatomaceous beds in the spillway area. In addition, beach pebbles and rare large chunks of tufa-cemented beach rock and snail coquina (unit Q10, Figs. 3 and 4D) are found on slopes at altitudes as high as 2280 m; they document a lake that in past time(s) stood much higher than that represented by the tephra-bearing deposits. Though exposures are poor, we infer that the tuffaceous beds and the higher beach rock and lag pebbles represent at least two different lacustrine episodes that reached ~2220 m and 2280 m, respectively. Several northeast-striking faults progressively displace the 2280 m deposits of beach gravel down to the south (Fig. 3), decreasing their maximum altitude to as low as 2228 m.

Beach pebbles within a similar range of altitudes in areas of less faulting (Dohrenwend, 1983) on the north side of Mono Basin (sites are shown as dots with altitudes in Fig. 2) confirm that Lake Russell once stood at very high levels. At one site southwest of Bridgeport Canyon, a dissected berm at an altitude of ~2212 m forms a tombolo between a small outlier of bedrock and the Bodie Hills; the tombolo slopes upward to a bedrock notch, resembling a shoreface angle, at ~2225 m. Abundant beach gravel crops out below the tombolo and also occurs as lag on bedrock as high as 2274 m. At a second site, north of Kirkwood Spring, beach pebbles and cobble-sized clasts shaped into ovoid rollers are scattered on a gentle bedrock slope to as high as 2252 m. Beach pebbles occur as high as 2219 m in the southeast corner of Alkali Valley. The preservation of geomorphic shoreline features and the similarity of the elevated beach pebbles to those of known highstands strongly suggest that these deposits are Pleistocene in age. We have not made an exhaustive search around Mono Basin, but anticipate that many similar sites exist with evidence for shorelines above the 2188 m altitude of the last overflow.

Tectonically Induced Opening of the Adobe Hills Spillway

The topography of the Mono Basin rim near the Adobe Hills spillway must have been very different when Lake Russell reached altitudes of 2210–2280 m, because no topographic barrier at or near the spillway is high enough to enclose so large a lake (Fig. 3). Numerous, closely spaced faults cutting 3 Ma basalt flows (Gilbert et al., 1968) have evidently lowered the landscape in this area, eventually creating a gap through which Lake Russell could drain south. This inference is consistent with the apparent down-to-the-south displacement of the highest level of beach pebbles east of the Adobe Hills spillway, totaling at least 52 m across several faults.

EVIDENCE FOR OLDER NORTHWARD OVERFLOW OF LAKE RUSSELL

As noted by Russell (1889), a low divide in northeastern Mono Basin lies at an altitude of 2237 m in the northwest corner of Alkali Valley, a northeastern appendage of Mono Basin (Fig. 2). At the time of the ca. 13 ka highstand at 2155 m, Lake Russell inundated the southern part of Alkali Valley but did not reach the higher northeastern part of the valley. Subdued lake terraces in the northern part of the valley below 2180 m probably were deposited during the 2188 m highstand of Lake Russell that did enter the valley. The late Tertiary and Quaternary sedimentary and volcanic stratigraphy (Tables 4 and DR2) of the area between Alkali Valley and Mud Springs Canyon (Figs. 5 and 6), records a history of older northward overflow from Lake Russell. This record can be deduced from the ages of deposits that were incised by paleodrainages carrying lake discharge and subsequently plugged by volcanic flows (Figs. 7 and 8).

Mount Hicks Spillway and Overflow Channel

A broad, gentle ridge ~0.5 km wide forms the lowest point in the northern rim of Alkali Valley (Fig. 7C) at an altitude of ~2237 m. This ridge is composed of andesite flow Qa (Fig. 5). The andesite ridge locally bears patches of subrounded pebble to cobble gravel of various lithologies, including granitic, rhyolitic, and basaltic clasts, suggesting a former drainageway across the ridge. We informally refer to this feature as the Mount Hicks spillway. The andesite flow emanated from Mount Hicks and is similar in age to flow Qb1, also

TABLE 4. AGES AND RELATIONSHIPS OF CRITICAL MAP UNITS IN THE AREA OF MOUNT HICKS SPILLWAY AND MUD SPRING CANYON

Unit	Approximate age*	Description and relationships to other units†
Qb5	<0.08 Ma, late Pleistocene	Elongate flow of hornblende andesite. West end of flow Qb5 emplaced in paleovalley incised into units Tp and Ts; east end buries Qp2 and Tg.
Qb4	0.2–0.5 Ma, middle Pleistocene	Andesite flow erupted from Aurora Crater. Includes older andesite at north edge (Lange and Carmichael, 1996). Buries western end of flow Qb3 and paleovalley occupied by Qb3. Appears inset below surfaces of Qp2 and Qt.
Qp3	Middle(?) Pleistocene	Unconsolidated, granitic pediment gravel. Inset below surface of Qp2; truncates higher Qt terrace.
Qt	Middle(?) to early Pleistocene	Well-washed, bedded fluvial sand and gravel. Lap onto fan/pediment gravel along Mud Spring Canyon west of flow Qb5. Two nested terraces lie 12–20 m above present drainage; higher terrace appears slightly inset below Qp2.
Qp2	Early Pleistocene	Unconsolidated granitic gravel, mapped as pediment gravel (Gilbert and Reynolds, 1973) but locally an aggradational unit. Inset below Qp1 surface.
Qb3	1.3 Ma (early Pleistocene)	Elongate flow of basaltic andesite. Emplaced in west-draining paleovalley incised into units Qp1 and Ts and floored by alluvium.
Qb2	>1.3 Ma, <1.6 Ma (early Pleistocene)	Basaltic andesite erupted from cone on northern edge of flow Qb1. Younger than Qb1; appears older than Qb3, but field relationships are unclear.
Qp1	>1.3 Ma (early Pleistocene to late Pliocene?)	Unconsolidated, granitic pediment gravel. Mainly overlies unit Ts; likely younger than unit Tg, but field relationships are unclear.
Qa	1.4–1.6 Ma (early Pleistocene)	Andesite; forms broad ridge that contains Mount Hicks spillway. Flowed north from spillway area down paleovalley incised into unit Tg and older rocks.
Qb1	1.5–1.7 Ma (early Pleistocene)	Basaltic andesite erupted from Mount Hicks, coeval with flow Qa.
Tg	2.9–2.5 Ma (late Pliocene)	Poorly consolidated, pale tuffaceous alluvium, including debris flows and air-fall pumice layers. Inset within unit Ts and highest pediment Tp.
Tp	>2.9 Ma, <4.8 Ma (Pliocene)	Unconsolidated, granitic pediment gravel. Overlies unit Ts but not younger deposits.
Ts	ca. 4.8 Ma	Upwardly coarsening sequence of diatomite, mudstone, siltstone, sandstone, and pebble conglomerate; indurated, flat to gently dipping (about 10°), and cut locally by small faults. Overlies Miocene andesite.

*See Tables 1–3 for details of age control.

†See Data Repository Table DR-2 (described in footnote 1 in text) for detailed descriptions and Figs. 5 and 6 for locations.

from Mount Hicks: A K-Ar age on unit Qa at the spillway is 1.44 ± 0.06 Ma (EL-KA-2, Table 2; E. McKee, U.S. Geological Survey, 1995, written commun.), overlapping the age of Qb1 (1.6 ± 0.1 Ma (Gilbert et al., 1968); age corrected by using new isotopic ratios of Robinson and Kistler, 1986) within 2σ error limits.

A small valley ~40 m deep drains north from the spillway. This valley narrows from 200 m wide at its head to only ~25 m wide near its north end, where it is incised into Miocene and older rocks as well as tuffaceous alluvium of unit Tg of late Pliocene age (Tables 1 and 4). Tightly paralleling the valley on the west side is the same andesite as that underlying the spillway ridge (Figs. 5 and 7C). A sample from this down-valley part of flow Qa was dated by $^{40}\text{Ar}/^{39}\text{Ar}$ at 1.6 ± 0.02 Ma (EL-KA-7, Table 3). This flow is at least 20 m thick and has a gently sloping top with a gradient like that of the modern valley floor. The shape of flow Qa and its relationship to the present valley clearly indicate that the an-

desite flowed down a paleovalley that resembled the present valley in shape, size, and orientation. The paleovalley must postdate 2.5 Ma, the youngest tephra identified in unit Tg. For ease of reference, we term the andesite-filled paleovalley “Hicks Valley 1” (HV1) and the present valley “Hicks Valley 2” (HV2, Fig. 8).

Hicks Valley debouches onto a broad alluvial surface that is ponded on the north and west by flows Qb5 and Qb3. A small terrace, beveled across both unit Tg (at tephra locality 105-MS, Fig. 5 and Table 1) and Miocene andesite, stands ~10 m above the eastern valley floor just south of the debouchment. The terrace top and flank are littered with very well rounded, polished pebbles and cobbles (identical to those already described herein as “beach gravel”) of a variety of volcanic types (Fig. 7D). Less well-rounded boulders derived from Miocene andesite and granitic rocks could have been transported from nearby sources (unit R, Fig. 5). Clasts in modern alluvium, in older fan deposits, and in Miocene

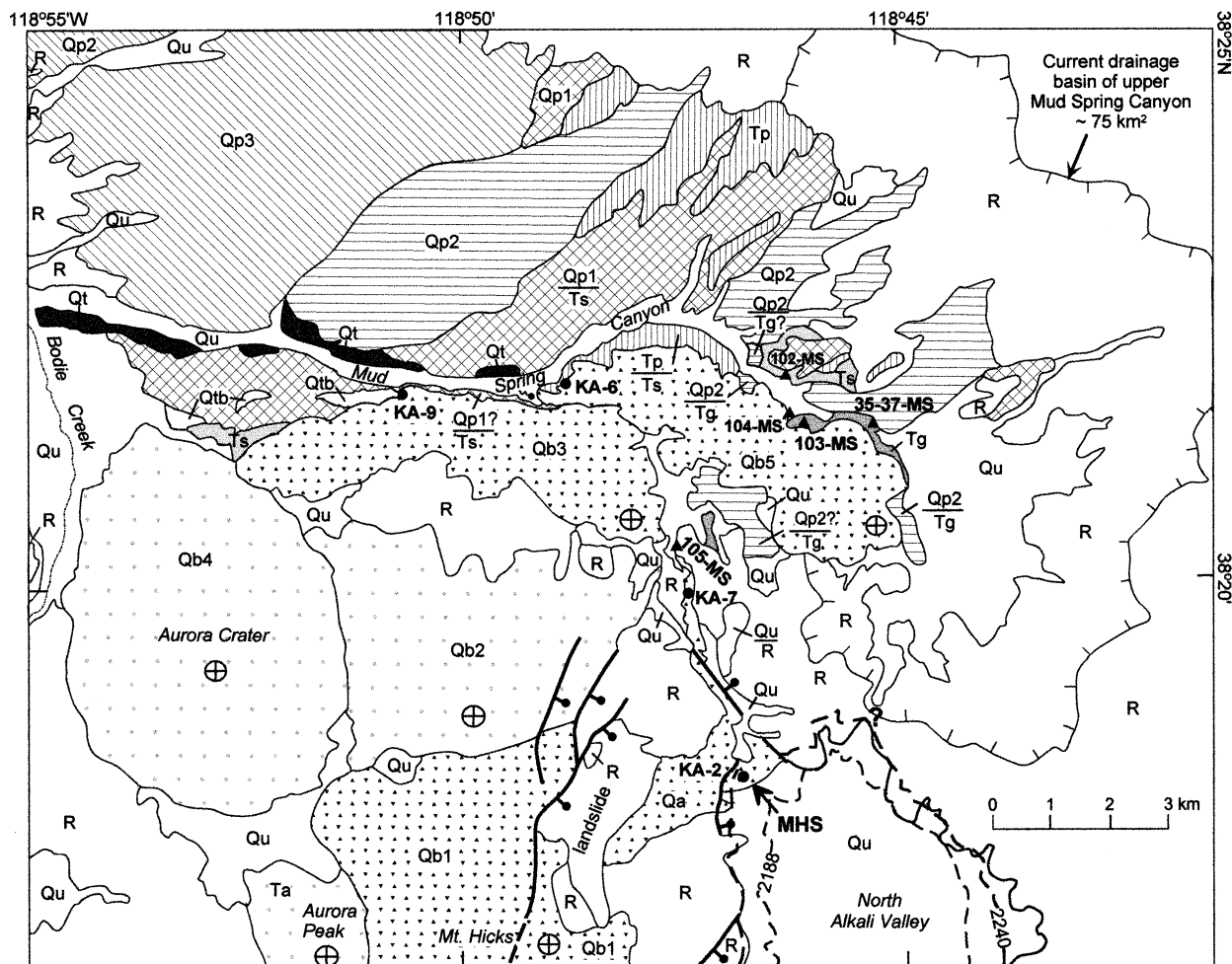
alluvium in the Mud Spring Canyon drainage basin are much less well rounded than most of the clasts on this terrace. The terrace gravel must represent erosion and transport by a competent stream flowing down Hicks Valley, which is currently dry, has no springs, and drains a very small basin. Furthermore, the polished clasts suggest that this stream was carrying beach gravel from an overflowing Lake Russell.

Mud Spring Canyon and Its Paleodrainages

Mud Spring Canyon is dry but for one small spring. The flat-floored, steep-walled canyon is 300–400 m wide and ~40 m deep for much of its length (Figs. 5 and 7B), but narrows quickly above a point ~3 km up-canyon from the western end of flow Qb5. The modern drainage area above the spring is ~75 km², in our judgment disproportionately small compared to the size of the valley. We suggest that much larger discharge is necessary to explain the size of Mud Spring Canyon than could be generated from the modern basin, even during pluvial periods. The modern canyon is cut into alluvial deposits of units Ts and Tg, of Pliocene age, and several capping pediment gravels. Flows Qb5 and Qb3 divide west-draining Mud Spring Canyon from north-draining Hicks Valley.

On the south side of Mud Spring Canyon, the surface of flow Qb3 slopes westward at the same gradient as the canyon floor (Fig. 7A). A $^{40}\text{Ar}/^{39}\text{Ar}$ age on a sample from the west end of flow Qb3 is 1.32 ± 0.08 Ma (EL-KA-9, Table 3). This flow originates from a cone adjacent to the north end of Hicks Valley. All along its northern margin, flow Qb3 overlies gravel of unit Qp1 and the underlying unit Ts (Fig. 7, A and B; Table 4); a baked contact is apparent in one road cut. West of where flow Qb3 diverges from the modern canyon, the flow thickens southward. This thickening is due to descent of the base of the flow toward the axis of a west-draining paleovalley, south of and subparallel to Mud Spring Canyon, that headed near the exit of north-draining Hicks Valley. The paleovalley was floored by alluvium and was incised into units Qp1 and Ts; at the western end of flow Qb3, the basalt-alluvium contact is inset nearly 10 m below the surface of Qp1 to the north. We term this paleovalley “Mud Spring 1” (MS1, Fig. 8)

At least two pediment levels as well as two fluvial terraces are inset below the surface of Qp1 on the north side of Mud Spring Canyon (Table 4, Fig. 5). Adjacent to the western



Sedimentary Units

Qu	Alluvium, undifferentiated (Holocene and late Pleistocene)
Qp3	Youngest pediment deposits (middle Pleistocene)
Qtb	Terrace deposits derived from flow Qb3 (late to middle Pleistocene)
Qt	Terrace deposits of Mud Spring Canyon (late and middle? Pleistocene)
Qp2	Intermediate pediment deposits (middle? to early Pleistocene)
Qp1	Intermediate pediment deposits (early Pleistocene to late Pliocene?)
Tg	Tuffaceous gravely alluvium (late Pliocene)
Tp	Oldest pediment deposits (late Pliocene)
Ts	Alluvial and lacustrine deposits (Pliocene)

Volcanic Units

Qb5	Valley-filling hornblende andesite (<0.08 Ma; Table 3 and Lange et al., 1993)
Qb4	Hornblende andesite of Aurora Crater (0.2-0.5 Ma; Silberman and McKee, 1972; Lange et al., 1993)
Qb3	Valley-filling basaltic andesite (1.3 Ma; Table 3)
Qb2	Basaltic andesite (early Pleistocene)
Qa	Andesite (1.4-1.6 Ma; Tables 2 and 3)
Qb1	Basalt of Mt. Hicks (1.6 Ma; Gilbert et al., 1968)
Ta	Dacite of Aurora Peak (2.7 Ma; Silberman and McKee, 1972)
R	Older volcanic and granitic rocks

Symbols

	Projected 2188-m shoreline (last S overflow)		Eruptive cone
	Projected 2240-m shoreline (last N overflow)		Fault; bar and ball on downthrown side; dotted where concealed
	▲ 35-MS Tephra locality and sample number; EL- prefix deleted for ease of drafting		● KA-7 K-Ar or Ar-Ar sample locality for this study; EL- prefix deleted for ease of drafting

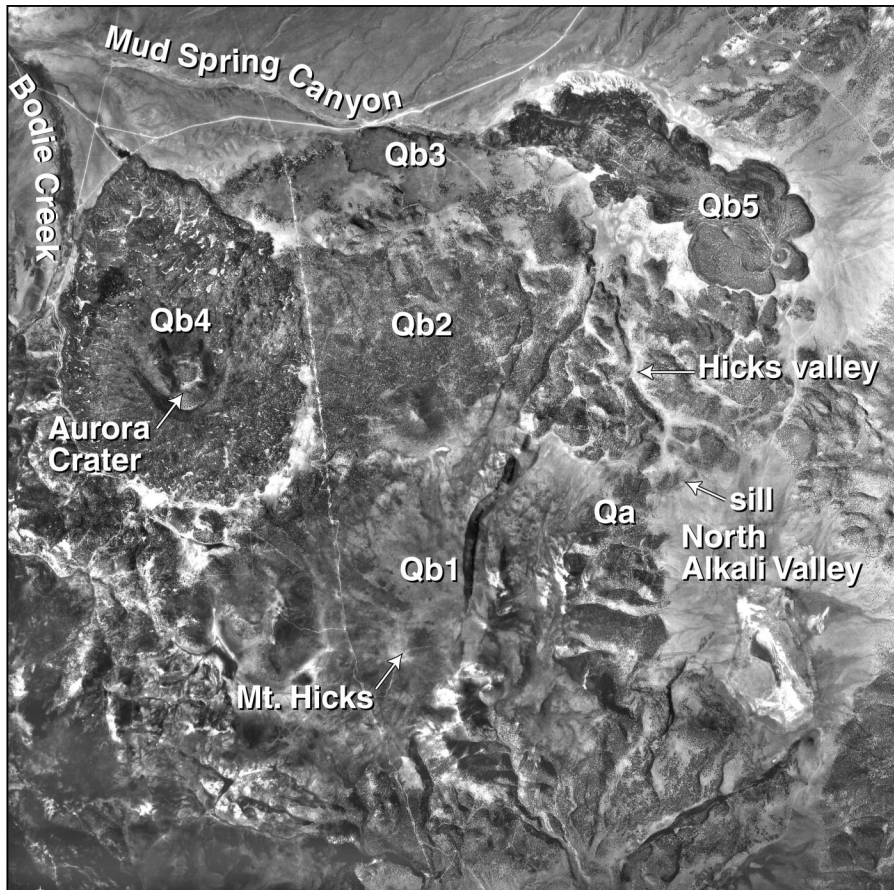


Figure 6. Aerial photograph of the area of the Mount Hicks spillway in northern Alkali Valley and Mud Spring Canyon, showing surface morphology and relationships among basalt flows; photo covers most of the area shown in Figure 5.

reach of the modern canyon, west of flow Qb5, unit Qp2 abuts terrace deposits of unit Qt. The terraces lie 12–20 m above the present drainage; there appear to be at least two nested terrace levels. The higher terrace is only slightly (<2 m) lower than the Qp2 surface. The terraces consist of well-washed, relatively well-bedded fluvial sand and gravel that lap onto reddish, more poorly sorted and bedded fan gravel. Similar terraces do not exist farther upstream along the eastern reach of Mud Spring Canyon. This distribution of terrace remnants suggests that the stream that transported and deposited these fluvial sediments along the western reach of Mud Spring Canyon did not flow down the eastern reach. In addition, the stratigraphic relationships indi-

cate that pediment Qp2 postdates flow Qb3 and that the upper Qt terrace was deposited within a valley incised into pediments Qp2 and Qp1. Pediment Qp3 is lower and younger than Qp2; Qp3 appears to crosscut, hence is younger than, the upper Qt terrace (Fig. 5).

Like flow Qb3, flow Qb5 parallels and slopes westward at the same gradient as Mud Spring Canyon. At its western end, flow Qb5 was clearly emplaced in a west-draining paleovalley defined by abrupt thickening of the flow from ~2 m on the north side to nearly 40 m against flow Qb3. The width of this thick part of the flow is ~300 m, the same as the width of the modern canyon. In contrast, the thickness of the eastern part of the flow is more uniform. Stratigraphic relationships in-

dicate that this paleovalley, termed “Mud Spring 2” (MS2, Fig. 8), was incised on the east and north sides of flow Qb3 after its eruption blocked Mud Spring 1. The northern valley wall of Mud Spring 1 was preserved by its burial beneath the resistant cap of basalt. The modern valley, Mud Spring 3, was incised into the northern valley wall of Mud Spring 2 adjacent to flow Qb5; west of flow Qb5 the courses of Mud Spring 2 and 3 coincide. The east end of Mud Spring 1 and the west end of Mud Spring 2 coincide spatially with the north end of Hicks Valley. We interpret these relationships to mean that Hicks Valley and Mud Spring Valley formed a combined paleodrainage that carried discharge from the Mount Hicks spillway at times when Lake Russell stood much higher than 2188 m.

No drainageway connects the upper part of the Mud Spring basin, east of flow Qb5, with spillways of Lake Russell. The canyon north of flow Qb5 narrows greatly east of sample site EL-KA-6 and merges east of the flow with a smooth piedmont slope covered with young (post-Qb5) fan alluvium. In addition, the late Quaternary age of flow Qb5 indicates that this flow postdates the opening of the southward drainage of Lake Russell. Thus, the eastern, narrow reach of present-day Mud Spring 3 was probably formed by local discharge from the drainage basin of upper Mud Spring Canyon rather than by post-Qb5 overflow from Lake Russell.

DISCUSSION OF OVERFLOW HISTORY OF LAKE RUSSELL

Mono Basin is uniquely situated in a topographically high area with active faulting and volcanism. This setting, combined with climatically induced changes in lake level, has led to a complex Quaternary history of changes in lake level and discharge direction. Although the geomorphic and sedimentary record of these changes is fragmentary, enough evidence exists to place broad constraints on lake levels and on the timing of overflows and incision of discharge channels. The following discussion provides a framework for this history and presents alternative hypotheses where appropriate. We also present implications of the overflow history for the distribution of

Figure 5. Geologic map of the area of the Mount Hicks spillway (MHS) in northern Alkali Valley and Mud Spring Canyon (modified from Stewart et al., 1981; Dohrenwend, 1982; Stewart, 1982; and Lange and Carmichael, 1996). Ages of volcanic rocks from this study (Qa, Qb3, and Qb5, Table 2) and from Lange et al. (1993) and Robinson and Kistler (1986). Dual map units (e.g., Qly/Tv) indicate thin surficial deposit overlying older unit.

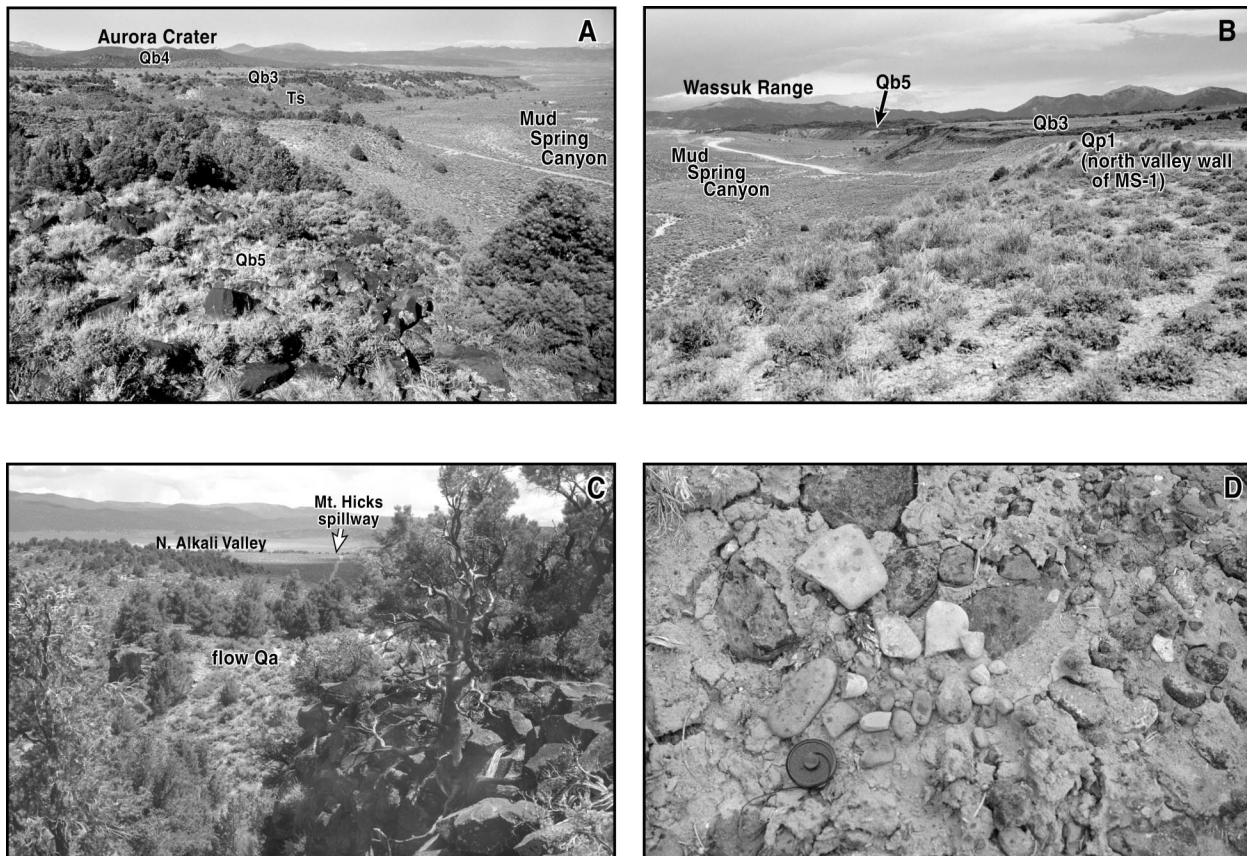


Figure 7. Photographs of features in vicinity of Mount Hicks spillway (scale shown by roads except in D). (A) View west from rugged top of flow Qb5 along north flank of flow Qb3; Qb3 is thin atop northern paleovalley wall of Mud Spring 1 and thickens southward. Qb3 overlies thin gravel capping whitish sediments of unit Ts. (B) View east along Mud Spring Canyon from top of northern valley wall of Mud Spring 1. Note that flow Qb3 in middle distance is lower than foreground as it follows a paleovalley diverging southwest (to right). Rugged surface in distance is flow Qb5 overlying whitish sediments of unit Ts. (C) View south from top of flow Qa in Hicks Valley toward Mount Hicks spillway (fringe of trees in middle ground). (D) Well-rounded and polished pebbles collected from within a 1 m radius on unit Qg at north end of Hicks Valley; pebbles were transported by discharge from Lake Russell; 50 mm lens cap for scale.

fossil and modern fish along the eastern side of the Sierra Nevada.

Pliocene Lakes in Mono Basin

Gilbert et al. (1968) documented a lake that existed between ca. 4 and 3 Ma in the eastern part of Mono Basin and eastward as far as Huntoon Valley (Fig. 2). Fossil mollusks and fish indicate that this lake was fresh, and the fish species suggest an aquatic connection to the north (Hubbs and Miller, 1948; Miller and Smith, 1981). However, Gilbert et al. (1968) thought that the lake had an outlet to the east or south. If so, this outlet must have been obstructed eventually by the voluminous outpouring of basalt east and south of Mono Basin (Gilbert et al., 1968). The apparent absence of deep-water mud and the presence of oolitic sand suggest that this lake was rel-

atively shallow and quiet. Deposits of the Pliocene lake are represented in our study by unit Tl near the Adobe Hills spillway (Fig. 3). A single tephra correlation (EL-102-MS, Table 1) suggests that unit Ts in Mud Spring Canyon (Fig. 5) represents somewhat older fluvial and lacustrine deposition at ca. 4.8 Ma, possibly in a different basin that extended north along the modern valley of the East Walker River (Gilbert and Reynolds, 1973).

Coarse tuffaceous lacustrine deposits constitute the upper part of unit Tl near the Adobe Hills spillway (Fig. 3). Tephra correlations yield ages of 2.5–2.0 Ma (Reheis et al., 2002). The beach gravels in these deposits suggest a higher-energy environment than do the oolite beds in the lower part of unit Tl. The tuffaceous lacustrine beds overlap in age with the tuffaceous alluvium of unit Tg in upper Mud Spring Canyon and Hicks Valley (Fig. 5). The

coarse bedding, poor sorting, and abundance of locally derived clasts in debris-flow beds of unit Tg indicate local derivation rather than deposition by an overflowing lake. Frequent rhyolitic eruptions in the Long Valley–Glass Mountain area south of Mono Basin (Fig. 2; Metz and Mahood, 1985) combined with the older basaltic eruptions probably blocked southward drainage during the late Pliocene. These relationships suggest that Mono Basin may have become internally drained through structural deepening and volcanism between 2.5 and 2.0 Ma, but much more work could be done to elucidate basin history during this period. Clearly, Mono Basin must have been closed on the south and east before 1.6 Ma (the age of flow Qa in the Mount Hicks spillway), and faulting in the structural knee around the Adobe Hills spillway was limited or had not begun.

Early Pleistocene Lakes and Northward Discharge

Hicks Valley 1 is incised into Miocene and Pliocene volcanic rocks and deposits of unit Tg (2.8–2.5 Ma). The paleovalley must have been incised prior to the emplacement of flow Qa. We interpret this incision as the first recorded overflow of Lake Russell sometime after 2.5 Ma and before 1.6 Ma (Figs. 8A, 9). Because flow Qa had not yet been emplaced, the lake need only have risen to ~2219 m (on the basis of modern topography). However, the lake could have been higher because the fault that displaces flow Qa may have also displaced the preexisting sill downward (Fig. 5). Mud Spring 1 conducted lake discharge west from the northern end of Hicks Valley 1 toward the East Walker River.

Flow Qa filled Hicks Valley 1 at ca. 1.6 Ma (Figs. 8B, 9), likely raising the Mount Hicks spillway by at least 20 m (estimated thickness of flow Qa beneath the current sill) and possibly by 60 m (altitude of flow Qa on the west side of the sill). Subsequent dissection, presumably by an overflowing Lake Russell, created Hicks Valley 2. Flow Qb3 filled Mud Spring 1 at ca. 1.3 Ma, following which Mud Spring 2 was incised along the north edge of the flow. Whether discharge occurred between 1.6 and 1.3 Ma, causing incision of flow Qa to form Hicks Valley 2 and reutilizing Mud Spring 1, is unknown.

We suspect that the initial flow over the Mount Hicks spillway after it was raised by flow Qa occurred at the altitude of the highest beach gravels known, which now lie at 2277 m near the Adobe Hills spillway and near 2280 m north of Mono Lake (Fig. 2). To overflow the raised spillway today would require that Mono Lake reach a minimum altitude of 2237 m (the modern sill altitude). However, the altitude of the surface of flow Qa adjacent to the spillway on the footwall block of the fault is 2280 m (Fig. 5). Thus, the pre-faulting surface of flow Qa likely controlled the shoreline altitude required to discharge northward. The sill was probably lowered to its present altitude by a combination of faulting and erosion by overflow. A shoreline altitude of 2280 m suggests an additional possibility: One or more overflows through a pass at 2264 m (marked by a bold query [?] on Fig. 5) east of the Mount Hicks spillway. However, no channel leads north from this pass, and we have not found any other evidence, such as well-rounded gravels, that suggests overflow by this route. The apparently unused pass must have been higher in altitude during early Pleistocene discharge events.

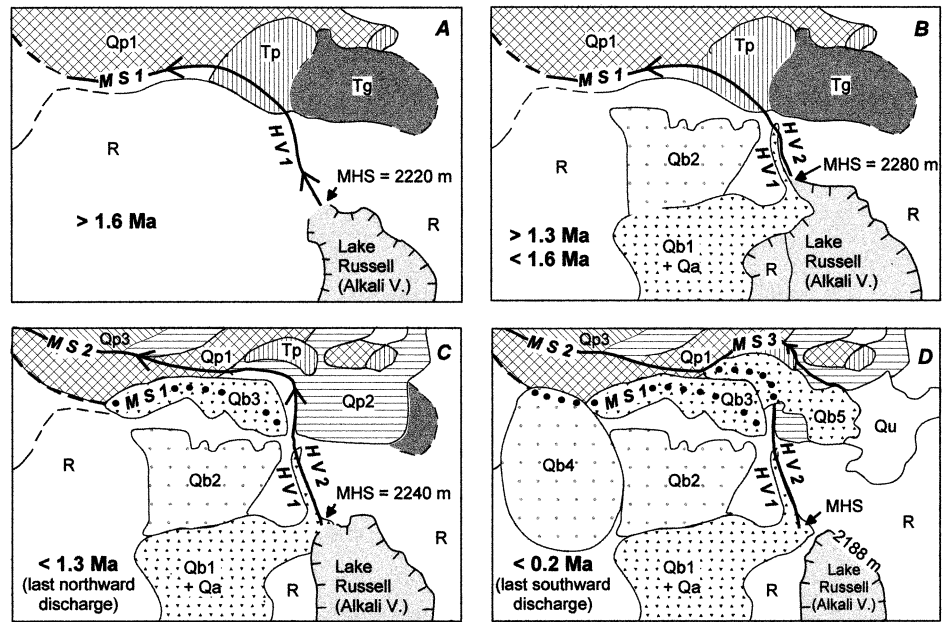


Figure 8. Schematic reconstruction of the sequence of valley incision and volcanic eruptions (A–D) in the area of Mount Hicks spillway (MHS) and Mud Spring Canyon. Map area and units same as in Figure 5, except that northernmost part of Figure 5 is not shown. Heavy lines show drainageways of Hicks Valley (HV1, HV2) and Mud Springs (MS1, MS2, MS3); dashed where uncertain, dotted where buried by volcanic flows. Numbers in meters are estimated altitudes of the spillway and coeval lake levels.

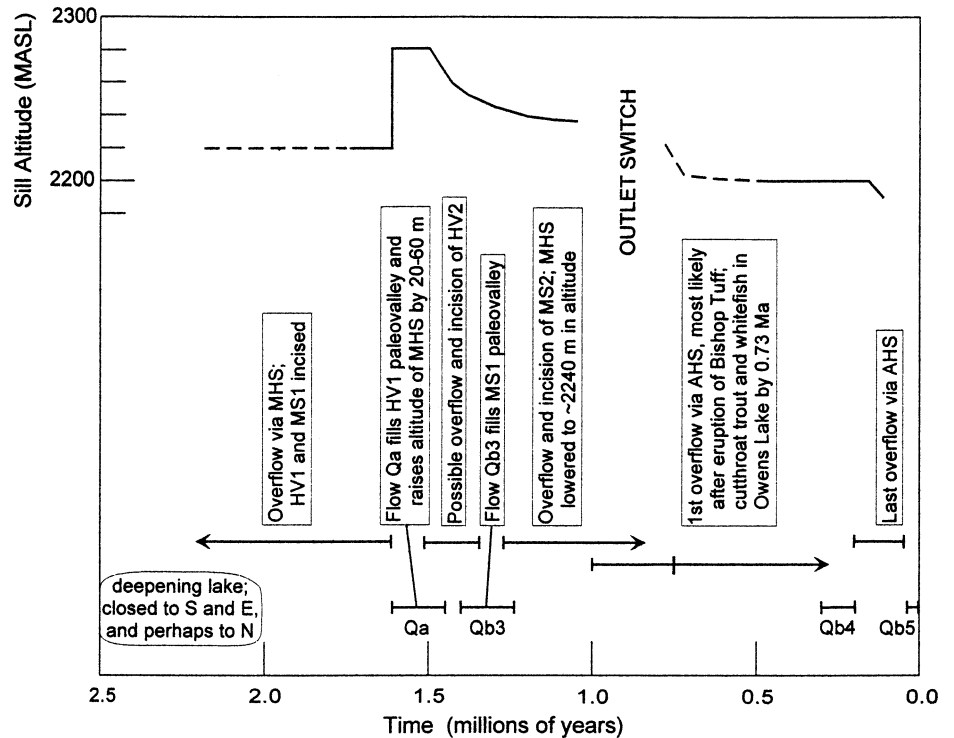


Figure 9. Timeline showing proposed timing of volcanic and discharge events near the Mount Hicks spillway and of discharge events through the Adobe Hills spillway. Lines at top show suggested changes in sill altitude with time.

One or more periods of lake discharge through the Mount Hicks spillway incised Mud Spring 2 after 1.3 Ma (Figs. 8C, 9). The number and timing of these discharges are unknown. Occurrences of beach pebbles at different altitudes between ~2200 m (the maximum level of beach pebbles that seem to be associated with discharge southward) and 2280 m (Fig. 2) could reflect either (1) a single protracted high lake episode and gradual erosion of the northern sill, (2) two or more highstands at significantly different times (our preferred hypothesis), and/or (3) faulting. At several sites with altitudes near or below 2240 m, beach pebbles are particularly abundant and are associated with either a berm with preserved morphology (tombolo west of Bodie Creek, Fig. 2) or snail coquina and beach rock (east of the Adobe Hills spillway, Figs. 3 and 4D). The present altitude of the Mount Hicks spillway is close to 2240 m; thus, berms near this altitude may record the youngest northward discharge. For comparison, the oldest berms with preserved morphology in the Lake Lahontan drainage basin are between 0.8 and 0.6 Ma (Reheis, 1999).

Was northward discharge over the Mount Hicks spillway episodic owing to climatically driven changes in lake level, or was discharge more or less continuous and interrupted only by volcanic eruptions that caused several periods of valley incision? The two pediment surfaces, Qp2 and Qp3, and the two Qt terraces inset below flow Qb3 on the north side of Mud Spring Valley provide evidence of alternating episodes of stability and incision by overflow. There is no reason to suppose that Lake Russell behaved differently during the early Pleistocene than it and all other large lakes throughout the Great Basin behaved during the middle and late Pleistocene, when lake levels fluctuated over hundreds of meters owing to millennial-scale changes in precipitation and temperature (e.g., Morrison, 1991; Reheis, 1999; Oviatt et al., 1999). This logic is supported by stratigraphic and chronologic data from the Walker Lake basin, to which Lake Russell was tributary during the early Pleistocene. These data indicate a minimum of three distinct large-lake episodes separated by intervals of low lake levels (recorded by unconformities, paleosols, and deposition of nonlacustrine sediment) in Walker Lake drainage basin prior to 760 ka: an episode prior to 1.4 Ma, one between 1.4 and 1.0 Ma, and one at ca. 1.0 Ma (Reheis, 1999; Reheis et al., 2002). During some or all of this time, the Walker River may have been tributary only to the Walker Lake basin (Fig. 1; King, 1993); this basin may not have become part of the

Lahontan Basin until ca. 1 Ma (Reheis et al., 2002). Additional input of water from Mono Basin at intervals during the early Pleistocene, if confined to the Walker Lake basin, could help to account for highstands that exceeded the late Pleistocene highstand of Walker Lake (Lake Lahontan).

Mud Spring 2 was filled by flow Qb5 (Fig. 8D), but the youthfulness of this flow provides little information on the timing of the last northward discharge. As faulting continued on the southeast side of Mono Basin, the rim near the Adobe Hills spillway was lowered; when this lowering reached 2237 m, it was then possible for Lake Russell to discharge south.

Opening of the Southward Outlet

Little age control is available on the opening of the Adobe Hills spillway, but geomorphic characteristics of the sill and shoreline terraces and the bouldery terraces downstream of Adobe Valley provide evidence for relative age. For instance, shoreline features near 2188 m in altitude associated with the last overflow south are relatively well preserved, though not as fresh in appearance as those of the late Pleistocene highstand (Putnam, 1949). Radiometric ages on features at or near 2188 m indicate that the last overflow occurred before ca. 50 ka. Multiple periods of discharge should have caused extensive headward erosion by the bedrock channel into the friable lacustrine deposits at the south end of the southern subbasin (Fig. 3).

The two bouldery terraces below Adobe Valley are different in height and weathering characteristics. We interpret these terraces to represent two large-volume discharges from Lake Adobe that were likely caused by overflow from Lake Russell. We hypothesize that the younger and lower terrace (Fig. 4C) correlates with the last southward overflow and that the older and higher terrace represents a much older discharge. Deposits of the older terrace are inset within the Bishop Tuff (Krauskopf and Bateman, 1977), suggesting that the first southward discharge for which evidence is preserved postdates 760 ka (Fig. 9). We suspect that southward discharge into Adobe Valley was not frequent or sustained and may have been limited to only two events. A corollary to this hypothesis is that once faulting had lowered the area of the south outlet to 2237 m (the altitude of the Mount Hicks spillway), Lake Russell must have remained at relatively low levels for a long time so that faulting could lower the Adobe Hills spillway to near 2200 m without erosion by multiple

discharges, which would have produced a more mature-appearing drainage system.

IMPLICATIONS FOR MIGRATION OF AQUATIC SPECIES

Fish and mollusks within the Pliocene lacustrine deposits in eastern Mono Basin have been interpreted to indicate an aquatic connection north into the area of the present-day Lahontan basin at or before 4–3 Ma (Miller and Smith, 1981). The fish fossils—including *Chasmistes* sp., *Catostomus* sp. (suckers), and *Ptychocheilus* sp. (minnow)—are similar to other Pliocene and early Pleistocene fossils found in widely distributed sites east of the Sierra Nevada from the Snake River Plain south to China Lake (Fig. 1). Fossil snails show a similar distribution (Taylor, 1985). One species of modern springsnail, *Pyrgulopsis wongi*, lives in several isolated basins along the eastern side of the Sierra Nevada south of the Carson River (Fig. 1), suggesting a widespread Pliocene population subsequently isolated by tectonic disruptions (Hershler and Sada, 2002). These fish and snail species represent ancient lineages that probably spread through the Great Basin during the Miocene and early Pliocene, when relief was generally lower and drainage basins had different configurations from those of today (Minckley et al., 1986).

Different dispersal timing is suggested by other species of fish and mollusks having more limited distributions or more recent arrivals in the area south of Mono Basin. For example, *Oncorhynchus clarki* (cutthroat trout) fossils occur in Pliocene and early Pleistocene deposits around the Lahontan basin from Walker Lake north (Minckley et al., 1986; Reheis et al., 2002; Smith, 2002), but have not been found in the Pliocene deposits of Mono Basin nor in early Pleistocene deposits exposed near China Lake. In contrast, fossils in a deep core taken from the floor of Owens (dry) Lake (Fig. 1) indicate that *Oncorhynchus* cf. *clarki* and *Prosopium* sp. (whitefish) were present in Owens Lake by 730 ka (Firby et al., 1997). These occurrences could be interpreted to mean that cutthroat trout and whitefish migrated south through Mono Basin during the early to early-middle Pleistocene, at or before 730 ka (though we recognize that a fossil occurrence is a minimum age for the species to be present). In addition, three living species of springsnails in Owens Valley are more closely related to a Lahontan species (Hershler and Sada, 2002) than to springsnails in the basins to the south

and east, suggesting that the Owens Valley species arrived there more recently.

Our reconstructions of the late Pliocene and Pleistocene history of Mono Basin place some constraints on dispersal of aquatic life along the eastern slope of the Sierra Nevada. For example, the coarse tuffaceous lake deposits on the east side of Mono Basin represent a lake that existed between ca. 2.5 and 2.0 Ma. Debris-flow deposits of similar age on the north suggest that the basin was closed on that side, and it seems likely that frequent volcanic eruptions in the Long Valley–Glass Mountain area had closed the basin to the south as well. Thus, easy aquatic dispersal along the east flank of the Sierra Nevada probably had ended by the late Pliocene, and some populations became isolated. A connection to the north via the Mount Hicks spillway was apparently re-established before 1.6 Ma and was sporadically available for aquatic migration until sometime after 1.3 Ma. This connection may have been viable only for species adapted to travel upstream in fast-flowing water, such as cutthroat trout and whitefish.

Sometime after 1.3 Ma, following incision of Mud Spring 2, the Mount Hicks spillway was abandoned and highstands of Lake Russell began to flow through the Adobe Hills spillway, permitting fish and mollusks then living in the lake to move south into the Owens River drainage. Mapping for this study indicates that the new outlet was created by intense faulting in the area of the Adobe Hills spillway. The relatively youthful appearance of many of the fault scarps suggests that this faulting might have accompanied and/or followed the 760 ka (Sarna-Wojcicki et al., 1991) eruption of the Bishop Tuff from Long Valley caldera (Bailey et al., 1976); several studies have concluded that vertical slip rates of fault zones in the area increased markedly after this eruption (dePolo, 1989; Gillespie, 1991; Reheis and Sawyer, 1997). If this scenario is correct, a migration route south from Mono Basin opened sometime after 760 ka.

The two bouldery terraces in the canyon east of Adobe Valley combined with the geomorphic appearances of the canyon and the Adobe Hills spillway suggest two high-volume discharge and incision events—a relatively young event and a much older event that postdates the 760 ka Bishop Tuff. In addition, the presence of cutthroat trout and whitefish remains in the Owens Lake core indicates at least one episode of southward discharge prior to 730 ka. Of course, such fish could have migrated through the Mono Basin during the early Pliocene before the basin became closed on the south, and fossils record-

ing this migration simply have not been found. Nevertheless, it is interesting to speculate on the effect that the 760 ka eruption of the Bishop Tuff from Long Valley caldera might have had on waters and aquatic species in the area. The presence of nearly 40 m of rock and ash interpreted to be the Bishop Tuff beneath Mono Lake (Lajoie, 1968) suggests a catastrophic event that would have heated the water and raised lake level significantly; if the lake were high at this time, the added volume of rock combined with eruption-induced faulting could have triggered an overflow. The Bishop Tuff also flowed into Adobe Valley and far south down the present valley of the Owens River past Bishop (Fig. 1). It seems unlikely that fish could have survived such chaos; thus, the appearance of trout and whitefish in Owens Lake may reflect either colonization prior to 760 ka or rapid dispersal southward between 760 and 730 ka. Could Lake Russell fish have ridden a seiche of water, created by the ash flow pouring into the Mono Basin, south into the Owens River?

In conclusion, this paper provides a foundation for future work. Much more could be done in Adobe and Benton Valleys on possible downstream evidence for large discharges from Lake Russell. Very few areas around Mono Basin have been examined for evidence of higher-than-last-overflow shorelines. Cosmogenic dating techniques could be applied to the beach gravels that represent fragmentary evidence of such highstands. Lastly, the ancient, very high lake levels documented in this study provide additional evidence for climates wetter than those of the late Quaternary along the eastern Sierra during the early and middle Pleistocene (Reheis, 1999; Reheis et al., 2002); the several factors that separately or

together may have produced these wetter climates should be investigated and tested.

APPENDIX

Original analytical data from electron-microprobe analysis of volcanic glass shards separated from tephra layers are shown in Table A1.

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Ted McKee of the U.S. Geological Survey (USGS) kindly provided K-Ar ages for one basalt sample, and the New Mexico Geochronology Research Laboratory provided $^{40}\text{Ar}/^{39}\text{Ar}$ ages of other samples; Elmira Wan (USGS) prepared tephra samples for analysis. We thank Mitchell W. Reynolds (USGS), Alan Gillespie (University of Washington), and Doug Clark (Western Washington University) for helpful reviews and suggestions; Gail Mahood (Stanford University) for pointing out relevant references; and Nadine Calis (USGS Volunteer for Science) for her help in the field.

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TABLE A1. ORIGINAL ANALYTICAL DATA FROM ELECTRON-MICROPROBE ANALYSIS OF VOLCANIC GLASS SHARDS SEPARATED FROM TEPHRA LAYERS

Sample	SiO ₂	Al ₂ O ₃	FeO	MgO	MnO	CaO	TiO ₂	Na ₂ O	K ₂ O	Total
EL-38-AV,T335-10										
Mean (<i>n</i> = 18)	72.98	12.59	0.649	0.031	0.052	0.414	0.063	3.624	4.689	95.10
Std. deviation	1.00	0.43	0.076	0.012	0.041	0.031	0.044	0.208	0.134	1.06
EL-102-MS,T433-6										
Mean (<i>n</i> = 20)	69.96	12.99	1.916	0.114	0.047	0.883	0.163	4.083	3.625	93.78
Std. deviation	0.88	0.16	0.101	0.012	0.029	0.049	0.025	0.360	0.459	0.92
EL-103-MS,T433-7										
Mean (<i>n</i> = 28)	72.31	12.05	0.514	0.049	0.051	0.472	0.086	2.551	5.625	93.71
Std. deviation	0.46	0.14	0.045	0.012	0.028	0.021	0.025	0.341	0.515	0.54
EL-104-MS,T433-8										
Mean (<i>n</i> = 29)	75.59	12.80	1.011	0.027	0.060	0.529	0.051	3.681	4.587	98.34
Std. deviation	0.54	0.14	0.080	0.007	0.033	0.025	0.020	0.115	0.065	0.67
EL-105-MS,T433-9										
Mean (<i>n</i> = 25)	73.01	12.29	0.535	0.054	0.053	0.590	0.081	2.887	5.124	94.63
Std. deviation	0.48	0.12	0.060	0.011	0.030	0.021	0.028	0.243	0.208	0.64

Note: Shards were analyzed by using the JEOL8900* instrument; *n* = number of shards analyzed for each sample. Varying fluid contents account for differences in totals from 100%. Analyses performed by C.E. Meyer, U.S. Geological Survey, Menlo Park, California, 1996–1999.

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