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QUATERNARY LAKES OF THE EASTERN MOJAVE DESERT, CALIFORNIA

EXPANSION BAR
VELOCITY & DEPTH DROPS —
→ DEPOSITION

FRIENDS OF THE PLEISTOCENE

PACIFIC CELL

1985 ANNUAL MEETING

FIELD TRIP GUIDEBOOK

QUATERNARY LAKES
OF THE EASTERN MOJAVE DESERT
CALIFORNIA

FIELD TRIP GUIDE

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Prepared for the 1985 annual meeting of the

Friends of the Pleistocene, Pacific Cell

October 25–27, 1985

QUATERNARY LAKES OF THE EASTERN MOJAVE DESERT, CALIFORNIA

FRIENDS OF THE PLEISTOCENE PACIFIC CELL
1985 ANNUAL MEETING

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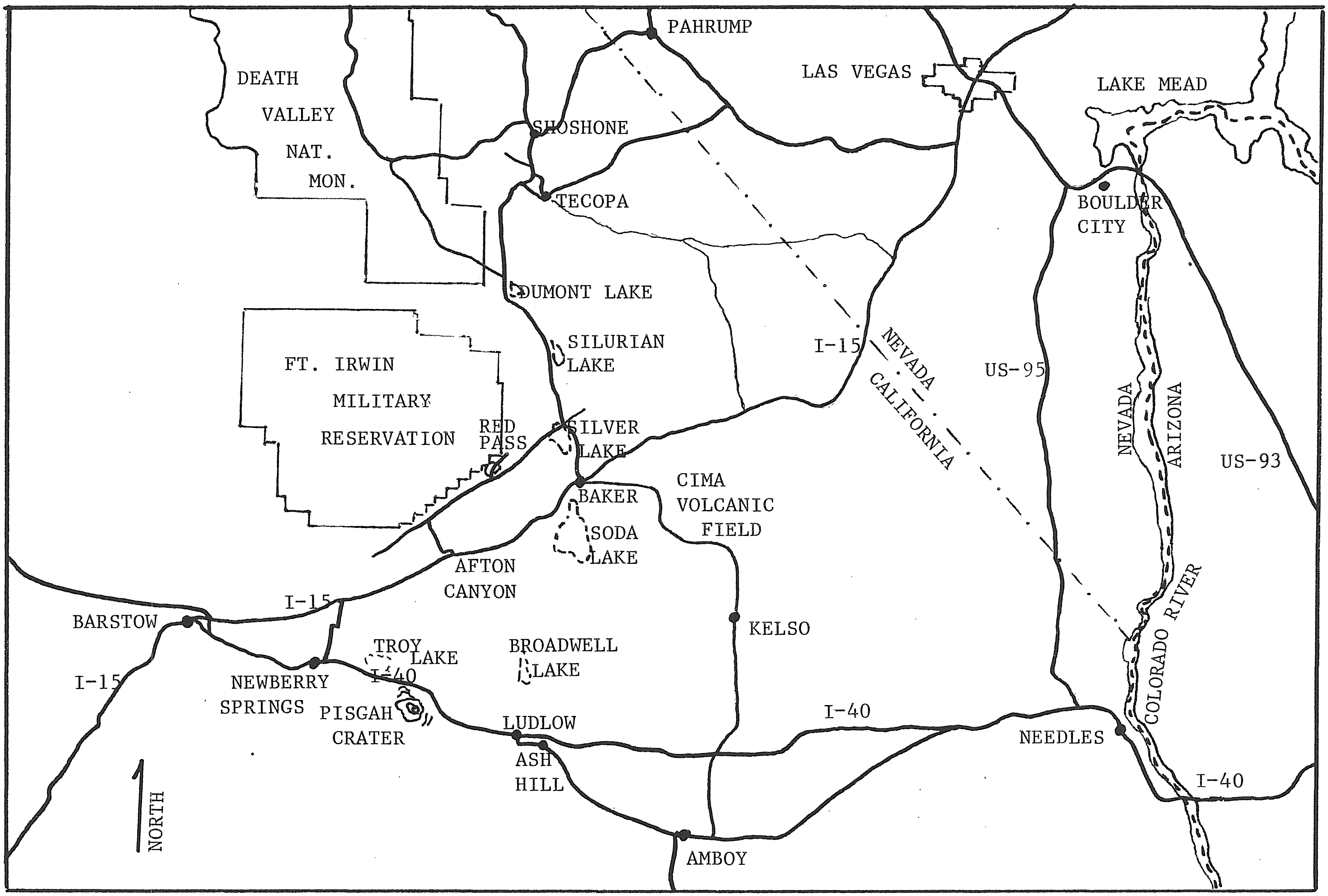
COVER PHOTO: View Northwestward toward Broadwell Lake Basin. Visible in the photo is the Ash Hill Paleochannel thought to have carried overflow discharge from Death Valley as it spilled toward the Colorado River in the Mid-Pleistocene. T. M. Oberlander photo.



MAP OF FIELD TRIP AREA

MID-PLEISTOCENE OVERFLOW OF DEATH VALLEY TOWARD COLORADO RIVER

FIRST DAY ROAD GUIDE

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Friday, October 25, 1985

Departure Time: 8:00 A.M.
Distance: 97.9 miles
Stops: 4

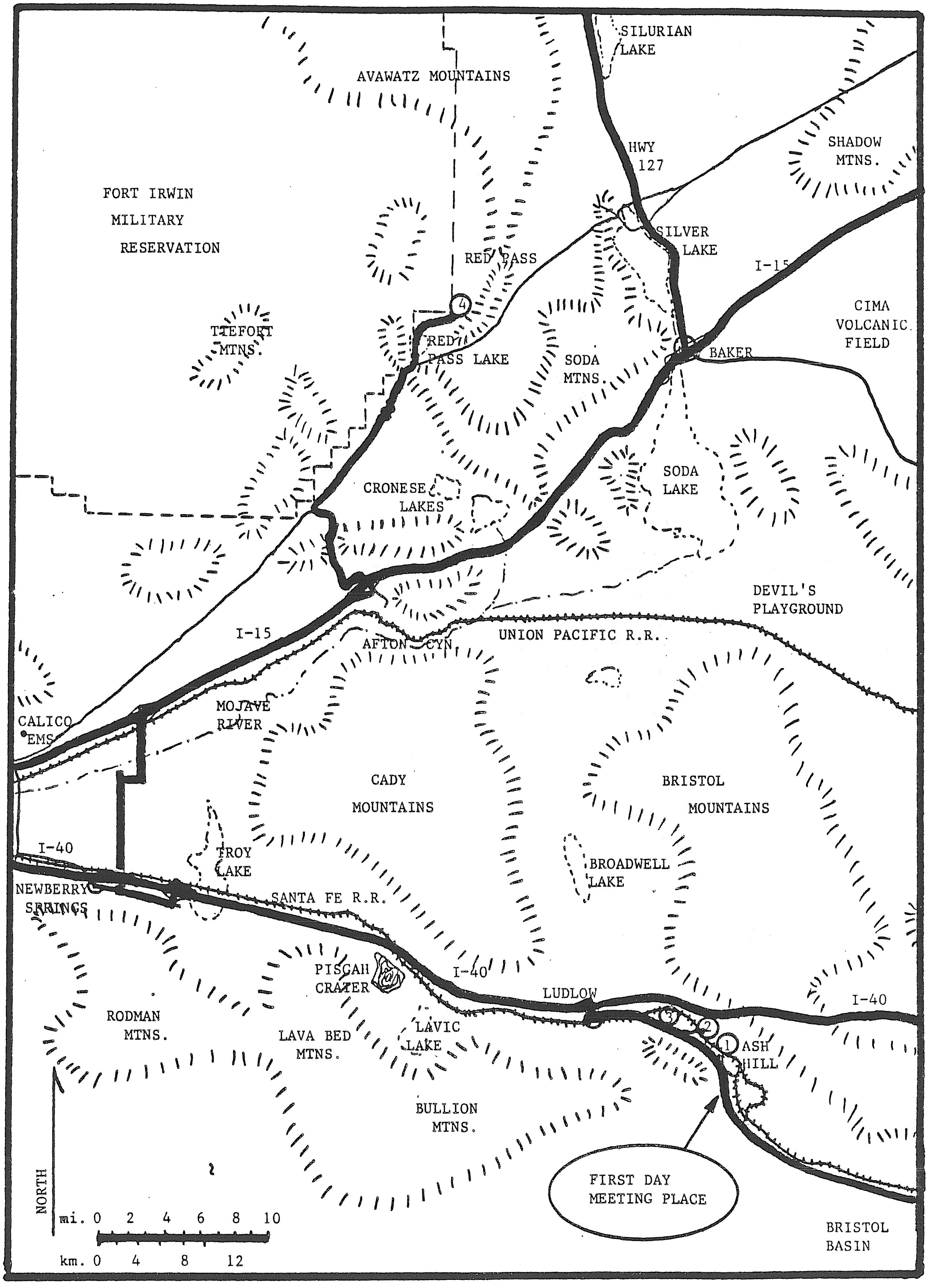
On the first day of the FOP trip, the group will examine geomorphic features in the central Mojave Desert interpreted as evidence of a Pleistocene overflow-stand lake in the Death Valley Basin. As it spilled toward the Colorado River, this lake would have been approximately 300 km in length extending from the town of Ludlow to the north end of Death Valley National Monument. The group will examine a large paleochannel incised in Miocene basalt at the lake's outlet, a long-abandonded (mid-Pleistocene?) alluvial fan cemented by exceptionally-developed calcrete, a relict beach at overflow elevation, and an outcrop of quiet water sediments thought to have been deposited in an estuary of the high stand lake.

DAY 1 - ASH HILL AREA TO RED PASS VALLEY, CALIFORNIA

This road log commences at the first day meeting point. The meeting point may be reached as follows: Drive 8.8 miles east from the center of Ludlow, California, on Old National Trails Highway (formerly US HWY 66). After rounding a gentle curve to the right that drops over the horizon, turn left off the highway onto a straight gravel road that diverges gradually from the highway. Travel .4 mi along this road to the meeting point.

Mileage

0.0	0.0	Start road guide - first day meeting point. Return to Old National Trails Highway heading northwest.
0.4	0.4	Merge with highway heading northwest; proceed a short distance on the highway.
0.8	0.4	Immediately after rounding the first bend to left, bear right off highway onto a wide, straight, graded dirt road that diverges at about 15^{o} from the highway.


SILURIAN LAKE
AVAWATZ MOUNTAINS
HWY 127
SHADOW MTNS.
FORT IRWIN MILITARY RESERVATION
SILVER LAKE
RED PASS
I-15
CIMA VOLCANIC FIELD
TIEFORT MTNS.
RED PASS LAKE
SODA MTNS.
BAKER
SODA LAKE
CRONESE LAKES
DEVIL'S PLAYGROUND
I-15
AFTON CYN
UNION PACIFIC R.R.
CALICO EMS
MOJAVE RIVER
CADY MOUNTAINS
BRISTOL MOUNTAINS
I-40
TROY LAKE
BROADWELL LAKE
NEWBERRY SPRINGS
SANTA FE R.R.
PISGAH CRATER
I-40
LUDLOW
I-40
RODMAN MTNS.
LAVA BED MTNS.
LAVIC LAKE
ASH HILL
BULLION MTNS.
FIRST DAY MEETING PLACE
NORTH
mi. 0 2 4 6 8 10
km. 0 4 8 12
BRISTOL BASIN

0.9 0.1 Intersection of several graded dirt roads, make a 90° turn onto another dirt road that drops toward the Santa Fe tracks.

1.1 0.2 Upon approaching the railroad tracks, turn left onto an irregular dirt road beside the rail line.

1.4 0.3 Day 1, Stop 1, Examination of Death Valley Overflow Paleochannel. Leave vehicles in the vicinity of the railroad cut. Walk 400 m north to the bank of the Ash Hill Paleochannel. Use caution crossing the railroad tracks, as this is the busy Transcontinental line of the Santa Fe.

Basalt ventifacts are abundant here and throughout the Ash Hill Trough. The Trough is a topographic gap for winds as well as for water.

ASH HILL BASALT 5.5 MY OLD

Stop on the south brink of the Ash Hill Paleochannel at a vantage point 20 or 30 meters east of a large pile of old railroad ties. Before you is the straight reach of the paleochannel. The 1884 alignment of the Santa Fe Railroad is visible on the opposite bank-top. To the north beyond the old railroad is the Ash Hill Alluvial Fan which is characterized by an exceptionally-developed pedogenic calcrete. The hills to the north beyond the fan are the Bristol Mountains from which the fan drainage orginally came. Out of sight at the fan apex (seen crossed by Interstate 40) is the diversion that long ago captured the fan's source area. Important features observable from this point are: two riffles and two bedrock pools; one point bar (at the upstream end of the reach); alteration by scour and deposition of the pavemented bank top where you are standing; and a large expansion bar at the downstream end of the straight reach. Mean velocity and discharge at bankfull in this reach are estimated to have been 6.4 m/sec and 2000 m^3/sec (70,000 CFS; Hale, this volume.), respectively. NEED 400mm HK

WATERSHED AREA REQUIREMENT

RIFFLE & POOL SEQUENCE 150 m

Climb to the channel bed. The bed is characterized by a marked dearth of debris. Abraded and jointed basalt bedrock lies immediately beneath the thin veneer of aeolian sand. Note the imbrication of the few large clasts in the bed.

.015 GRADIENT

Walk up the channel past the point bar where the straight reach begins. Bed forms such as riffles, pools, and a point bar in this channel are bedrock anologies to the forms observed in alluvial channels. Presumably, the processes at work to shape the channel bed operate similarly while plucking jointed bedrock as they do when transporting and depositing alluvial bed materials.

Continue up the channel into a long, hook-shaped bend (Hale, this volume.). As you pass between the Sandstone Bridge abutments built in 1884 and abandonded 40 years later, you pass by century-old construction debris that remains unscoured across the channel. Also note the lack of abrasion on the fine sandstone masonry. Forty meters farther upstream in the north bank there is an exposure of a ropy pahoehoe flow surface. Ventifaction is an important sculpting process throughout the Ash Hill area. However, excellent preservation of detail on this constructional feature suggests that ventifaction works quite slowly to alter significantly the shapes of exposed surfaces.

Continue up-channel another 150 meters until reaching the confluence of the second of three fan channels that join the bedrock channel from the north. This is the largest, widest, and (in aerial photography) most prominent of channels coming from the apex of the Ash Hill Fan. For this reason it is interpreted to be the last active channel to carry Bristol Mountain runoff to the bedrock channel. Discussion of the importance of this fan channel can be found in the accompanying article by Hale. Two facts argue strongly that significant flows in these two channels ended about the same time. First, the fan channel is graded to the bed of the rock-cut channel. Second, neither channel has succeeded in clogging the other with slack water deposits. Presumably, large flows on the fan stopped as it was captured. Further erosional and depositional activity on the fan would logically have ceased or at least slowed greatly at that time also. The Stage V pedogenic calcrete sample described in the article by Hale was taken from the bed of this fan channel higher on the fan. The K-horizon conforms everywhere to the surface of the fan and its channels. It is not truncated by channel incision. It appears, thus, to have formed since the last major discharge in both the fan and bedrock channels.

1.4	0.0	Return to the vehicles via the basalt ridge south of the hook-shaped curve. On the way you will pass two (rail) roadcuts exposing (5.5 K-AR MA) Ash Hill Basalt. Once in the vehicles, return to the highway where you left it.
2.0	0.6	Turn right onto the highway and head west. As you proceed, the apex of the Ash Hill Fan is visible at two o'clock in the distance where it is crossed by Interstate Highway 40. The hills bordering the Ash Hill Trough to the north and south are resistant members of a thick Tertiary volcanic section that dips moderately to the southwest. Visible also is the crest of the Ash Hill Divide which can be seen where it is crossed by the railroad at one o'clock.
4.5	2.5	Slow and turn sharply to the right onto a dirt road immediately before the highway drops through a small roadcut.
5.2	0.7	Stop at this dirt road's intersection with the railway at Ash Hill Divide. This is the watershed divide over which the overflowing Death Valley Lake is thought to have spilled.
5.2	0.0	Proceed with caution across the rustic (three or four track) grade crossing. To reiterate, this is the Santa Fe main transcontinental line. It is very busy. Turn slightly to the right and continue across another single track (rarely used).
5.3	0.1	Turn left onto a graded dirt road beside this single track and follow it.
5.8	0.5	Stop at a sharp left turn of this dirt road (in an area of well-exposed calcrete). Park vehicles.

5.8	0.0	<u>Day 1, Stop 2, Examination of advanced stage pedogenic calcrete</u>. The calcrete on which you are standing meets the criteria for Stage V development established by Bachman and Machette (1977). At present there is no direct age control on this calcrete. However, at seven locations in the Southwestern U.S. Stage V calcretes identified by Bachman and Machette (1977) were interpreted to be between .4 MA and 2.0 MA in age. At the two sites nearest to Ash Hill (one at Vidal Junction, 150 km east) Stage V calcretes are associated with surfaces no younger than 1.0 MA. This magnitude of age, if applicable to Ash Hill, is consistent with interpretations by others of most recent (Lake Manly) filling of Death Valley to 75 m elevations (G.I. Smith, personal communication, 1985; and Butler, 1981). The low hills lying 3 km distant west-northwest of where you stand are the Ash Hill Volcanic Vents--our next (and lunch) stop.
5.8	0.0	Return to vehicles. Retrace your route to the Old National Trails Highway. Again, <u>use great caution in crossing the Santa Fe tracks</u>.
7.0	1.2	Upon reaching the highway, turn right onto it and proceed west.
8.7	1.8	Slow and turn onto a graded dirt road departing perpendicularly to the right.
9.2	0.5	At the fork in the road, bear right.
9.5	0.3	Turn right at the T junction onto a graded dirt road that parallels the railroad.
9.7	0.2	Follow dirt road as it detours to the right through a stream bed and resumes its direction parallel to the railroad.
9.9	0.2	Park vehicles at the closest point to the west end of the small hills lying south of the road.
9.9	0.0	<u>Day 1, Stop 3, Wave-rounded pebble site</u>. Leave vehicles and walk 400 m southeast to the top of the first hill at its west end (head toward the large stone claim marker there). Meet for lunch at vantage point 100 m east of the claim marker. This site is one of five at overflow elevation (1950') where the pebble rounding described in the article by Hale was observed. Occasionally, subrounded pebbles can be observed by careful examination of the ground at this site. The rocky outcrop south of the cairn exhibits subrounded particles on its south side. The outcrop is the westernmost surviving remnant of the smaller volcanic vent rim. Approximately one-third of this vent's rim has weathered and eroded away in the 5.5 MA since its formation. Remnants of both the plug and the rim of the larger vent (the western half of which is gone) can be seen south of the lunch site. The highest point on the rim of the larger cone is 615 m elevation and shows no sign of the pebble-rounding observable here. It is located south-southeast of this site at a distance of 700 m.

Broadwell Lake Playa is visible 15 km to the north-northwest. Beyond it and to the right is the 420 m elev., 2.2 km wide pass connecting the Broadwell Basin with Soda Lake Basin and Death Valley beyond. The point where you are seated is interpreted as having been lapped by the waves of the mid-Pleistocene overflowing Death Valley Lake mentioned earlier.

If this lake filled today, it would be 300 km in length and an average of 25 km wide. Its greatest depth--at Badwater in Death Valley--would be 681 m (2234 ft).

9.8	0.0	Return to vehicles. Retrace your route to the Old National Trails Highway.
11.1	1.2	Upon reaching the highway, turn right onto it and head west.
15.4	4.3	Downtown Ludlow is endowed with three gas stations, one cafe, and one motel. There is a mini-grocery store at the Chevron gas station. For several decades, Ludlow's claim to fame was that it was the southern terminus of the Tonopah and Tidewater Railroad. This railroad was built in 1907 to bring borates from Nevada to the Santa Fe Railway. It ceased operation in the 1930s. Remnants of its roadbed can be seen crossing the Broadwell, Soda, Silver, Silurian, Tecopa, and Amargosa Lake basins.
15.4	0.0	Leave Ludlow, heading north out of town.
15.5	0.1	Pass under the freeway and enter the Interstate highway heading west toward Barstow.
15.7	0.2	As the highway begins to climb out of the Broadwell Basin, the wide gap connecting the Broadwell Lake Basin with Death Valley is visible at three o'clock in the distance beyond Broadwell Playa.
23.5	7.8	The freeway passes out of the tectonically less active Broadwell Basin into the Troy and Manix Lake drainages. The Cady Mountains on the right and the Bullion Mountains to the left are mainly Tertiary volcanics. The Pisgah Basalt flow comes into view on the left soon after the rise as does Lavic Dry Lake behind it.
29.0	5.5	Pass Pisgah Crater close by the freeway on the left just before crossing over the Santa Fe Railroad.
34.0	5.0	Pass Hector exit. Continue west on I-40. One-and-one-half miles after the Hector overcrossing the road crosses the Pisgah Fault. This fault is one of many northwest-trending, right-lateral faults that appear to be absorbing the stress created by northward movement of the San Andreas Fault's constraining "Big Bend" on the southern border of the Mojave Desert. Similar activity characterizes three other parallel fault systems between this point and the town of Helendale near Victorville. The generally recognized activity of these faults and of others in this part of the Mojave Tectonic Block effectively prevents mid-Pleistocene lake basin reconstruction here based upon modern topography.

38.5	4.5	Exit at rest area for a brief stop.
38.5	0.0	Return to vehicles. Enter freeway heading west. Soon after leaving the rest stop, the highway crosses the dry bed of Troy Lake. This lake was the south arm of Late Pleistocene Lake Manix.
42.0	3.5	Exit at Fort Cady Road. Turn left, crossing over the Interstate highway. Continue southward.
42.1	0.1	Stop at intersection. Turn right onto Old National Trails Highway.
45.1	3.0	Slow to make right turn onto Newberry Road which crosses Interstate 40 heading north. (Groceries and gasoline are available in Downtown Newberry Springs, 2 miles west of this intersection on Old National Trails Highway).
45.6	0.5	Cross the Santa Fe Railway.
48.0	2.4	Pass Newberry School as you continue north.
51.1	3.1	Slow for right turn onto Riverside Drive.
52.1	1.0	Slow for left turn onto Harvard Road.
53.0	0.9	Cross the channel of the Mojave River. This river drains the north slopes of the San Gabriel Mountains. Like many desert rivers, the Mojave is limited most of the time to subsurface flow. It drains by way of Afton Canyon toward Cronese, Soda, and Silver dry lakes (Pleistocene Lake Mohave), Silurian Dry Lake, Dumont Lake, and Death Valley. Mojave River peak discharge upstream at Victorville during the flood of March 1938 was 2000 m^3/sec (70,600 CFS; see Hale article). On that occasion, in 1916 and in 1983 enough water flowed through this channel to penetrate to Silver Lake (Wells, this volume.).
55.7	2.7	Cross Union Pacific Railroad and Yermo Road. Continue north.
55.8	0.1	At Interstate 15 onramp turn right onto freeway heading east toward Baker and Las Vegas.

Ten kilometers east of this junction is the San Bernardino County Museum - Calico Early Man Archaeological Site.

For the last 22 miles the parade route has traversed the bed of Pleistocene Lake Manix. This lake filled and desiccated on and off through the Pleistocene. It spilled toward Soda Lake several times over a divide south of Afton Canyon (Jefferson, personal communication, 1984). It is generally believed that its basin was breached by fault displacement in Afton Canyon at some time after 21,000 B.P. If Death Valley filled to overflow elevation today, the Manix Basin would form an arm of the overflowing lake. The Calico site would be situated 1.5 km beyond its edge.

59.0 3.2 Manix overcrossing (no exit). At two o'clock 5 km from the freeway is the type section for the Manix lacustrine sequence.

66.0 7.0 I-15 rest area. Between one and two o'clock at a distance of 8 km is the Afton Canyon outlet from the Manix Basin. Do not stop.

70.0 4.0 The road passes among exhumed lacustrine sediments of Lake Manix.

71.8 1.8 Take the Afton Canyon Road exit from the freeway. Afton Canyon Road rests on a long lacustrine bar built by Lake Manix. Turn left crossing over the freeway. Gasoline is sometimes available at the station here.

71.0 0.1 Turn left, heading west onto the frontage road that parallels the freeway.

73.4 1.5 Prepare for a sharp right turn at the end of the frontage road. The pavement ends here. Head northwest on this dirt road. This road winds through unnamed small hills until it meets a private powerline maintenance road associated with transmission lines connecting Hoover Dam with the City of Los Angeles.

80.8 7.4 At the T junction with the powerline road, turn right--heading northeast. Remain on the main branch of the road. The road crosses the Cronese Basin. It is bordered at close range on the left by the Fort Irwin National Training Center.

86.9 6.1 Road crosses Dill Creek at the nadir of its basin transect. It climbs steeply over a gap in the Soda Mountains. As it climbs, there are several small sandy spots. These have been consistently passable to two-wheel drive vehicles. However, keep your vehicle moving as you pass through these.

92.3 5.4 The road crests. The panorama to the northeast includes Red Pass Dry Lake at the base of the slope; the Avawatz Mountains to the north; the Soda Mountains stretching to the east and southeast; and Red Pass Valley--the day's final stop--extending beyond Red Pass Lake toward a narrow gorge in the Southern Avawatz Mountains. The field trip route crosses the middle of the dry lake bed.

92.9 0.6 Junction of several dirt roads at the south end of Red Pass Lake. The route to Red Pass Valley site makes use of a recently established mine access road that crosses the lake. NOTE: Be sure the lake surface is dry before venturing onto it. The graded road drops onto the playa surface at which point it forks. Take the right-hand fork so as to head generally northeastward across the lake bed. The graded road becomes a faint track as it crosses the playa. Stay to the rightof the military reservation boundary signs posted in the lake bed.

94.3 1.4 At the northern margin of the playa, find a bulldozed dirt road with a well-defined berm on each side. Continue north-northwest on this road.

95.1 0.8 Turn right onto an intersecting dirt road that crosses nearly perpendicularly. Follow it generally eastward.

96.7 1.6 Traverse a sandy stretch of the road that is roughly 0.1 mile long. If possible, keep moving in this stretch. If your vehicle has low clearance, you may want to ride one wheel up on the center berm a bit.

97.9 1.2 Day 1, Stop 4, Examination of Quiet Water Sediments at Red Pass Valley
Road drops several meters onto an old strath terrace. Park on the road close to where it drops precipitously into a deeply (10 m) incised fluvial channel. Walk north on the channel bank top about 150 m from the vehicles to a vantage point for viewing the opposite channel bank. This vantage point is at 1880 ft elevation. Drainage from this point is generally northeastward through Red Pass, a bedrock gorge near the south end of the Avawatz Mountains. Red Pass drains a 350 km^2 catchment which includes most of the south slope of the Avawatz Mountins. The gorge at Red Pass is similar in proportions to one at Afton Canyon which is thought to have been carved by the Mojave River sometime after 20,000 BP (Jefferson, et al. 1982). At Afton Canyon the Mid-Pleistocene Mojave River catchment was probably 9,370 km^2 (Snyder and others, 1964). It may have been nearly twice that size if the Antelope Valley was Tributary to the Mojave River. The bank opposite the vantage point exhibits a complex history of deposition and erosion. Deposition of at least 10 m of buff-to-pink quiet-water (probably lacustrine) sediments was followed by as much as 3 m of erosion which truncated the pink beds. The eroded top of the pink sediments can be seen several meters below the bank top. Erosion was followed by alluvial deposition that culminated in a smoothly graded fan surface. This surface or one closely related to it was isolated and remained inactive long enough to permit development of a varnished, pavemented "jogging" surface on the top of the bluff. Evidence of soil development, including a weak K-horizon, can be seen in the bluff half a meter beneath the present upper surface. Although this soil carbonate is clearly much less developed than that observed at Ash Hill, soil formation is only one in a series of geomorphic events all of which have happened serially since deposition of the quiet water sediments at this site. Finally, downcutting has proceeded to at least the 10 m depth observable today exhuming the features described above.

From this vantage point proceed to the pavement surface on top of the bluff, which will be the last stop of the day. The best (safest) access to the top of the bluff is at its upstream end and approximately 250 m upstream. At the upstream end of the bluff a linear alignment can be observed among (1) the bluff's upstream slope, (2) the dogleg bend in the channel west of it, (3) the upstream limit of at least one other erosionally isolated interfluve (visible to the east-northeast) and (4) Red Pass itself. It is not presently certain if this alignment is the result of faulting that might also have opened Red Pass or if it is the result of the vagaries of dissection in a landscape with a complex history. It is worth noting that desert pavement is moderately well-developed on the upstream slope of the bluff. Whether originally disturbed by faulting or erosional dissection, this surface has remained undisturbed for enough time to form this pavement.

The pavement on top of the bluff consists of an exceptionally smooth and closely packed mosaic with occasional cobbles protruding above the surface. It is well varnished and highly reflective. This degree of reflective sheen is characteristic of all the older fan surfaces in this valley.

The two principal questions that are unanswered at this site are the following: First, was the gorge at Red Pass open or closed when the pink quiet water sediments were deposited? Second, were they deposited at the same time as the overflow-stand lake that has been the principal subject of this roadguide?

Support for interpretation of these quiet water sediments as belonging to an overflow of Death Valley comes from the following: The near coincidence of elevation between sediments at Red Pass and at the Ash Hill Divide; the unlikelihood that a watershed as small as that of Red Pass could have excavated as large a gap as Red Pass since deposition of the quiet water sediments; and finally, the possible consistency of age magnitude between the Red Pass Valley sediments and Ash Hill Fan.

The official FOP campsite for Friday evening will be at the Red Pass site. Motels and package stores are in Baker about an hour's drive 17 miles east of Red Pass Lake on the powerline road then south 9 miles on US 127. Camping is acceptable anywhere between here and US 127 except in the immediate vicinity of the powerlines and on the military reservation. The Saturday program begins at 9:00 a.m. at the intersection of the powerline road and US 127. Please park on the gravel powerline road and not on the paved highway.

LATE QUATERNARY GEOMORPHIC HISTORY OF THE SILVER LAKE AND SALT SPRING HILLS AREAS EASTERN MOJAVE DESERT, CALIFORNIA

SECOND DAY ROAD GUIDE

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Saturday, October 26, 1985

Departure Time:	9:00 a.m.
Distance:	62.1 miles
Stops:	6

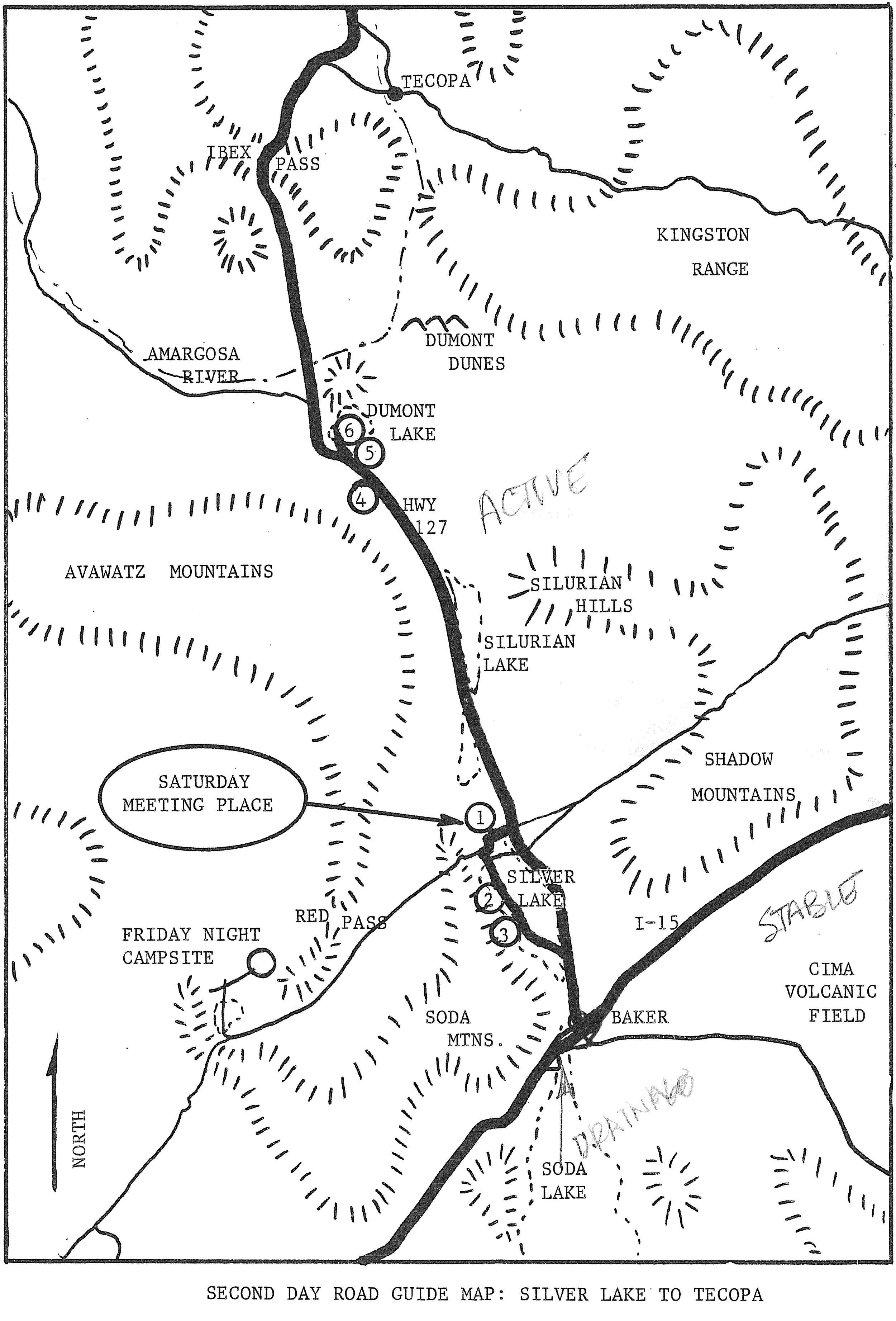
The second day of the trip will be spent examining Late Quaternary soil development and surficial processes on desert piedmonts adjacent to Silver Lake and Lake Dumont. Silver Lake, a playa in the path of the Mojave River, is well suited to the study of such processes, because nearly two dozen radiocarbon and cation ratio dates of Late Quaternary lacustrine features provide good age control. A variety of geomorphic evidence at Lake Dumont reveals a complex history. The group will examine shorelines, alluvial fan surfaces, and outcrops of several ages. Discussion will center on rates of soil development, evolution of piedmont surfaces, and the sequence of Late Quaternary climatic fluctuations in this part of the Mojave Desert.

The meeting point may be reached as follows: Drive 9.3 miles north of the town of Baker, California, on Highway 127; or, start at the first night campsite in Red Pass Valley. Retrace your route (5 miles) to the Los Angeles Water and Power utility road at the south end of Red Pass Lake. Turn left, heading east, onto the utility road. Travel 17.1 miles to the meeting place at the junction of California Highway 127 and the utility road. Be sure to stay generally under the transmission lines the whole way. Park off the paved highway when you reach the meeting point.

Mileage

0.0 0.0 Start road guide - Day 2, Stop 1. Leave vehicles and walk to vantage point to view Silver Lake and the vicinity of its outlet.

0.0 0.0 Return to vehicles. Head west on utility road.



SECOND DAY ROAD GUIDE MAP: SILVER LAKE TO TECOPA

2.5 2.5 Intersection with graded dirt road that crosses diagonally. This is the old Silver Lake-Randsburg Road. Turn sharply left descending gradually onto Silver Lake Playa.

3.0 0.5 As road curves to left on Sliver Lake Playa, continue straight ahead onto a poorly marked dirt track that departs toward the south. Follow this track southward along the western margin of Silver Lake.

4.1 1.2 <u>Day 2, Stop 2 - Geomorphology of Silver Lake and latest Pleistocene history of Lake Tecopa</u>.

The following summary of late Quaternary geology and geologic history in the Silver Lake area is synthesized from Ore and Warren (1971) and Wells et al. (this volume).

Lake Mojave is the name given to a shallow late Pleistocene pluvial lake that occupied the basin of Soda Lake and Silver Lake playas. Local drainage into this basin is predominantly from the west-sloping piedmonts of the Halloran Hills and the Cima Volcanic Field; however, water presently reaching these two playas comes almost entirely from the Mojave River system which originates in the San Bernardino Mountains about 200 km to the southwest. Only unusually heavy precipitation in this source area causes significant flooding in Silver Lake Playa. During this century, periods of significant high water in Silver Lake have been limited to January 1916 to July 1917, March 1938 to September 1939, June 1969, and February to April 1983.

The 26 km^2 floor of Silver Lake is extremely flat and lies at a uniform elevation of 907 feet. (This is 15 to 25 feet lower than the floor of Soda Lake south of Baker; consequently, Silver Lake must be filled to a depth of at least 15 feet before any significant amount of water can accumulate in Soda Lake.) A narrow channel at the north end of Silver Lake Playa was the outlet for Lake Mojave, and overflow through this channel drained northward along the Salt Creek drainage into Lake Manly at the southern end of Death Valley. The original elevation of this outlet was approximately 943 feet, and the maximum depth of Lake Mojave was only 36 feet.

Several prominent beaches, beach ridges and wave cut cliffs, at elevations between 935 and 945 feet along the north and west margins of Silver Lake Playa, mark the higher levels of Lake Mojave. The beaches, veneered with stone pavements and locally partly dissected by shallow drainageways, are underlain by varying sequences of intercalated clean fine to coarse sands, silty sand and clayey silt, and well-sorted to poorly sorted gravel. Pelecypod shells and lithoid tufa fragments are locally abundant. The wave-cut cliffs are fringed by aprons of coarse angular blocks locally coated with thick rinds of lithoid tufa.

NO HISTORY > 15,500 yr BP.

Radiocarbon dating of these shells and tufa rinds forms the basis of the following late Pleistocene lacustrine chronology. A major lacustrine episode filled the basin from before 15,500 yr B.P. until least 10,500 yr B.P. Water overflowed through the north outlet during this period and maximum water depths were approximately 36 feet. During the latter part of this period the channel outlet was lowered to 936 feet. A final, somewhat lower lake filled the basin from sometime after 9500 yr B.P. until about 8000 yr B.P. (Wells et al., this volume).

Early man apparently occupied the Lake Mojave area several times and artifacts are abundant on beaches and pre-lake alluvial fans. At least 26 archeological sites dating from the Lake Mojave period are located in the Silver Lake - Soda Lake area. These sites, distributed on level areas dispersed around the shoreline of Lake Mojave, form a pattern of seasonal camps about a resource center, the pluvial lake supporting vegetation along its margins and attracting wildlife to its shores. Artifacts include projectile points, knives, scrapers, drills, hammer stones, anvils, and heavy chopping tools (Warren et al., 1980).

4.2 0.0 Return to vehicles. Proceed south along margin of Silver Lake towards Stop 3.

5.0 0.8 <u>Day 2, Stop 3 - Latest Pleistocene and Holocene alluvial stratigraphy at Silver Lake - (1.4 km or 0.9 mile walk)</u>. This part of Silver Lake Playa and the adjacent piedmont (Fig. 2) provides an excellent opportunity to examine landscape-stratigraphic relations between late Pleistocene and Holocene lacustrine shorelines and deposits, alluvial fan surfaces and deposits, hillslope deposits, eolian deposits, and soils (Wells et al., this volume). Date shorelines of Lake Mojave (Ore and Warren, 1971, and Wells et al., this volume) and dated alluvial fan surfaces (Dorn, 1984) allow time constraints to be placed on latest Quaternary lacustrine, alluvial, colluvial, and eolian events.

Five major alluvial fan surfaces have been assigned approximate ages based on relations with radiocarbon dated shorelines and cation-ratio dating of rock varnish on fan surfaces. The oldest fan surface, unit Qf3, is dated at greater than 15,500 yr B.P. (Wells et al., this volume) and at approximately 37,000 yr B.P. (Dorn, 1984). This surface is characterized by darkly varnished, interlocking stone pavements and less than 0.1 m of original depositional relief (Fig. 3). It supports a moderately-developed soil with a 40 cm thick Bw horizon and Stage II to Stage III secondary carbonate. The next younger fan surface, unit Qf4, cuts the 15,500 to 10,500 yr B.P. shoreline and locally grades to the 9,500 to 8,000 yr B.P. shoreline. This surface is characterized by moderately to darkly varnished stone pavements on remnant bar and swale topography and a weakly to moderately developed soil.

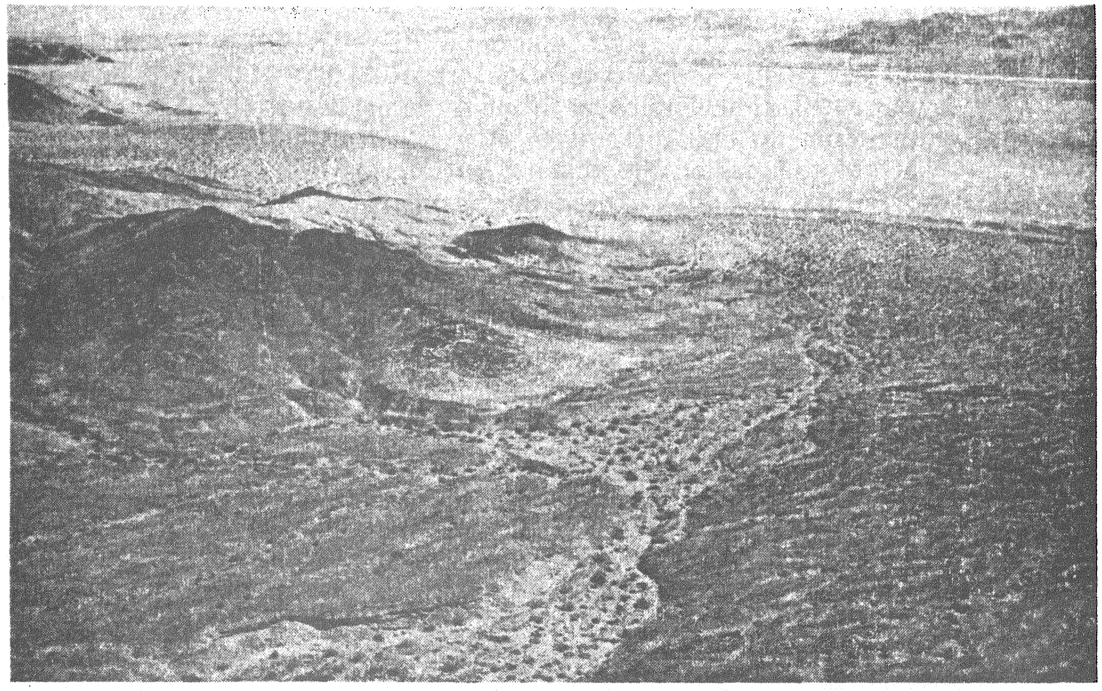


Figure 2. Holocene and late Pleistocene alluvial fan surfaces on the west shore of Silver Lake playa (area of Day 2 - Stop 3). Aerial view northeast down the distal piedmont of the Soda Mountains and across Silver Lake playa. Photograph by J. C. Dohrenwend.

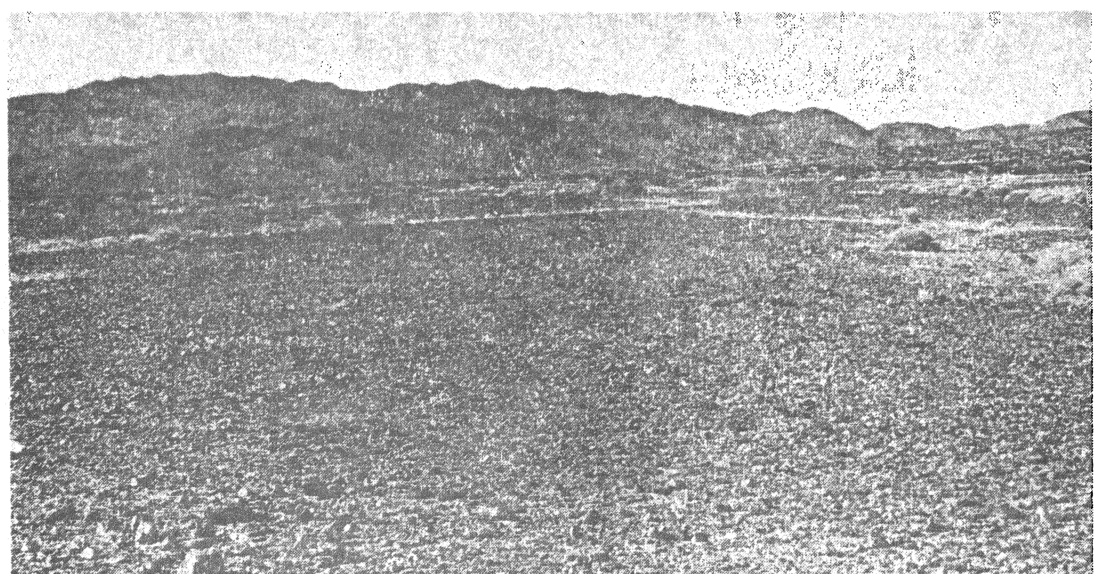


Figure 3. An absence of bar and swale topography and an interlocking stone pavement on a late Pleistocene unit Qf3 alluvial fan surface, west shore of Pleistocene Lake Mojave, Silver Lake, California. View west upfan. Photograph by J. C. Dohrenwend.

An eolian unit is buried by Qf4 deposits, and the Qf4 surface grades to and interfingers with a colluvial mantle developed on hillslopes of metamorphic bedrock. Several strandlines occur below the 9,500 to 8,000 yr B.P. shoreline. These strandlines were formed either during the dying stages of Lake Mojave or during prehistoric and historic floods. Alluvial unit Qf5 cuts across the higher strandlines but does not cut the lower strandlines. Unit Qf5 surfaces are characterized by prominent bar and swale topography with up to 0.37 m depositional relief, incipient varnishing of clast surfaces, and weakly developed soils with Stage I carbonate morphology. Units Qf6 and Qf7 are morphologically similar to unit Qf5 but display no rock varnish on clast surfaces. Soils on unit Qf6 exhibit only a thin vesicular A horizon, and no soils occur on unit Qf7.

After the first major lake decline at approximately 10,500 yr B.P. and prior to a lower, short-lived lake stand between approximately 9,500 and 8,000 yr B.P., an eolian depositional event occurred on the piedmonts and hillslopes of the Silver Lake area. During the lower lake stand, hillslope deposits became unstable and a period of hillslope stripping and alluvial fan building occurred. Following the final demise of Lake Mojave at approximately 8,000 yr B.P., several playa flooding episodes and at least two fan building periods have occurred in the area.

5.0 0.0 Return to vehicles. Continue southeast on the faint track across Silver Lake.

8.0 3.0 Track crosses the abandoned grade of the Tonopah and Tidewater (T & T) Railroad. The realization of F. M. "Borax" Smith's dream to develop his Death Valley borate properties, the T & T (locally referred to as the "tired and tardy") was completed in 1907 and operated between Ludlow, California, and Beatty, Nevada, by way of Silver Lake, Tecopa, and Death Valley Junction. Tracks for the Tonopah and Tidewater were laid through this area in 1906 and abandoned in 1939 (Warren et al., 1980). The grade has been flooded at this location at least four times since its construction. Using the railroad grade as a baseline and facing west across the playa, the following panorama can be seen: Baker, 9:00; Soda Mountains, 9:30 to 1:00; Avawatz Mountains, 1:00 to 2:30; and the Lake Mojave outlet area, 3:00. Proceed east across the playa to Highway 127.

8.3 0.3 Junction with California 127. Turn right and proceed south towards Baker.

9.3 1.0 Enjoy good views of the Cima Volcanic Field at 9:00 to 10:00 and the deeply-dissected piedmont high on the east flank of the Soda Mountains at 2:00 to 3:00.

11.7	2.4	Stop sign at the intersection of California 127 and Business I-15, Baker, California. Use this opportunity to replenish groceries, gasoline, and water as needed.
11.7	0.0	Junction California Highway 127 and Business I-15 in Baker. Bid adieu to Baker and proceed north on Calif. 127 toward Shoshone and Death Valley. (The Field Trip route is shown on Map 3.)
12.5	0.0	Road curves left (west). Soda Mountains at 7:00 to 11:00; Avawatz Mountains at 11:00; and Halloran Hills (Hollow Hills) at 1:00 to 2:00. On the left, the Soda Mountains are cut by a north-northwest trending fault zone, one to two miles wide, that is topographically expressed as a zone of low to moderate relief just below the range crest at approximately 9:00. East of the fault, the hills and low mountains between Silver Lake and the range crest are composed of Precambrian metamorphic rocks that have been intruded by Mesozoic plutonic rocks. West of the fault zone (and largely out of sight beyond the range crest) the range is underlain by upper Paleozoic and Mesozoic rocks that have been folded and faulted and then intruded by Cretaceous granitic rocks. On the right, a Mesozoic structural dome in Precambrian rocks intruded by Mesozoic plutons forms the west flank of the Halloran Hills (Troxel, 1982).
13.3	0.8	Baker Airport turnoff on the left.
14.7	1.4	Mile post 3.0. Abundant basalt clasts can be seen on the distal alluvial fan surface east of the road. These clasts, up to 0.9 m in diameter, have been transported across 1.0 to 2.5 degree slopes from the northern part of the Cima Volcanic Field approximately 12 miles to the east. For the next 7 miles, the road traverses Holocene distal alluvial fan deposits issuing from the west flank of the Halloran Hills (on the right) as it curves gradually around the east shore of Silver Lake Playa (on the left). Lake Mojave, the ancestor of Silver Lake, attained a maximum elevation of approximately 943 feet and a maximum depth of 36 feet during latest Pleistocene time (Ore and Warren, 1971). Had it existed at that time, the road at this point and for the next several miles north would have lain under 20 to 30 feet of water. Refer to Wells et al. (this volume) for a more complete discussion of Lake Mojave.
19.8	5.1	Intersection with an unmarked dirt road. This dirt road runs up across the piedmont of the Halloran Hills (on the right) and straight across the north end of Silver Lake Playa (on the left). Site of Silver Lake, the village (elevation 909 feet). This settlement, a stopping point along the Tonopah and Tidewater Railroad, served as the major supply point for mines in the surrounding hills. Built after the railroad was laid through here in 1906, this town of about 100 people boasted a depot with a telegraph, a post office, a general store, and several saloons and boarding houses. In 1916 Silver Lake (the playa), the Tonopah and Tidewater Railroad (at last living up to its name) and even Silver Lake (the village) were flooded. Despite a program of

urban renewal and relocation to higher and dryer environs, most business had departed by the 1920's, the post office by 1933, and even the railroad by 1939 (Mendenhall, 1909; Paher, 1973).

20.7 0.9 Road crosses a prominent Lake Mojave beach ridge (Ore and Warren, 1971).

21.1 0.4 Road passes beneath high voltage transmission lines from Hoover Dam. The power line access road (on the left) occupies the crest of another Lake Mojave beach ridge and crosses the outlet channel bay of the old lake approximately 0.6 miles west of this point.

21.3 0.2 Crest of the divide between Silver Lake and Silurian Valley (approximate elevation 950 feet). For the next twenty miles the road follows the course of Salt Creek as it descends gradually to 390 feet at its confluence with the Amargosa River at the extreme south end of Death Valley. During late Pleistocene time, water from the Mojave River flowed across Soda Lake into Silver Lake, then overflowed Silver Lake into the Salt Creek drainage through a narrow cut (approximately one-half mile west of this point beyond the low ridge on the left). From here it flowed into Silurian Lake, about 8 miles further north, and finally into Death Valley's Lake Manly.

22.3 1.0 Road curves left and heads north towards the east flank of the Avawatz Mountains. These mountains will dominate the western skyline for most of the next 20 miles. The Avawatz Mountains (9:30 to 11:30) are composed of Precambrian gneiss and Precambrian to Mesozoic plutonic rocks and overlain by Cenozoic sedimentary and volcanic rocks (Troxel, 1982). The south end of the range is dominated by an extensive and deeply dissected pediment complex developed on a moderately deformed sequence of Tertiary and early Quaternary (?) sediments. The position and state of preservation of this pediment suggest significant southward tilting of the range during lates Cenozoic time.

25.7 3.4 Mile post 14. Good view of the dissected pediment surfaces on the south end of the Avawatz Mountains (9:00 to 9:30).

26.7 1.0 Mile post 15. Road curves right and for the next 12 miles follows the axis of Silurian Valley as it curves gradually around the distal edge of the Avawatz piedmont. Here as in many other areas of the eastern Mojave, remnants of older, more dissected alluvial fans are concentrated along or near the mountain front and all levels of alluvial surfaces converge towards the basin axis. In the vicinity of the highway, morphostratigraphic criteria indicate that the piedmont surface is veneered by lates Pleistocene and Holocene deposits.

27.7 1.0 Mile post 16. On the left, well-paved early Holocene and latest Pleistocene (?) fan surfaces extend to within 50 m of the road.

30.1 2.4 For the next 2.5 miles Silurian Dry Lake lies just a few hundred yards east of the highway (on the right). Also on the right, the Silurian HIlls (3:00), the Kingston Range (2:00), and Alexander and Sperry hills (12:30 to 1:00) form the eastern skyline. The Silurian Hills contain a remarkably well-preserved and well-exposed section of Middle Proterozoic through Middle Cambrian sedimentary rocks. The Kingston Range is a mass of Tertiary granitic rocks flanked on the east and north by many of the same Precambrian and Cambrian formations that underlie the Silurian Hills. The Alexander Hills and that part of the Sperry Hills visible from this vantage point are mostly coarse granitic gravels of late Tertiary age and local origin (Troxel, 1982).

The Dumont Dunes, the largest of four small dune fields that form a 25 km long chain from southernmost Death Valley across northern Silurian Valley, lie nearly straight ahead in the distance at the base of the Sperry Hills. Pyramidal dunes and long, sinuous dunes predominate in this dune field; the tallest dune rises to a height of approximately 120 m. Although most eolian features in this dune field indicate opposing winds, barchan migration and seasonal wind streaks indicate that winds from the south predominate over winds from other directions. Along the west side of the field, small barchans moved N. 23°E. at an average rate of 3.5 m/yr between 1952 and 1978; and along the southeast edge of the field, barchans moved due north at an average rate of about 5 m/yr between 1945 and 1952 (Smith, 1984).

33.7 3.6 Mile post 22. On the right, well-developed examples of climbing dunes partly bury the Valjean Hills (3:00). The Kingston Range fills the skyline beyond.

34.7 1.0 Mile post 23. On the left, Quaternary faulting is apparent near the front of the Avawatz Mountains. The age of the dissected upfaulted piedmont remnant immediately adjacent to the mountain front at 9:00 has not been determined.

37.7 3.0 Mile post 26. The Salt Spring Hills lie across the field trip route from 9:00 to 1:30. The Dumont Dunes lie at 1:30 to 2:00. This view affords a fine perspective of the continuous sand sheet that stretches 5 km eastward across northern Silurian Valley from the Dumont Dunes to the Valjean Hills. The Valjean Dunes, at the base of the Valjean Hills, are mostly north-facing barchanoid dunes less than 10 m high. Between 1945 and 1978 these dunes moved N. 19° E. at average rates of 5 to 6 m/yr (Smith, 1984),

38.4 0.7 Road gradually curves left about 30 degrees and follows Salt Creek through the Salt Spring Hills, an east dipping section of Cambrian carbonate and clastic sedimentary strata.

38.6 0.2 Day 2, Stop 4 - Late Quaternary stratigraphy in the Salt Spring Hills Area - (0.1 km walk). Turn left onto raised gravel road. Walk to alluvial fan to left of road. Qf2 deposits, which dominate central portion of fan, are truncated by lacustrine shoreline at an elevation of 175 m. Secondary fans comprised of Qf4a deposits have formed on the north and south distal flanks of this fan complex.

Latest Pleistocene and Holocene alluvial fan deposits have been identified on the basis of soil profile development and surface morphology, including depositional bar relief, pavement development, and varnish cover. Seven alluvial fan units have been described in the Salt Spring Hills area (Ritter, this volume).

Unit Qf1 is the oldest subaerially exposed fan deposit in the study area, prading into hillslopes in proximal fan areas. Soils developed in Agl deposits exhibit a 50-cm thick Bt horizon and Stage 3 secondary carbonate morphology. Depositional bars have been reduced to single clast relief and strong interlocking pavement, with greater than 90 percent varnish, cover on surface clasts. Surface morphology of units Qf2 and Qf3 is characterized by recognizable bar and swale topography, interlocking desert pavement, and moderately to darkly varnished surface clasts. Soils developed in Qf2 have a 26- to 52-cm thick Bw horizon with Stage 2 carbonate morphology, whereas Qf3 soils have a 7-cm thick Bw horizon and Stage 2 carbonate morphology. Lacustrine shorelines cut the Qf2 surface but are apparently truncated by Qf3 deposits. Soils in Qf4a, the net youngest deposit, exhibit a vesicular A horizon and Stage 1 carbonate morphology; depositional bar relief ranges from 10 to 39 cm and surface clasts are lightly to moderately varnished. Qf4b deposits, which have similar soil development and surface characteristics, are inset into unit Qf4a deposits. Qf5 deposits have maximum bar and swale relief (24.0-63.5 cm) and a weakly-developed pavement with incipient varnish on exposed clast surfaces. Soild development is limited to a thin vesicular A horizon. Active washes have been designated unit Qf6.

Alluvial units in the Salt Spring Hills area have been correlated with the Silver Lake Piedmont deposits, and the timing of major geomorphic events has been compared (Ritter, this volume).

38.9 0.3 Day 2, Stop 5 - Fan evolution in the Salt Spring Hills Area--external and internal controls on incision and deposition - (0.5 km traverse). Pull vehicle off on right side of road as far as possible. Traverse across middle to late Holocene alluvial units and alluvial-lacustrine relationships.

39.4 0.5 Day 2, Stop 6 - Dumont Lake versus Lake Manly high stand - (0.5 km walk). Turn right onto abandoned road. Walk northeast to Salt Creek - Dumont Lake (?) incision through Salt Spring Hills.

39.6 0.2 Crest of a low pass through the Salt Spring Hills. Southern Death Valley and the Panamint Range lie straight ahead. The site of the abandoned mining settlement of Salt Springs lies about 0.5 miles to the east. This locality first served as a stopping point for pack trains traveling the Old Spanish Trail from New Mexixo and Utah to Southern California in the 1830's. J. C. Fremont stopped here in 1844; "A very poor camping place - a swampy, salty spot, with a very little unwholesome grass. The water rose in the springs entirely too salt to drink." In fact, these waters are an almost saturated solution of Glauber and Epsom salts, and persons delirious from thirst could easily drink a fatal dose. Salt Springs has been discontinuously active as a gold mining and milling site since the 1850's (Mendenhall, 1909; Paher, 1973).

39.9 0.3 Mile post 29. Road begins a gradual 60 degree curve to the right and drops into the valley of the Amargosa River.

40.5 0.6 Junction with dirt road into Death Valley. Monument commemorates the Wade Party's 1849 "escape" from Death Valley via this route. Continue straight ahead. On the left, the Saddle Peak Hills are at 11:00 to 12:00. On the right, climbing dunes veneered by varnished colluvium and alluvium, bury lower parts of the west flank of the Salt Spring Hills.

41.9 1.4 Mile post 31. For the next 2.7 miles the road crosses the active fan of the Amargosa River. Low active sand dunes of the Little Dumont Dune field lie a few hundred meters east of the road (2:30 to 3:30). Parallel, east-southeast-trending ridges, typically less than 10 m high, 60 m apart, and approximately 500 m long occupy most of the field. No obvious movement of this dune field can be detected by comparing ten sets of sequential aerial photographs taken between 1945 and 1982, but several dunes at the field's north end have diappeared (Smith, 1984).

42.8 0.9 Main channel (?) of the Amargosa River. The Amargosa drains an area of approximately 15,000 km^2 and terminates in the Badwater salt pan of Death Valley (approximately 240 feet below sea level).

44.6 1.8 North side of the Amargosa fan. Road crosses onto the piedmont of the Saddle Peak and Sperry Hills and begins its ascent to Ibex Pass. The Saddle Peak Hills lie on the left (9:00 to 12:00), the Sperry Hills straight ahead and to the right (12:00 to 3:00), and the Dumont Dunes can be seen at 3:00 to 4:00 across the Amargosa River fan. The Saddle Peak Hills are made up of Cambrian sedimentary rocks and the Sperry Hills are underlain by granitic rocks capped by thick and extensive deposits of very coarse granitic gravel (Troxel, 1982).

48.6 4.0 For the next 2.3 miles the road climbs across a bedrock pediment that comprises the upper piedmont of the Sperry Hills.

50.5 1.9 Road on the left to a microwave station located several hundred meters west of the highway.

50.9 0.4 Road begins to wind through deeply-weathered granitic rocks and coarse granitic gravels near the crest of the Sperry Hills. Not surprisingly, these granitic rocks and their coars sedimentary derivatives are not easily distinguishable by casual observation.

52.4 1.5 Crest of Ibex Pass (elevation 2,090 feet, 637 m) on California 127. For the next 6.5 miles the road descends into the Tecopa Basin. The Tecopa Basin is a roughly triangular lowland of about 500 km^2 surrounded by varied and irregular uplands. The Nopah and Resting Spring ranges form the eastern margin; the Dublin Hills (ahead) and the Ibex Hills (on the left) flank the basin on the west; and the Sperry Hills (to the right) flanks the basin on the south.

52.8 0.4 Elevation 2,000 feet. Straight ahead the south end of the Resting Spring Range forms the eastern margin of the Tecopa Basin. The Nopah Range fills the skyline beyond.

The Resting Spring Range is an east-dipping fault block composed of approximately 5,000 m of latest Precambrian and Cambrian sedimentary strata. Dark gray dolomite of the Bonanza King Formation forms the crest of the southern part of this range; predominantly clastic strata underlie the Bonanza King (Burchfiel, et al., 1982). The Nopah Range is also underlain by the latest Precambrian and Cambrian strata, an unusually complete section dominated by dolomite and limestone with minor amounts of quartzite and other non-carbonate clastic rocks (Troxel and Cooper, 1982).

Dissected beds of Lake Tecopa spread westward from the foot of the Resting Spring Range. Lake Tecopa is the name given to the lake that occupied the Tecopa Basin for most of Late Pliocene through Middle Pleistocene tome. Deposits of the lake underlie nearly 240 km^2 or approximately one half of the area of the basin. The lake beds occur at elevations as high as 560 m around the periphery of the basin and slope gently basinward with dips of generally less than one degree (the combined effect of original depositional attitude and post-depositional compaction).

54.8 1.6 Mile post 2. Road begins a gradual curve to the left and descends across the Sperry Hills piedmont.

56.8 2.0 Mile post 4. Road curves right approximately 45 degrees. Dissection of the Sperry Hills piedmont begins at this point.

57.2 0.4 Lacustrine deposits of Lake Tecopa appear beneath the dissected alluvial cover. From this point to Shoshone, 15 km to the north, the road winds its way through bluffs and low hills carved from these lake beds.

Sometime betweenapproximately 0.5 and 0.3 m.y.B.P., the barrier of thick late Tertiary fanglomerate that had closed that basin at its southern end was breached and Lake Tecopa was drained. Subsequent dissection has sculptured the lake beds into complex bedlands near the basin axis and deeply dissected gravel-capped piedmonts along basin margins. These lake beds attain a stratigraphic thickness in excess of 70 m (Sheppard and Gude, 1968; Hillhouse, in press). In the center of the basin, they are predominantly mudstone and claystone; towards basin margins, they grade into siltstone and sandstone with interbedded conglomerate. Ledge and caprock forming calcareous strata are also common near basin margins.

57.8 0.6 The Lava Creek Tuff crops out along both sides of the road as a prominent greenish-yellow bed approximately 0.5 m thick. The distinctive greenish-yellow color is caused by alteration of the tuff of Phillipsite (Sheppard and Gude, 1968). This unit has been traced almost continuously along the west margin of the basin from this area northward to Shoshone (Chesterman, 1972; Hillhouse, in press).

At least 12 tuff beds occur within the lacustrine deposits of the Tecopa Basin. From Late Pliocene to Middle Pleistocene time, the 240 km^2 Lake Tecopa was the sump for approximately 6,000 km^2 of the upper Amargosa drainage and served as a highly effective trap for concentrating air-fall tuffs blown into the region. The three thickest of these ash beds have been comprehensively described by Sheppard and Gude (1968), carefully mapped by Hillhouse (in press), and reliably correlated with isotopically dated volcanic sources by Izett et al. (1970), Izett and Wilcox (1982), and Sarna-Wojcicki et al. (1984). From youngest to oldest, these three ash beds are designated: Tuff A, which correlates with member B of the Lava Creek Tuff of the Yellowstone caldera (K-Ar age of 0.62 m.y.); Tuff B, which correlates with the Bishop Tuff of the Long Valley caldera (K-Ar age of 0.73 m.y.); and Tuff C, which tentatively correlates with the Huckleberry Ridge Tuff of the Yellowstone caldera (K-Ar age of 2.02 m.y.). These tuffs provide the basis for reliable time control of the early and middle Pleistocene lacustrine stratigraphy of the basin.

58.3 0.5 Road curves to the right. Impressive outcrops of the Tecopa lake beds form imposing walls along both sides of the road.

59.2 0.9 Junction to the right with the Old Spanish Trail Highway, a paved road to Tecopa (the town) and Tecopa (the hot springs).

61.2 2.0 Scattered, poorly preserved ruins (on both sides of the road) mark the site of the Amargosa Borax works. Borax was discovered here in 1882 and the site was purchased by William Coleman, owner of Harmony Borax Works in Death Valley. Coleman worked these deposits during the summers of 1883 to 1887 in preference to the more extensive deposits of Death Valley. The mere 110 degree mid-summer temperatures here were more suitable for borax extraction than the 120 to 130 degree temperatures on Death Valley's floor (Paher, 1973).

61.6 0.4 Road passes onto the Holocene-late Pleistocene alluvial fan of Greenwater Valley Wash.

62.1 0.5 Junction with the Tecopa Hot Springs Road (paved road on the right) near the apex of the Greenwater Valley Wash fan. End of the second day Field Trip Guide.

You are now in the middle of the Tecopa Basin, one of the more delightful areas for desert camping in the Northern Mojave. Peaceful and picturesque campsites can be found in almost any of the washes and small canyons that dissect the Tecopa lake beds along the west side of the basin from here to Shoshone. To help preserve the area, please stay on established roads and tracks, and try to leave your camping spot in better condition than when you found it. So stop, relax, and enjoy the revealing exposures of the Tecopa lake beds all around you.

For those who are drawn by bright city lights, Tecopa Hot Springs is located about two miles down the Tecopa Hot Springs Road. Hot springs bathing, RV hookups, ice cold beer, and--believe it or not--wholesome, home-style meals can be had for a price. Or you might consider continuing north for approximately 5 miles along California 127 to the bustling metropolis of Shoshone, a famous desert watering hole since 1907 when the Tonopah and Tidewater Railroad was completed through the area. At that time, the Pacific Coast Borax Company contracted for an eating establishment at Shoshone "for the convenience of passengers and crew....Wooden benches and table covered by a tent were presided over by a Chinese cook....The sign read 'All You Can Eat for 50¢" (Keeling, 1976). Today, the Crow Bar continues this long tradition of dining excellence. Shoshone also boasts a naturally-heated outdoor swimming pool (It's owned by the trailer park across the road, so if you're in the mood for an evening swim, stop by at the office and obtain permission. It will probably cost you a dollar or so--including shower--but it's a bargin at twice the price after a hot, dusty day in the desert).

THIRD DAY FIELD GUIDE
MIDDLE AND LATE QUATERNARY DISSECTION
OF THE TECOPA BASIN, CALIFORNIA
OPTIONAL BEFORE BREAKFAST EXCURSIONS

Unquestionably one of the finest ways to experience the surficial geology of the Tecopa basin is to enjoy an early morning run (preferably just before sunrise) around the lake beds or up and down any of the surrounding piedmonts. Weathered surfaces on the lake beds are generally soft and punky (easy on the ankles and knees) and the Pleistocene piedmonts also serve as reasonably comfortable running areas. These surfaces are almost always veneered with well sorted, interlocking stone pavements underlain by thick vesicular A horizons. For those inclined to such pursuits, the following alternatives are recommended:

For joggers: Start at the junction of California 127 and California 178 just south of Shoshone. Head west up the dirt road opposite California 178 keeping to the right. You will climb along a broad wash where several small chambers carved into a spectacular 2 to 3 m thick bed of almost pure Lava Creek ash stare vacantly from the low cliffs on both sides of the road. (If you care to explore, knock before entering. At least one of the chambers is still inhabited.) Continue on up towards the Dublin Hills. After a couple of kilometers or so you will enter a large embayment within the Dublin Hills that is floored by a pediment veneered with Holocene and late(?) Pleistocene alluvial gravels. In the early morning light, this spot can be quite captivating. Return by the same route or add some variety by running down the ridges (actually dissected remnants of middle Pleistocene piedmont surfaces) on either side of the road. Total elevation gain - approximately 150 m.

For runners: Start at the northern limits of Tecopa Hot Springs (the town) where the Tecopa Hot Springs Road crosses the basin floor. Leave the road and head northeast across a broad flat towards the Resting Spring Range, 7 or 8 kilometers distant. Pick your own course up the washes or along the dissected remnants of the post-lake erosion surface that cuts across the lake beds. Along the washes you will find abundant outcrops of the lower and middle parts of the lacustrine section and good exposures of Tuffs B and C. Along the ridges you will observe darkly varnished, well sorted gravel veneers (locally no more than one layer of clasts thick), patterned ground formed by micropiping, rounded 'popcorn' covered hills of weathered mudstone and claystone, and occasional sinkholes and small pipes along the ridge flanks. This is a good area for almost any length of run from 5 to 20 km (but if you reach the mountain front you have probably gone too far). Total elevation gain - 100 to 400 m depending on whether you go up and over the ridges or stick to the easy going along the washes.

For hard-core cross-country runners: Start at Stop 2 of the third day field guide, 1.9 miles southeast along the Old Spanish Trail Highway from its junction with California 127. Run southwest towards the highest peak in the Sperry Hills (along the northwest side of a low irregular ridge that extends from the highway to the Sperry Hills). Head up across modern and Holocene erosion surfaces and alluvial deposits towards the high, deeply dissected pediment surface that bevels the northeast flank of the Sperry Hills. Improvise the climbing of ridges and crossing of deep washes and small canyons as you go. The lower ridges (generally below 460 m or 1500 ft elevation) are classic examples of a landscape in the advanced stages of dissection by 'megapiping'. These ridges display an irregular topography that is reminiscent of dead ice terrane. Vertical pipes 0.5 to 2 m wide and as much as 15 m deep, sinkholes up to 30 m across and 10 m deep, and steep-sided blind valleys up to 15 m deep and several 100 m long conspire to block your way. (*WARNING* - these areas can be *EXTREMELY UNSTABLE* and should be avoided when running or when walking alone. Plan ahead and keep to the low ground as much as possible.) The upper ridges (above 490 m or 1600 ft elevation) are the dissected remnants of at least two middle Pleistocene erosion surfaces. These remnants are broad, flat to slightly rounded, and covered with darkly varnished stone pavements. The oldest surface is truncated by the high lake shoreline at approximately 535 m elevation. The presence of wave cut notches along the walls of canyons dissecting this surface indicate that it was partly dissected before the last high stand of the lake. Before returning to the basin floor, pause to admire the panoramic view of the Tecopa basin. Total elevation gain - 200 to 600 m (depending on how many canyons you cross and how many sinkholes you encounter unexpectedly).

THIRD DAY FIELD GUIDE
MIDDLE AND LATE QUATERNARY DISSECTION OF THE TECOPA BASIN, CALIFORNIA

MILEAGE

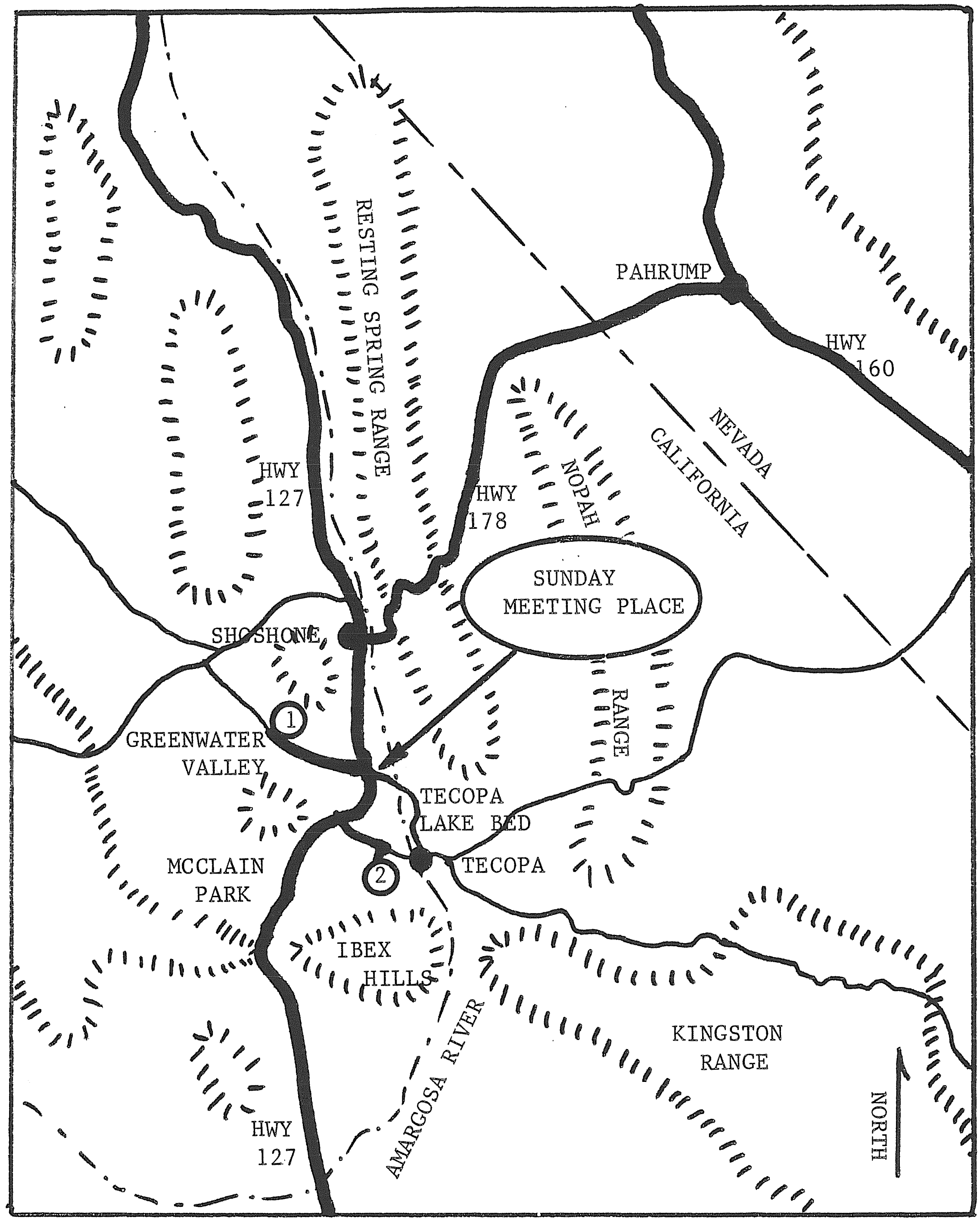
0.0 0.0 Intersection of Tecopa Hot Springs Road and California 127 on the alluvial fan of Greenwater Valley Wash. Drive north on California 127 towards Shoshone.

0.2 0.2 Turn left onto the Greenwater Valley road, an unmarked graded dirt road. Drive west up the fan into the mouth of the wash. For the next 0.8 miles, exposures along both sides of the road display the slight (generally less than 2°) discondancy between the lake beds and the alluvial gravel cap of the Greenwater erosion surface.

1.2 1.0 Road climbs out of the wash onto the Greenwater erosion surface, a post-lake pediment cut across the lacustrine deposits of the basin. Road continues up across the Greenwater surface for the next 2.3 miles (approximately).

Three general erosion surfaces occur within the Tecopa basin: (a) the Amargosa surface which includes the modern fluvial plain of the Amargosa River and the numerous lesser drainageways that incise the basin below the level of the high lake shoreline, (b) the Greenwater surface (between 0.25 and 0.15 m.y. old) which bevels the lacustrine deposits of the basin, and (c) the Sperry surface (somewhat older than 0.3 to 0.5 m.y.) which grades to the approximate level of the high shoreline but is truncated by that shoreline.

The Greenwater surface is the most extensive and prominent pediment surface in the Tecopa basin. Broad, nearly continuous remnants of this surface bevel lacustrine deposits on the piedmonts of the Resting Spring Range, Dublin Hills and Sperry Hills. However, the surface is best preserved south of the Dublin Hills in the area of Greenwater Valley Wash. Pediment capping gravels vary in thickness from as much as several meters in the vicinity of the high lake shoreline to as little as a single layer of



DAY THREE ROAD GUIDE MAP, GREENWATER VALLEY AND TECOPA.

gravel clasts on degraded surface remnants near the basin axis. Along Greenwater Valley Wash, the pediment gravels are generally less than one meter thick and the lake beds are locally exposed in shallow drainageways and scattered prospect pits. Remnants of the Greenwater surface are generally flat and are almost everywhere covered by darkly varnished, interlocking stone pavements.

1.8 0.6 For the next 1.7 miles, thin alluvial deposits and shallow drainageways of Holocene and late Pleistocene age locally modify the Greenwater surface.

2.8 1.0 On the right, a profoundly degraded and deeply embayed erosional scarp, 3 to 5 m high, marks a probable intermediate shoreline of Pleistocene Lake Tecopa.

3.5 0.7 Road runs along the north bank of Greenwater Valley Wash. Late Pleistocene and Holocene alluvial deposits within the wash are inset 5 to 10 m below the level of the Greenwater surface.

3.6 0.1 On the right, a second highly degraded and deeply embayed erosional scarp, approximately 5 m high, marks the level of the high Lake Tecopa shoreline (at approximately 520 m elevation along the west margin of the basin). Road climbs across this scarp onto a well-dissected remnant of the middle Pleistocene Sperry surface and continues up along the south edge of this surface for approximately 1.4 miles.

Dissected remnants of the Sperry surface grade to the approximate level of the high Lake Tecopa shoreline but are truncated by that shoreline. Isolated remnants of the Sperry surface and of coeval alluvial fans are preserved on the upper to middle parts of all the piedmonts flanking the Tecopa basin. Although broadly but gently rounded and deeply dissected by steep-sided drainageways, remnants of this surface are typically covered by darkly varnished, interlocking stone pavements.

5.0 1.4 Road curves left to avoid an enlogate inselberg about 50 m high. Stop just before the curve and park along the north side of the road on the darkly varnished, interlocking stone pavement of the Sperry surface.

Stop 1 - Surface morphology and relations between Quaternary piedmont surfaces along the west margin of the Tecopa basin. North of the road, the Sperry surface extends several 100 meters off to the northeast towards the south flank of the Dublin Hills. South of the road, latest Pleistocene and Holocene alluvial surfaces of Greenwater Valley Wash are inset 1 to 3 m below the level of the Sperry surface.

The Holocene alluvial surfaces of the Tecopa basin are morphostratigraphically very similar to the Holocene surfaces along the west margin of Silver Lake approximately 70 km to the south (described by Wells et al., 1984). These surfaces are at most only slightly dissected. They occur within or adjacent to (and generally less than one to two meters above) active drainage channels, and they display prominent bar and swale relief of as much as 0.4 m. Stone pavements and rock varnish are absent on the youngest surface. However the development of these features progressively increases on intermediate and oldest Holocene surfaces; and many of the older Holocene surfaces support substantial areas of moderately to darkly varnished, interlocking stone pavements.

The Pleistocene alluvial surfaces of the Tecopa basin are morphologically distinct from the Holocene surfaces. All of the Pleistocene surfaces are characterized by broad, nearly flat interfluves covered with interlocking stone pavements. Non-carbonate clasts on these surfaces are typically darkly varnished. Interfluve rounding, piedmont-sourced drainage development, piedmont-sourced drainage dissection, and incorporation of secondary carbonate fragments in surface pavements generally increase with surface age on these Pleistocene surfaces.

In upper piedmont areas, younger surfaces are typically inset into older surfaces; however in middle and lower piedmont areas, younger deposits commonly overlap older surfaces above the high lake shoreline but are inset into the lake beds below this shoreline. Thus the several piedmont surfaces of the Tecopa basin generally converge at the approximate elevation of the highest Lake Tecopa shoreline and relations between these surfaces become locally complex.

Relations between the high lake shoreline, the erosion surfaces truncated by this shoreline or cut across the lake beds, and the alluvial surfaces graded to and inset within these erosion surfaces enable the reconstruction of an approximate middle and late Quaternary erosional history: (1) existance of a lake until late middle Pleistocene time (the highest shorelines of this lake attained an average elevation of about 545 m); (2) breaching of the basin to an outlet elevation of 440 m and draining of the lake between approximately 0.5 and 0.3 m.y.B.P.; (3) formation of a gently sloping, nearly featureless pediment (the Greenwater erosion surface) that cut across the deposits of the lake and graded to the 440 m level; (4) deepening of the outlet canyon to an elevation of 410 m followed by renewed dissection of the basin; and (5) formation of low straths and local pediments veneered by late Pleistocene and Holocene alluvial gravels during the late stages of this second pulse of dissection.

5.0 0.0 Turn around and return along the Greenwater Valley road towards the center of the basin. Proceeding down piedmont, the Sperry surface and the late Pleistocene alluvial surface (inset within the main channel of Greenwater Valley Wash) gradually diverge.

9.8 3.8 Junction with California 127. Turn right (south) onto California 127.

10.0 0.2 Intersection of California 127 and the Tecopa Hot Springs Road. Continue straight ahead on California 127.

12.9 2.9 Intersection of California 127 and the Old Spanish Trail Highway. Turn left (southeast) onto the Old Spanish Trail Highway.

13.8 1.9 Just before the road passes across the end of a low (15 to 20 m high) irregular ridge carved out of deeply dissected lake beds, turn right onto a dirt track and drive for approximately 200 m along the north side of the low ridge. Park along the track.

WARNING: STOP 2 IS A STEEP, RUGGED, UNSTABLE AREA. PLEASE EXERCISE EXTREME CAUTION WHEN HIKING AND CLIMBING IN THE VICINITY OF THIS STOP.

Stop 2 - Degradational landscapes produced by concentrated subsurface flow processes (piping) in the lacustrine deposits of the Tecopa Basin.

Post-lake dissection of the lacustrine deposits of the Tecopa basin has been strongly affected by piping processes. Indications of long-continuing piping are widely and abundantly distributed across large areas of the basin. Although somewhat limited in area, the irregular ridge south of the track is a classic example of a landscape in the advanced stages of dissection by piping processes. This ridge is riddled with: (1) sinkholes as much as 30 m wide and 10 m deep, (2) vertical pipes up to 2 m in diameter and at least 15 m deep, (3) blind valleys as much as 15 m deep and several 100 m long, and (4) tunnels up to 2 m high and 4 m wide. These large scale piping landforms have coalesced to form a chaotic landscape of irregular low ridges and hills lacking integrated surface drainage. Similar landscapes have also developed in several other areas of the basin.

Several geologic, geomorphic, and climatic factors combine within the Tecopa basin to produce the necessary conditions for large-scale piping: (a) subhorizontally bedded, poorly to moderately indurated lacustrine deposits with low intrinsic permeabilities, (b) high infiltration capacities produced by extensive cracking, (c) high silt-clay contents with high percentages of expandable clays, (d) high potential evapotranspiration coupled with low mean annual rainfall and the occasional occurance of high intensity, short duration storms, (e) deep dissection providing abundant free faces and steep hydraulic gradients adjacent to these free faces, and (f) strongly indurated layers of carbonate-cemented mudstone which form and protect the roofs of the larger pipes. Where these factors come together, concentrated subsurface flow processes locally outstrip overland flow processes and large-scale piping landforms dominate the landscape.

13.8 0.0 Return to the vehicles. End of the third day field guide.

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MID-PLEISTOCENE OVERFLOW OF DEATH VALLEY TOWARD THE COLORADO RIVER

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ABSTRACT

A large fluvial channel incised in bedrock near Ludlow, California, is thought to be the product of overflow of a large Mid-Pleistocene Death Valley lake. The channel is located 2.6 km southeast of Ash Hill Siding beside the Santa Fe Railroad. Fluvial (as opposed to debris flow) discharge through this channel is suggested by imbrication of boulders of more than one meter diameter; existence of a large expansion bar, a pendant bar, riffles and pools; high water lines preserved on a long inactive fan surface; minimal rounding of large clasts; and comparisons of channel and fan particle size distribution. That discharge was from Ash Hill Siding rather than the hills nearby is suggested by the following: The fan is steeper than the bedrock channel, yet the channel is essentially free of fan debris; estimate of maximum channel discharge is unusually great (2000 m^3/sec based on boulder size and bankfull cross section which is seen as unlikely from the 18 km^2 watershed nearby when the world record for that size watershed is 650 m^3/sec); and a convex fan profile is consistent with removal at the toe by an exotic source. Elapse of considerable time since the channel was fully active is suggested by the following: Weathering of bedrock and transported boulders; exceptional (5+ m) depth and development of caliche over whole Ash Hill fan which is graded in its lower reaches to the floor of the incised fluvial channel and appears to have been beheaded not long before channel flow ceased.

Direct evidence for the hypothesized lake inside the Death Valley pluvial watershed includes weathered but clearly rounded pebbles at spillway elevation (1950 ft) on the rim of the smaller volcanic vent of the Ash Hill Basalt 3 km northwest of Ash Hill Siding (K-Ar age 5.5 my). In this location pebble rounding by means other than by wave action appears impossible. Similar rounding is not apparent 20 m higher on top of the larger Ash Hill vent.

The exact size and shape of the lake that would overflow at Ash Hill are not known. Calculations suggest this lake and its smaller tributary lake surfaces totalled about 12,000 km^2 or twice that of Lake Manly and its tributaries. The large lake would have extended 300 km from Ash Hill to the north end of Death Valley. Comparison of necessary precipitation and evaporation ratios suggests that if evaporation were the same for both the Manly and hypothesized lake systems, 100 mm precipitation per year beyond that which sustained Manly would fill and spill the hypothesized lake at Ash Hill.

Basing evaporation depression on the modern record at Mojave, California, mean watershed precipitation demanded by the hypothesized lake system is estimated to have been either, 660, 490, 330, or 230 mm at temperature depressions of 3, 5, 7, or 8°c from the present, respectively. The counterpart figures estimated for the Lake Manly system are 560, 430, 290, or 210 mm, respectively. Modern mean precipitation and evaporation are estimated to be 200 mm and 1600 mm, respectively.

A Mid-Pleistocene--probably Sherwin Glacial--age is postulated based on circumstantial evidence including rock weathering, fan age, survival of fluvial features, and the recent literature. The latter includes: (1) Quaternary faulting studies by Bull in the Mojave Desert suggesting tectonic stability of the Ash Hill site for several million years; (2) Examination of cores by Smith and others at Searles Lake suggesting a deep, fresh lake with frequent overflows from 1.28-1.0 M.Y.B.P. (correlated by them with the Sherwin Glaciation); (3) A stratigraphic column by Sheppard and Gude combined with volcanic ash dates by Sarna-Wojcicki and others for Tecopa Basin showing an essentially continuous lake there from 2-.6 M.Y.B.P; (4) Estimates by Hoover, Hay, and Hillhouse of Tecopa Basin climates from depositional environments (200-500 mm annual precipitation sometimes greater or less; 10°-18°C temperature for most of the Pleistocene); (5) The paleogeographic reconstructions of Huber and of Curry in the Central Sierra Nevada where uplift in the last million years is inferred to be 300 m or 600 m, respectively; (6) Earthquake wave studies by Hadley and Kanamori and gravity studies by Grannell suggesting absence of deep mountain roots under the Transverse Ranges of California which may mean they were significantly lower during the Mid-Pleistocene; (7) The work of Ehlig and of Crowell who have demonstrated 240-250 km northwestward movement of the San Gabriel Mountains during the last 4 M.Y. (interpolate 60 km in 1 M.Y.). Together, these recent contributions serve to: (1) provide a time (Mid-Pleistocene) when other parts of the greater Death Valley watershed were under pluvial conditions; (2) provide a mechanism (reduced rainshadow barriers) by which the additional precipitation required by an overflowing Death Valley lake could be obtained; (3) establish the tectonic stability of the overflow site so that its evidence may be interpreted as it lies.

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INTRODUCTION

More than a century has passed since the first discoveries that large lakes once occupied the internally drained basins of the now arid Great Basin (See Whitney, 1865; Simpson, 1876; and Gilbert, 1890). Not long after the initial discoveries, it was realized (Gilbert, 1890) that the large lake in the Bonneville Basin had received sufficient runoff to overflow the confines of its basin and spill to the sea. Further examination showed abundant evidence of Pleistocene large lakes in the Lahontan Basin of Nevada and in the smaller Mono Basin of Eastern California (Russell, 1889).

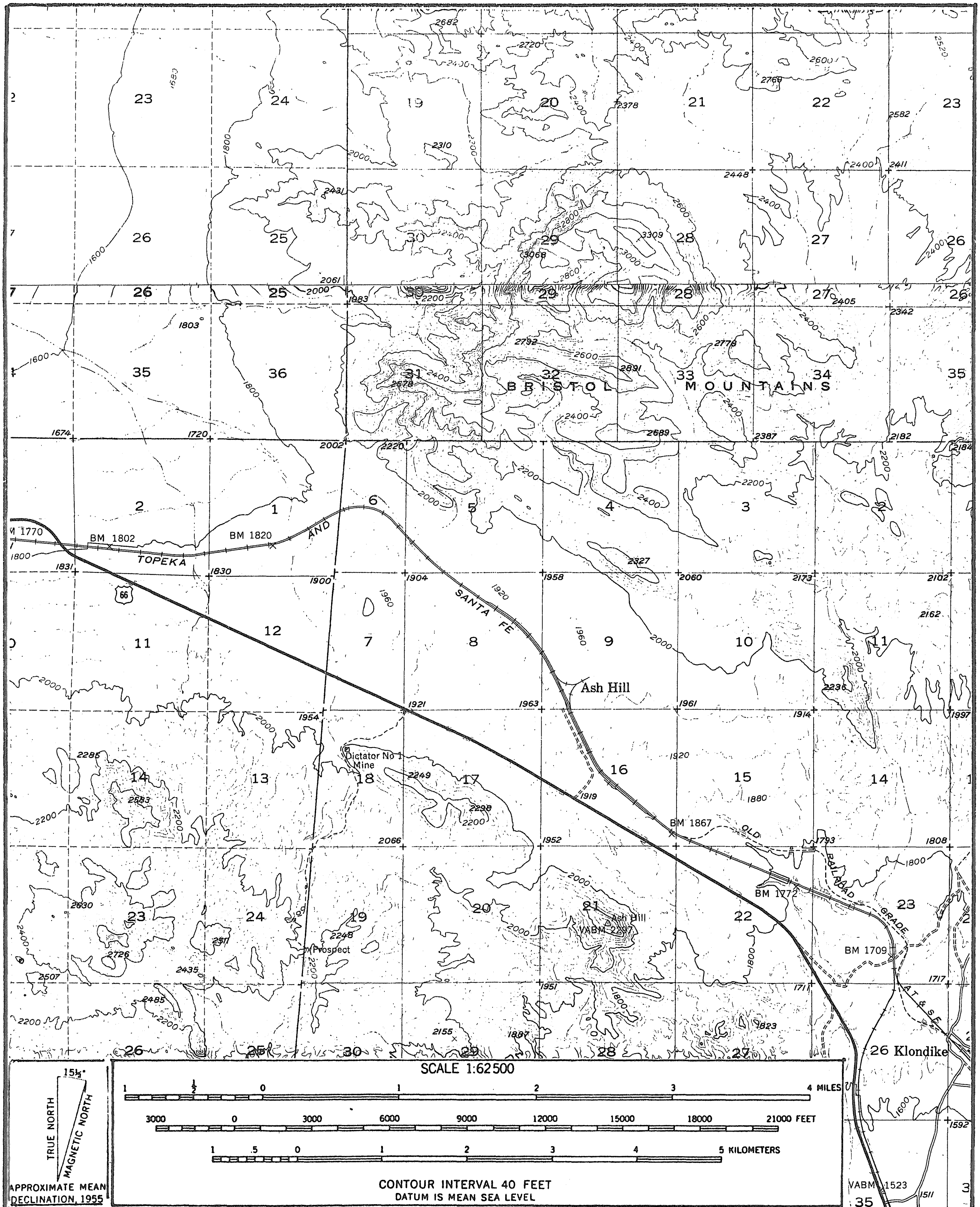
However, south of the latter, the evidence of extinct lakes was not so clear. Despite references to a large Pleistocene lake by Russell (1885, 1889), Gilbert (1890), and Bailey (1902), it was not until 1926 that Noble (1926, p. 69) discovered strand lines on what was later named Shoreline Butte and near Mormon Point in Death Valley, confirming the Pleistocene existence of a large lake. This lake was referred to as Lake Manly by Means (1932) and by Blackwelder (1933). Its maximum elevation was approximately 95 m.

The discovery that Lake Bonneville had overflowed brought up the question of whether or not the same had happened in Death Valley. Noble's strand lines are more than 400 m below the elevation where the Death Valley Basin would overflow were it filled today. Three points argued strongly against overflow of Pleistocene Lake Manly. First, the faintness of the Death Valley shorelines gave the impression of a short-lived and thus climatically marginal lake there. Second, Death Valley's position at both a lower latitude and a lower elevation than the Bonneville Basin made a large lake appear unlikely. Third, the apparent absence of shore features at, or close to, overflow elevation argued convincingly against an overflowing lake in the Death Valley Basin--especially when compared to the abundance of shore features in the Bonneville Basin.

Blackwelder and Ellsworth (1936) speculated about the possibility of Death Valley overflow based upon stratigraphy of lacustrine sediments in the Manix Lake Basin. Blackwelder (1954) further speculated that quiet water sediments at the north end of Death Valley might have been deposited in an overflowing Death Valley lake associated with the mid-Pleistocene Sherwin Glaciation of the Sierra Nevada. Hubbs and Miller (1948) and Miller (1981) argued for a former hydrologic connection between the Death Valley Watershed and that of the Colorado River based upon the distribution and form of fossil fishes. Bassett and Muessig (1959) reported lacustrine clays in the Soda Lake Basin. Barring minor crustal warpage, these would signal the former existence of a 250 km long, 400 m deep lake filling much of the Death Valley Rift. Butler (1982) reported shorelines in Death Valley above 275 m elevation on the northern foothills of the Avawatz Mountains.

PURPOSE

The principal purpose of this study is to determine whether or not a large lake overflowing toward the Colorado River occupied the Death Valley Basin at some time during the Pleistocene. If such an overflow can be demonstrated, the second purpose is to examine the evidence for its timing.



Topographic map
of Ash Hill Trough

Plate 1 shows the approximate outline of the hypothetical large lake that would overflow at the present basin sill at 594 m elevation. Also shown are outlines of the maximum Pleistocene Extent of other Pleistocene lakes in the Great Basin as interpreted by Snyder and others (1964).

The search for geomorphic evidence of an overflowing lake in the Death Valley Basin has led to the site where overflow would have occurred. If it is assumed that the present topography is unchanged from that of the past, spillage of such a lake would have occurred at Ash Hill in California's Mojave Desert. Ash Hill is 10 km east of the town of Ludlow at the watershed divide between the Bristol and Broadwell basins. The Ash Hill Divide lies at an elevation of 594 m in the middle of a broad structural trough connecting the Bristol and Broadwell basins. A large channel cut in basaltic bedrock, but apparently inactive at present, leads southeastward from this divide.

Three kilometers west of this divide is a pair of volcanic vents which would have been small, wave-worked islands protruding from the surface of any lake that would overflow at Ash Hill.

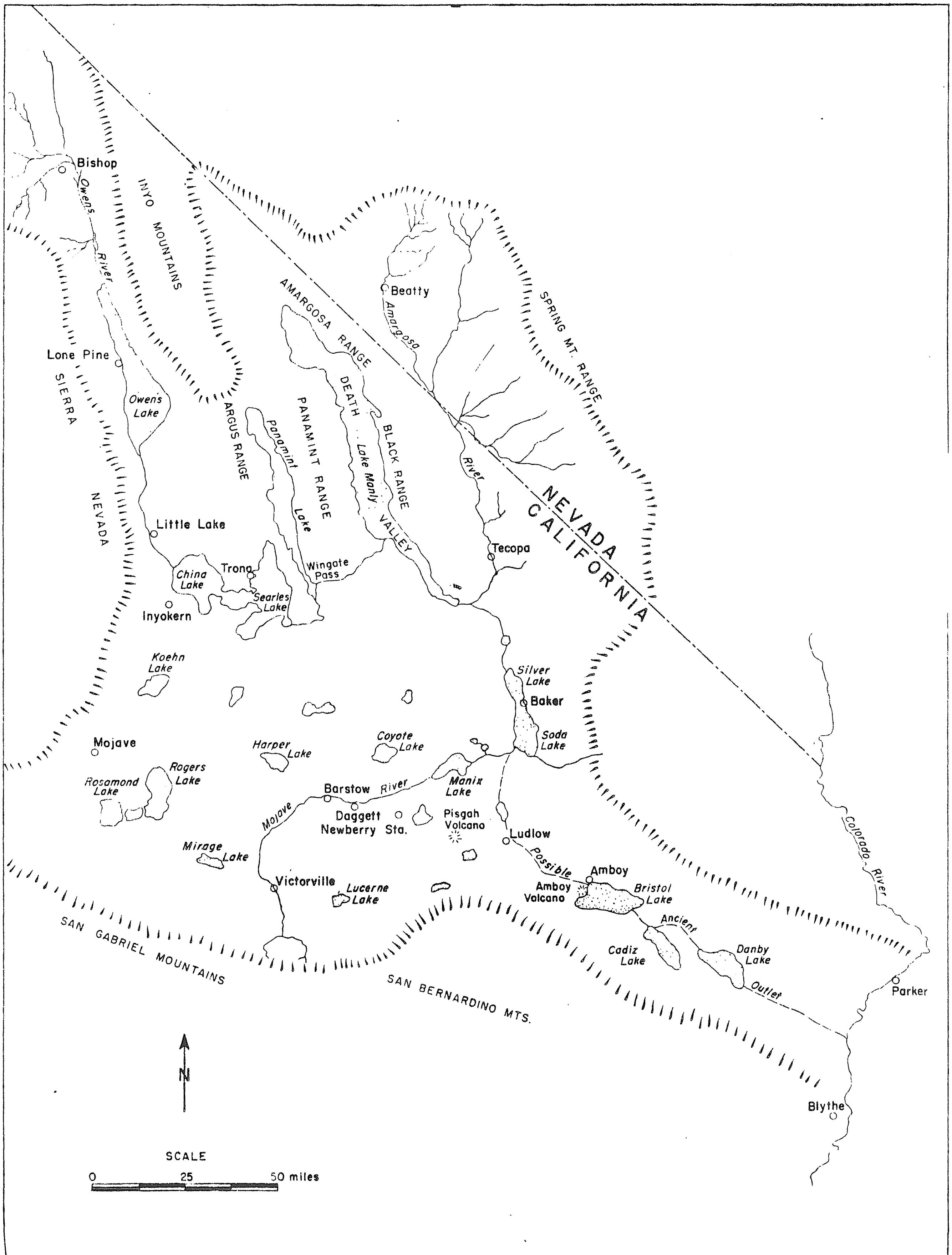
In addition, a thorough search of the overflow shore elevations has revealed a dozen sites where quiet water sediments may be found. Several of these are situated so that their origin is difficult to explain by means other than those of the hypothesized lake. The most intriguing of these is in the Red Pass Valley 63 km north of Ash Hill.

GEOLOGY OF ASH HILL

The broad trough that contains Ash Hill is underlain by volcanic rocks, older and younger alluvium, and aeolian sand. The area is depicted on the Ludlow 15' and Ash Hill 7.5' topographic quadrangles (USGS, 1955). The geology of this qûadrangle has been mapped by Dibblee (1967). Several points are pertinent to the present study.

First, most of the area is underlain by volcanic rocks of probable mid-Tertiary age. These rocks dip moderately and consistently to the southwest. The hills prominent in the landscape north and south of the trough are interpreted by Dibblee (1967) as remnants of erosionally resistant strata within a thick, tilted section of these rocks. In this interpretation, the Ash Hill Trough is seen as an eroded, less resistant stratum of the thick tilted section.

Lying unconformably and undeformed on these older volcanic rocks is the Basalt of Ash Hill. This extrusive rock covers the lower elevation parts of the trough. It appears to have issued from vents in the vicinity of Ash Hill and to have flowed to the southeast. It forms a resistant cap layer that covers much of the trough from west of its present crest at Ash Hill Divide to a point roughtly 6 km downstream on the east. The previously-mentioned bedrock channel is incised into this basalt. This basalt is $5.56 \pm .44$ K-AR MA (Turrin, unpublished data, 1983). Four parallel ridges of basalt near Ash Hill are believed to be dikes of the basalt filling fissures through which the flows erupted (Dibblee, 1967).



BLACKWELDER'S MAP OF PLEISTOCENE PLUVIAL LAKES OF THE DEATH VALLEY WATERSHED SHOWING "POSSIBLE ANCIENT OUTLET" (1954)

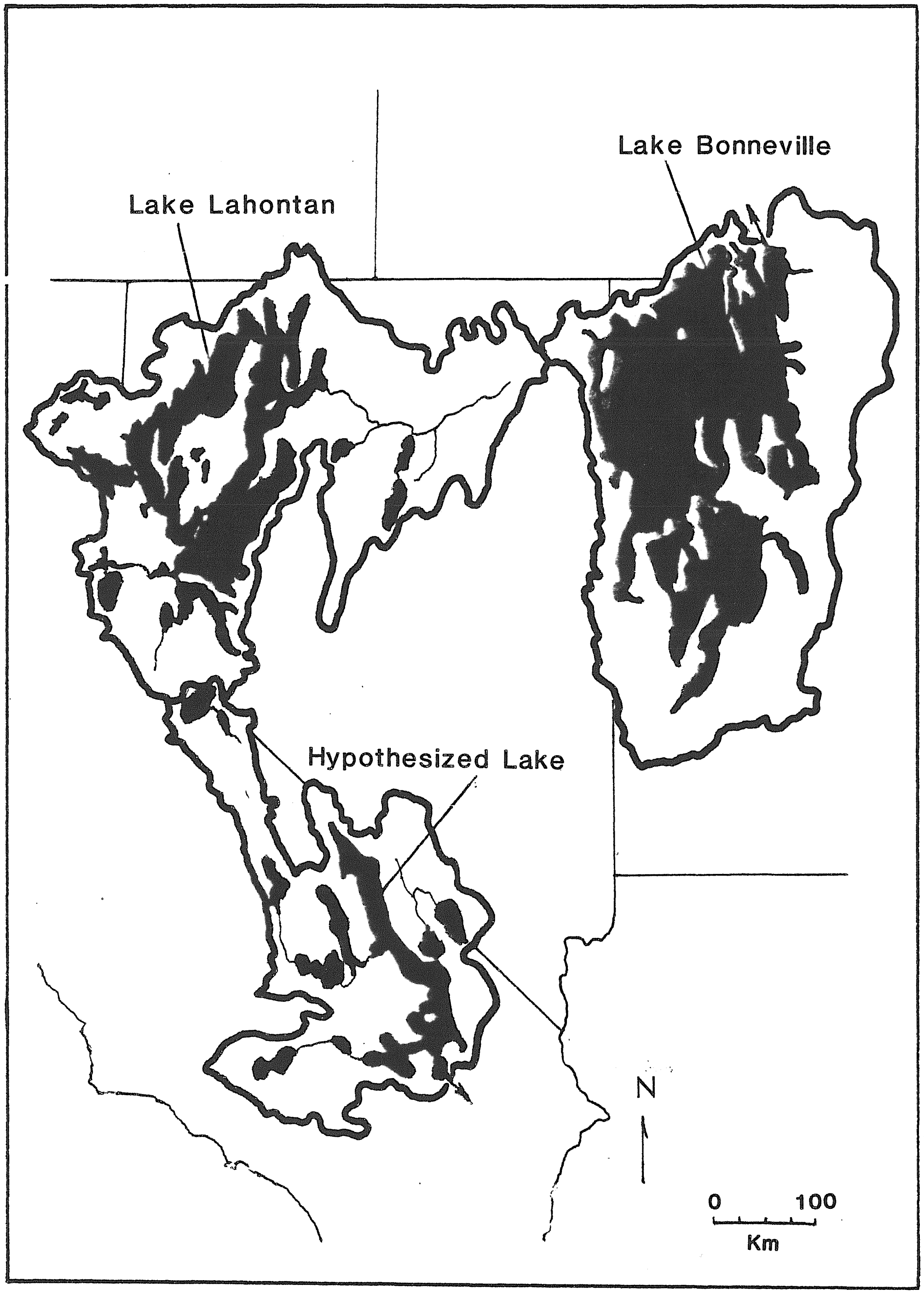


PLATE 1. MAXIMUM EXTENT OF PLEISTOCENE PLUVIAL LAKES IN THE GREAT BASIN INCLUDING LARGE LAKE HYPOTHESIZED IN THE PRESENT REPORT.

On the westernmost of these ridges are two small eruptive vents. The rim of the larger vent stands 50 m above the alluvium of Broadwell Basin to the west.

Older alluvial fan material that has come from the hills to the north and south of the Ash Hill Divide lies unconformably on top of the Ash Hill Basalt (Dibblee, 1967). While younger than the Ash Hill Basalt, the Ash Hill Alluvial Fan on the north side of the trough appears to be quite old. The age of this fan will be discussed later in this paper.

Bedrock fracturing is widespread in the Basalt of Ash Hill. Nowhere is there visible an unbroken expanse of this volcanic rock greater than 3 m. At the same time, however, there is no sign of faulting anywhere in the lithologic unit. Dibblee's (1967) investigation confirms this. Extensive studies of Quaternary faulting by Bull (personal communication, 1983) indicate that the Mojave Desert east of the Ludlow Fault is an area of "tectonic quiescence" during the last several million years. Little, if any, evidence for Quaternary faulting can be found east of the Ludlow Fault in the Mojave Desert. The Ludlow Fault itself exhibits displacement that is minor when compared with the relatively great fault activity farther west in the middle of the Mojave tectonic block. Although the Ludlow Fault passes within 8 km of the Ash Hill Trough, Bull believes that the watershed divide at Ash Hill Siding is a "...stable indicator relative to other tectonically stable sites for several million years."

The appearance of the Ash Hill area is deceptive because the Basalt of Ash Hill and older fan materials are covered by a thin veneer of aeolian sand. Standing on a small knoll overlooking the crest of the watershed divide at Ash Hill one does not have a sense of the considerable erosional resistance of the basalt-capped saddle itself. In fact, the opposite seems true. The appearance of the area is that the slightest bit of runoff would rapidly erode what looks like a fine, sandy fanglomerate that seems to comprise the floor of the trough.

GEOMORPHIC EVIDENCE AT ASH HILL

Beginning 2.6 km southeast of the spillway directly in the path that would be followed by overflow from a large Death Valley lake, is a 3.0-km-long channel incised as much as 10 m below the surface of the Basalt of Ash Hill. Where incised in basalt the width of this channel varies from 25 m to 40 m. This channel appears to have been formed by a discharge of water much greater than that likely to be generated within the watershed that presently drains toward it. The following sections evaluate the possibility that this channel was cut by water discharging from a large lake that overflowed at Ash Hill, as opposed to being a continuation of channeled flows originating in the Bristol Mountains nearby.

Ash Hill Bedrock Channel

The Ash Hill bedrock channel can be seen in Plate 2 crossing from the left center to the lower center of the plate. Runoff first enters a hook-shaped segment near the left margin of the photograph. Former active channels on the

PLATE 2. OBLIQUE AIR PHOTO OF ASH HILL CHANNEL SHOWING STRAIGHT REACH AND EXPANSION BAR. FEATURES ARE IDENTIFIED IN CARTOON ON FOLLOWING PAGE. THEODORE M. OBERLANDER PHOTO.

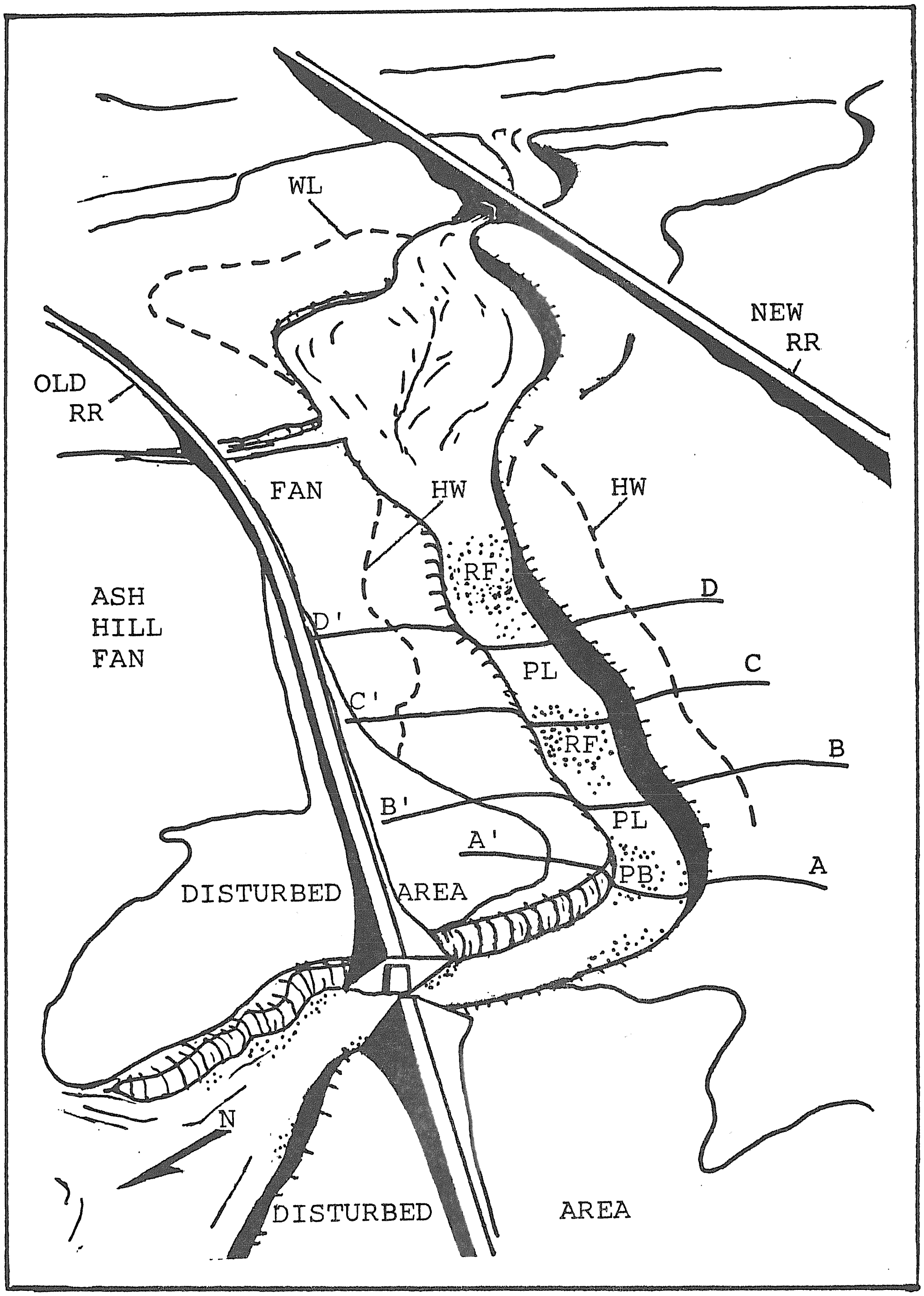


Fig. 1 Schematic drawing of Straight Reach and Expansion Bar showing several High Water Lines (WL and HW), locations of channel Cross Sections A-A' through F-F', Riffles (RF), Pools (PL), and Point Bar (PB). Discharge and velocity estimated at C-C'.

Ash Hill Fan enter this hook-shaped reach at right angles. After leaving the hook-shaped bedrock bend, the channel turns sharply into a .4-km-long bedrock straight reach. At the end of the straight reach, it widens into an expansion reach containing a large center bar. Downstream from this point, there is a series of constrictions and expansions as the channel crosses the irregular surface of the Ash Hill Basalt.

The Ash Hill Channel is incised deeply into the Ash Hill Basalt in several places. Its maximum depth is more than 10 m (33 ft.). The bed slope of the channel in the straight reach averages .015 (m/m) deviating very little from this figure. The channel averages 36 m width in the straight reach and is notable for its consistency of cross section there.

Riffles and Pools

Two gravel bars, or riffles, are observable in the straight reach. These are depicted in Figure 1 and Plates 2 and 3. Cross Section C-C' passes through the center of one. It is located 120 m downstream from the sharp bend and point bar that mark the beginning of the straight reach. The second is located approximately 150 m downstream from the first. Upstream from each of these riffles is a bedrock pool. These pools are depicted in Figure 1 and Plate 3. If channel width is taken at the point of bank inflection at Cross Section C-C', the channel is 30 m wide. The riffle spacing is, therefore, between 4 and 5 times channel width. This is generally consistent with values for the geometry of channels in alluvial floodplans given by Leopold and others (1964).

Expansion Bar

An unmistakable fluvial expansion bar can be found in the Ash Hill Channel. It is 400 m long, 150 m wide, and lies immediately downstream from the straight reach. The expansion bar has the dimensions and form of central bars produced in laboratory flumes. See Leopold, et. al. (1964) and Baker (1973).

Inspection of Plate 3 reveals a subtle, sinuous terrace (labelled WL in Figure 1). Precise leveling of this terrace and its associated scour marks in the field shows it to be at the same elevation and to have the same downstream gradient as the top of the opposite channel bank. Because of its contour-like character and its relationship to the opposite bank, this terrace is interpreted to be a fluvial high water line.

High Water Marks

The water surface slope at bankfull stage in the straight reach can be inferred from apparent high water marks seen on top of the channel banks. One pair of these is visible in oblique air photo (Plate 3). These marks are drawn in Figure 1 as dashed lines labelled HW. The mark on the south bank can be seen as a subtle boundary between differing surface textures and coloration. Informal field observations show that below the mark the desert pavement surface includes many cobbles and small boulders that protrude, whereas above it such is not the case. Many more rocks below the mark are, by virtue of color or compositional differences, unmistakably transported. That is, each shows some difference from the weathered basalt bedrock beneath. Although essentially all rocks in the area are volcanic and most are basaltic, there are noticeable differences in the tint of weathering rinds on clast surfaces from various upstream locations. These subtle color differences are much more apparent below the boundary than above it, suggesting transport from a variety of upstream sources. Also, occasional clasts of pink rhyolite are found below the boundary. This rock crops out 2½ km to the north of the Bristol Mountains. The pattern that these clasts make when seen from the air favors interpretation as fluvial deposits. The upper boundary of the depositional zone just described parallels the channel banks. It is essentially level with the opposite channel bank top, and its slope is the same as that of the channel bed.

At channel cross sections B-B', C-C', and D-D' (Figure 1), the elevations of the upper boundary of this depositional zone on the south bank correspond with another mark (HW) that can be seen on the flat area at the top of the opposite bank (Figure 1).

These high water marks on both banks of the straight reach are thought to have been formed by one flow.

Discharge in the Outlet Channel

Estimation of peak channel discharge is possible using Manning's equation for a reach characterized by uniform flow when hydrologic radius, channel cross section, and velocity are known or can be inferred.

Uniform flow in the straight reach at Ash Hill is suggested by similarity between channel bed and water surface slopes and by uniformity of channel cross-section areas. The dimensionless channel bed slope (fall per unit of horizontal distance) is .015. Minor deviations (.003 or less over 30 m or less) from this figure occur at the upstream and downstream margins of the two riffles described earlier.

Discharge is estimated at Cross Section C-C' (Figures 1 and 2). At this point, flow appears to have been essentially uniform. Evidence for this is taken from channel slopes and cross sections. The channel depth from the bed to the high water line at Cross Sections B-B' and C-C' is in each case 8.8 m. The water surface slope between these two sections along the high water line (HW) is therefore the same .015 that was measured for the channel bed. Water depth at Section D-D' is 8.6 m. This steepening of the water surface over the 50 m separating Cross Sections C-C' and D-D' adds .004 to the water surface slope, making it .019.

The second suggestion of uniform flow is found in the channel cross sectional areas at B-B', C-C', and D-D'. Areas of the cross sections shown in Figures 1 and 2 are the average of three repeated measurements, each taken with an electronic planimeter of cross sections measured in the field with an 8-inch Dumpy Level. These cross sections are plotted in Figure 2. The average values are 391 m^2 for B-B', 389 m^2 for C-C', and 392 m^2 for D-D'. The difference among these are of the same magnitude as the difference between repeated measurements of the same cross section. All three sections, therefore, are considered to be

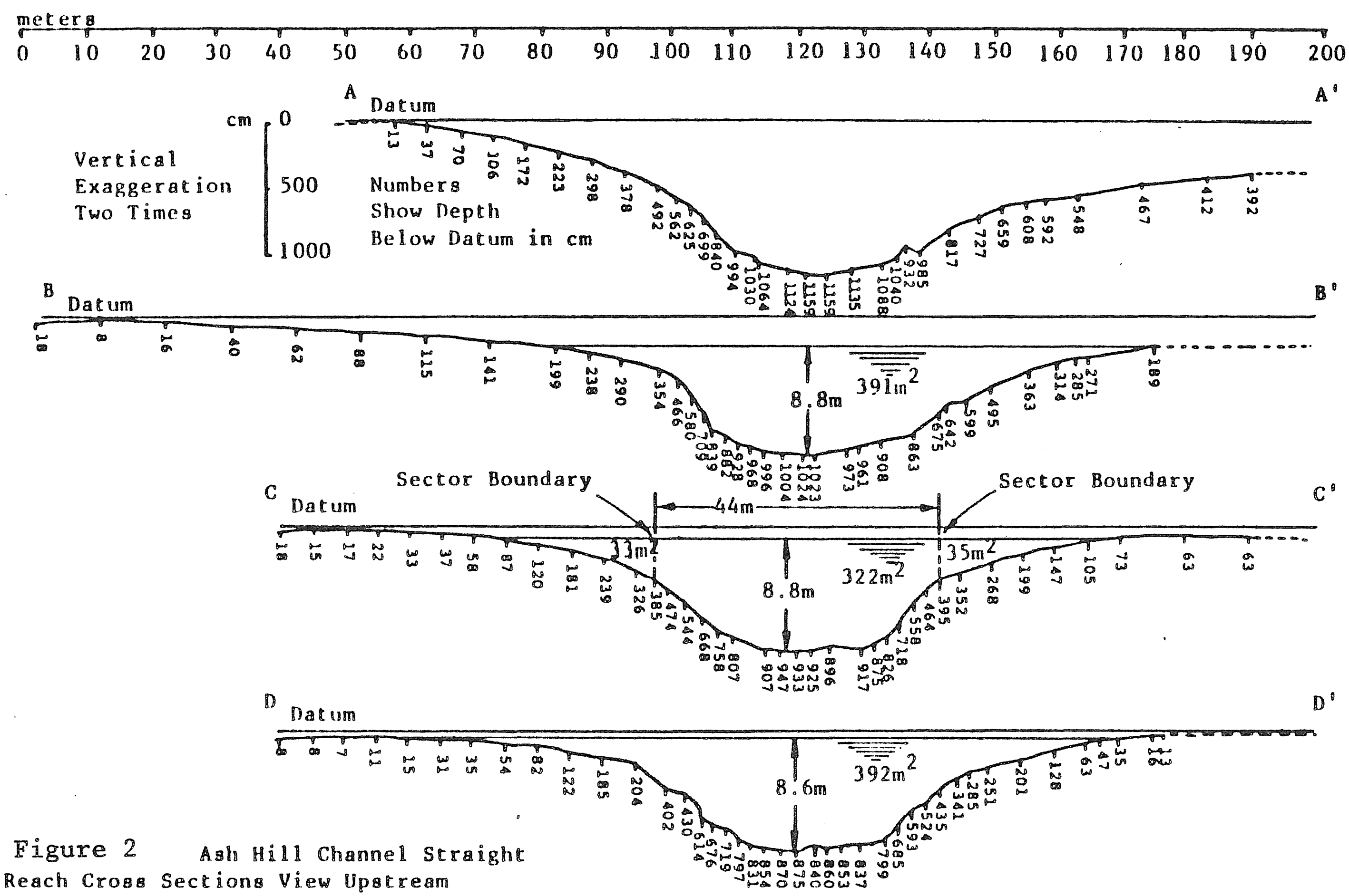


Figure 2 Ash Hill Channel Straight Reach Cross Sections View Upstream

the same--equalling approximately 390 sq. m. Consistency of cross sectional area, combined with the previously discussed similarity and constancy of water surface and channel bed slopes constitutes evidence for uniform flow--especially in the upstream third of the straight reach (that lying above Section C-C'). Criteria for indirect measurement of paleoflows in channels are presented by Costa (1983).

Mean Velocity at Peak Discharge

If the channel cross section includes one or several wide, shallow sectors, the river-crossing section is generally divided into several components.

Component sectors of the channel are established by dropping a vertical boundary from the water surface to the channel bed at the edge of each shallow zone. Separate velocity and discharge estimates are made for each of the sectors thus produced. Cross Section C-C' has two shallow sectors--one at each bank. If the boundaries of these are taken at the sharpest convexities in the channel bank slope (See Cross Section C-C'), the wetted perimeter of the central sector is 44 m. The cross sectional area of the central sector is 322 m^2. This gives a hydraulic radius (R) of 7 m. Flanking the center sector are two shallow panhandle sectors--one having an area of 33 m^2 and a wetted perimeter of 24 m (R = 1.38 m) and the other having an area of 35 m^2 and a wetted perimeter of 25 m (R = 1.40 m).

The ideal parameter for slope (S) in the Manning equation is the energy grade line of water in the channel (Leopold and others, 1964, p. 157). This is a value that is not directly available for ancient flows. However, the water surface slope is generally considered to be a reasonable approximation of the energy grade line (Baker, 1973, p. 23, and Leopold and others, 1964, p. 157). If water surface slope is not available, the channel bed slope is sometimes used as an alternate approximation (Baker, 1973, p. 23, and 1975, p. 975). For the present study, the close similarity between channel bed and water surface slopes and the similarity among cross section areas are taken as indications that the energy grade line was parallel to the other two measures of slope. The .015 value common to both will be used.

Manning's n is difficult to estimate. It is based on subjective interpretations of the appearance of roughness elements in the channel. Estimation of its value is usually made either from tables linking n values to verbal descriptions of roughness (See Chow, 1964, pp. 7-24) or from stereographic photographs (Leopold and others, 1964, p. 158). Although n is usually between .01 and .07, Leopold (personal communication, 1983) reports using values as high as .10 in cases of extreme channel roughness. Costa (1983) also reports empirical values greater than .1 for n in steep mountain channels where visual estimates would suggest lower values.

The straight reach channel bank roughness is great. The large basalt joint blocks jutting into the passing current would have applied considerable braking action to water in the channel. For these reasons n is assumed to be .07 in the straight reach.

Substituting the previous values for the central sector into Manning's equation,

$$\bar{V} = \frac{(7.0)^{\frac{2}{3}} \ (.015)^{\frac{1}{2}}}{.07}$$

$$\bar{V} = 6.4 \text{ m/sec}$$

Substituting the previous values for each of the panhandle sectors into Manning's equation,

$$\bar{V} = \frac{(1.38)^{\frac{2}{3}} \ (.015)^{\frac{1}{2}}}{.07}$$

$$\bar{V} = 2.2 \text{ m/sec}$$

and,

$$\bar{V} = \frac{(1.40)^{\frac{2}{3}} \ (.015)^{\frac{1}{2}}}{.07}$$

$$\bar{V} = 2.2 \text{ m/sec}$$

The mean velocity in the central sector of the channel at the depth indicated by high water lines is therefore estimated to have been 6.4 m/sec. The velocities in the two shallow panhandle sectors at the same time were much smaller, each being 2.2 m/sec.

Peak Discharge

Discharge is the product of velocity and cross section area for each sector of the stream. At Cross Section C-C', the estimated peak discharge for the central sector is

$$(6.4 \text{ m/sec})\ (322 \text{ m}^2) = 2062 \text{ m}^3/\text{sec}$$

The discharge estimated for the panhandle sectors at peak flow is

$$(2.2 \text{ m/sec})\ (33 \text{ m}^2) = 73 \text{ m}^3/\text{sec}$$

and

$$(2.2 \text{ m/sec})\ (35 \text{ m}^2) = 77 \text{ m}^3/\text{sec}$$

Together the panhandle sectors equal about 7 percent of the peak discharge for the central sector.

These figures are smaller than the likely error introduced into the calculations by uncertainties with respect to Manning's n. For this reason they will not be added to the figure for peak discharge in the central sector.

The central sector peak discharge figure (2062 m^3/sec) will also be rounded to 2000 m^3/sec because of the same uncertainty.

Channel Bed Clasts

The diameters of the largest stranded clasts on the channel bed tend to corroborate the contention based on high water marks that water depth in the channel at Cross Section C-C' was at one time on the order of 9 m.

Numerous studies have been made of the various relationships among depth, shear stress, velocity, and largest clast size (See especially Scott and Gravlee, 1968; Birkeland, 1968; Baker 1973; Baker and Ritter, 1975; Costa, 1983; and Andrews, 1983). Clast-based estimates of former water depth for the Ash Hill Channel made using criteria presented in these studies vary by almost an order of magnitude. However, the average of the largest and smallest of these (2.4 m and 13 m) approximates the depth interpreted from high water marks at Cross Section C-C' (8.8 m). At this site the intermediate diameter of the largest clast is 1140 mm. The slope is .015. The largest depth estimate is obtained from the regression line of Baker and Ritter (1975, p. 976). It suggests a shear stress of 201 kg/m^2. When substituted into DuBoys' equation, this figure produces a water depth estimate of 13 m. On the other hand, recent work by Andrews (1983) and by Costa (1983) suggests that water depths as small as 2.4 m or 2.8 m could move a boulder of that size. The smallest estimate comes from Andrews' (1983) finding that where the ratio of clast diameter to median subsurface particle diameter is larger than 5, the critical dimensionless shear stress approaches a constant value of .020. From this, a critical shear stress of 36.5 kg/m^2 is derived for Cross Section C-C'. Substitution into Du Boys' equation suggests that 2.4 m depth of water at Cross Section C-C' could possibly have moved the largest clast there. For this estimate the bedrock channel is assumed to be equivalent to an alluvial bed with a very small median subsurface particle diameter.

A similarly small depth estimate of 2.8 m is taken directly from a curve presented by Costa (1983, Figure 7). From this curve a clast diameter of 1140 mm at a slope of .015 is consistent with a depth of about 2.8 m. Baker and Ritter (1975, p. 977), Andrews (1983, p. 1229), and Costa (1983, p. 999) discuss several factors that may contribute to this variety of depth estimates.

Because of the considerable magnitude of difference among these depth estimates for the same event, clast size will not be used for a precise estimate of maximum water depth. The depth estimates (2.4 m to 13 m) based on boulder sizes are, however, useful to corroborate the magnitude validity of the depth estimate taken from high water marks (8.8 m) at Cross Section C-C'.

SOURCE OF DISCHARGE

Former Nearby Tributary Watersheds

The area of the watershed tributary to the incised bedrock channel is presently quite small when compared to its probable extent prior to incision of the channel. The area of the present watershed is 4.3 km^2. An additional contiguous watershed area of 14.1 km^2, formerly tributary to the bedrock channel appears to have been diverted away from the straight reach as the Ash Hill Fan was beheaded. This makes the maximum probable size of the watershed prior to capture 18.4 km^2. Even this larger watershed area appears insufficient to have produced 2000 m^3/sec of runoff.

Equivalent Rainfall in Nearby Watersheds

An unprecedented deluge would be necessary to produce a flow of 2000 m^3/sec from an 18.4 km^2 watershed. If it is assumed as follows that: (1) the soil surface was saturated so no infiltration could occur, causing all of the precipitation to run off; (2) evaporation was non-existent; and (3) precipitation was uniformly distributed throughout the watershed, precipitation at the rate of 397 mm/hr (16 in/hr) would have been necessary to produce the estimated discharge.

Comparison with Greatest Flood Discharges

Hoyt and Langbein (1955, Figure 3) have plotted greatest measured and estimated flood discharges for the priod of record against drainage areas for the United States. The curve shown for this plot suggests that the maximum likely discharge for an 18.4 km^2 watershed is on the order of 700 m^3/sec (25000 ft^3/sec). The peak discharge estimated from geomorphic evidence for Cross Section C-C' is nearly three times this figure.

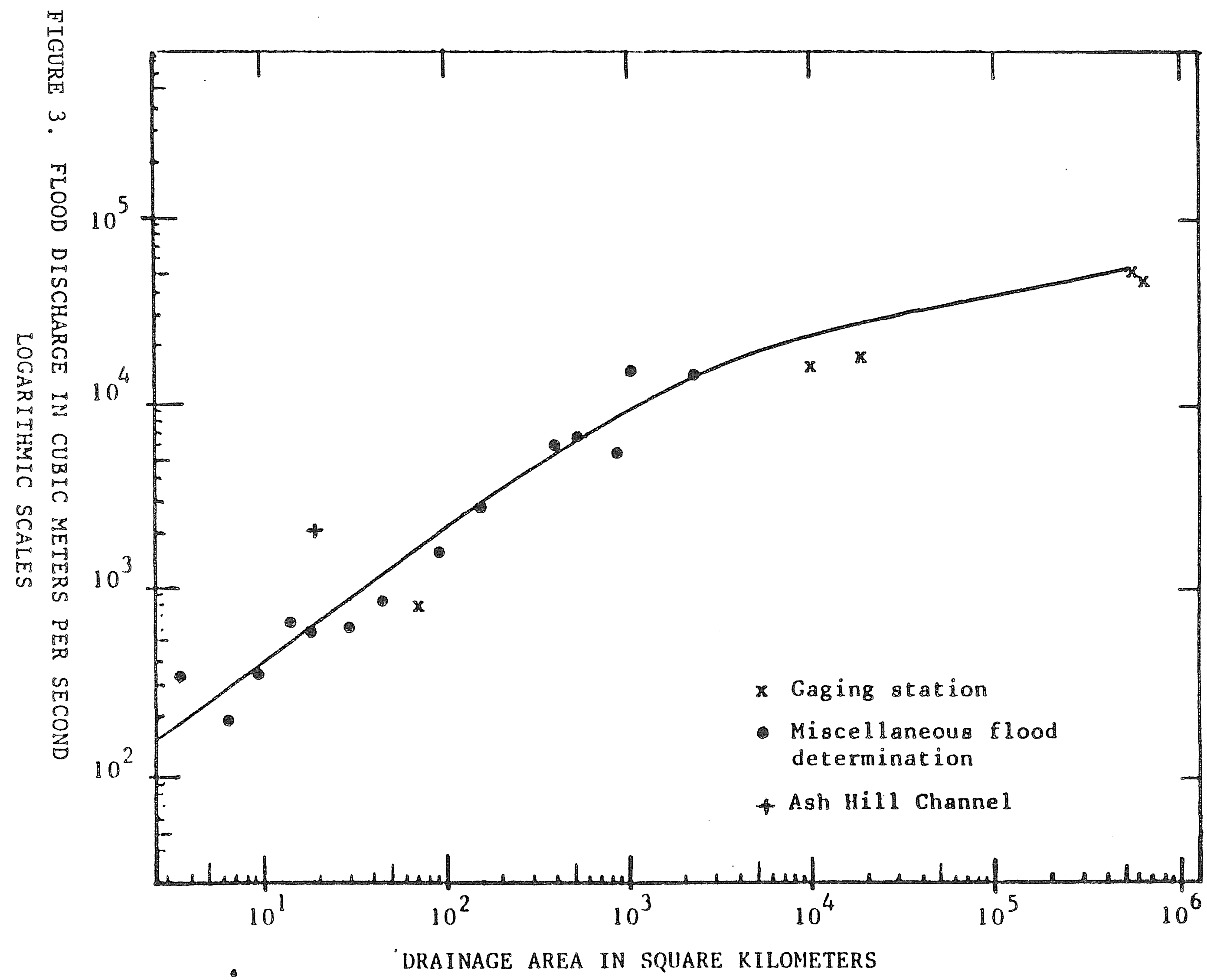
Comparison with Mojave River Flood

The estimated peak discharge (2000 m^3/sec) through the straight reach of the Ash Hill Channel is the same as that for the greatest flood in the period of record for the 1330-sq-km watershed of the upper Mojave River. That flood was estimated to be 2000 m^3/sec (70600 ft^3/sec) at the Victorville Narrows on March 2, 1938 (Grover, 1939). The recurrence interval for this flood based on a 46-year period of record (1931-1976) is reported by Kahrl and others (1979, p. 72) to be 25 years. The magnitude of the 1938 flood was roughly twice that of the next largest flood on the upper Mojave in the same period of record.

The two watersheds (of the Ash Hill Channel and the Upper Mojave River) are only 110 km apart. They are both within the Mojave Desert. However, one--the Upper Mojave--heads at elevations of more than 2500 m, while the other is nowhere higher than 1010 m elevation. The present-day maximum mean annual precipitation in the Upper Mojave River Watershed is almost an order of magnitude larger than that at Ash Hill. At the same time, the Upper Mojave catchment is almost two orders of magnitude larger in area than that of Ash Hill.

The demonstrated discrepancy between the estimatd discharge at channel Cross Section C-C' and that expectable from the local watershed is interpreted as strong evidence for a source of discharge other than the nearby watershed.

Plate 4 shows the hook-shaped curve of the Ash Hill Channel where it is crossed by the old railroad. At this site piles of weathered basalt fill obviously left from construction of the bridge abutments lie undisturbed across the full width of the channel. The sandstone blocks of the abutment are unscathed, even where they would protrude into the current. More than a century in this channel has not produced enough water to alter small piles of fill in the channel. On the other hand, another sandstone bridge abutment 2.5 km to the east on the same century old alignment has been abraded buried, and exhumed perhaps several times (including at least 2 m of fill and excavation--the latter between 1975 and 1982).



Comparison of peak discharge at Ash Hill with maximum discharge in relation to drainage area for other parts of the United States (modified after Hoyt and Langbein, 1955).

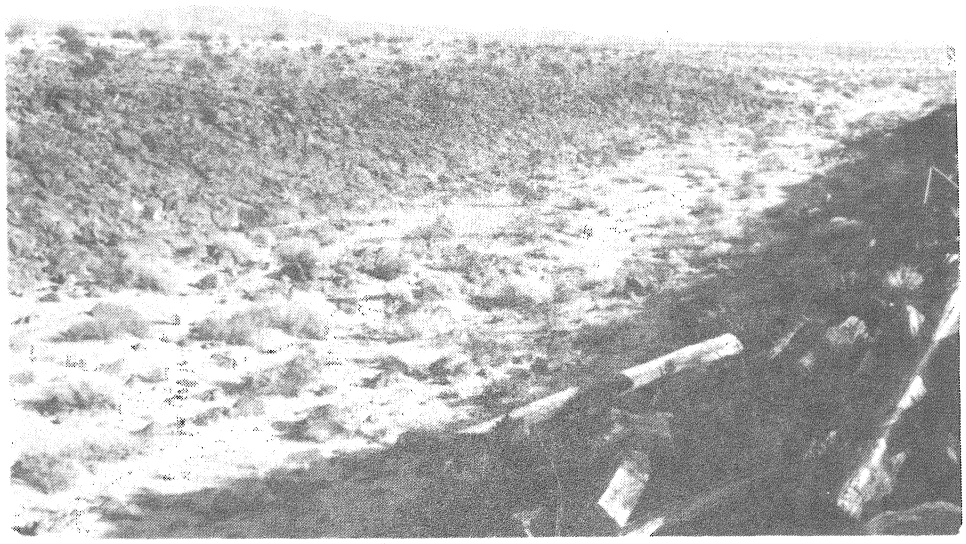


PLATE 5. VIEW DOWN STRAIGHT REACH SHOWS SEDIMENT STARVATION, THIN AEOLIAN VENEER, AND RIFFLE.

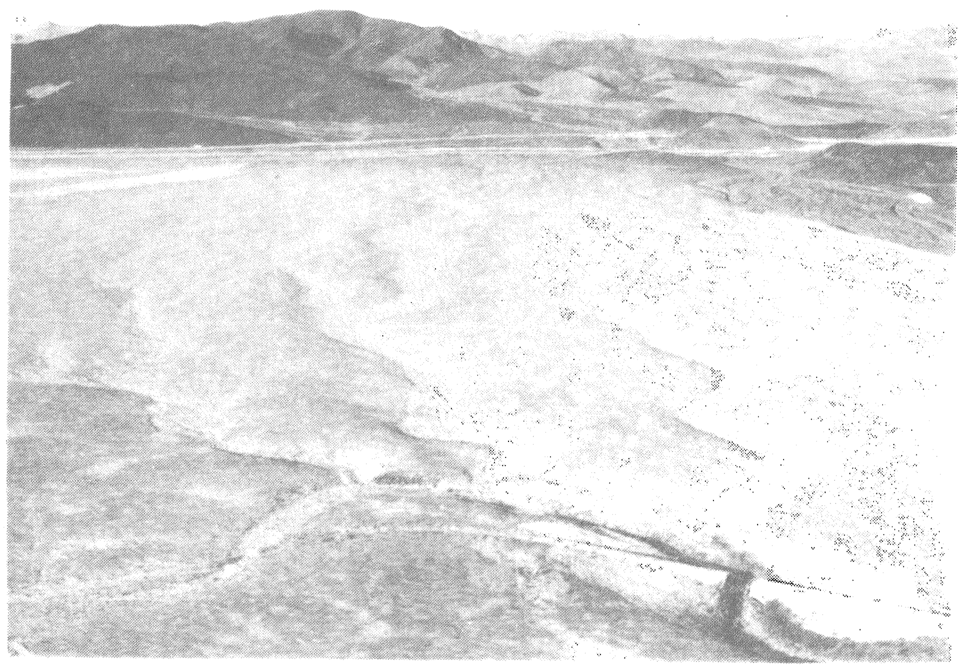


PLATE 4. BRISTOL MOUNTAINS & ASH HILL FAN ACROSS HOOK-SHAPED BEDROCK CHANNEL BEND. NOTE GRADING OF FAN CHANNELS AND TRIBUTARY RELATIONSHIP OF SAME TO BEDROCK CHANNEL.

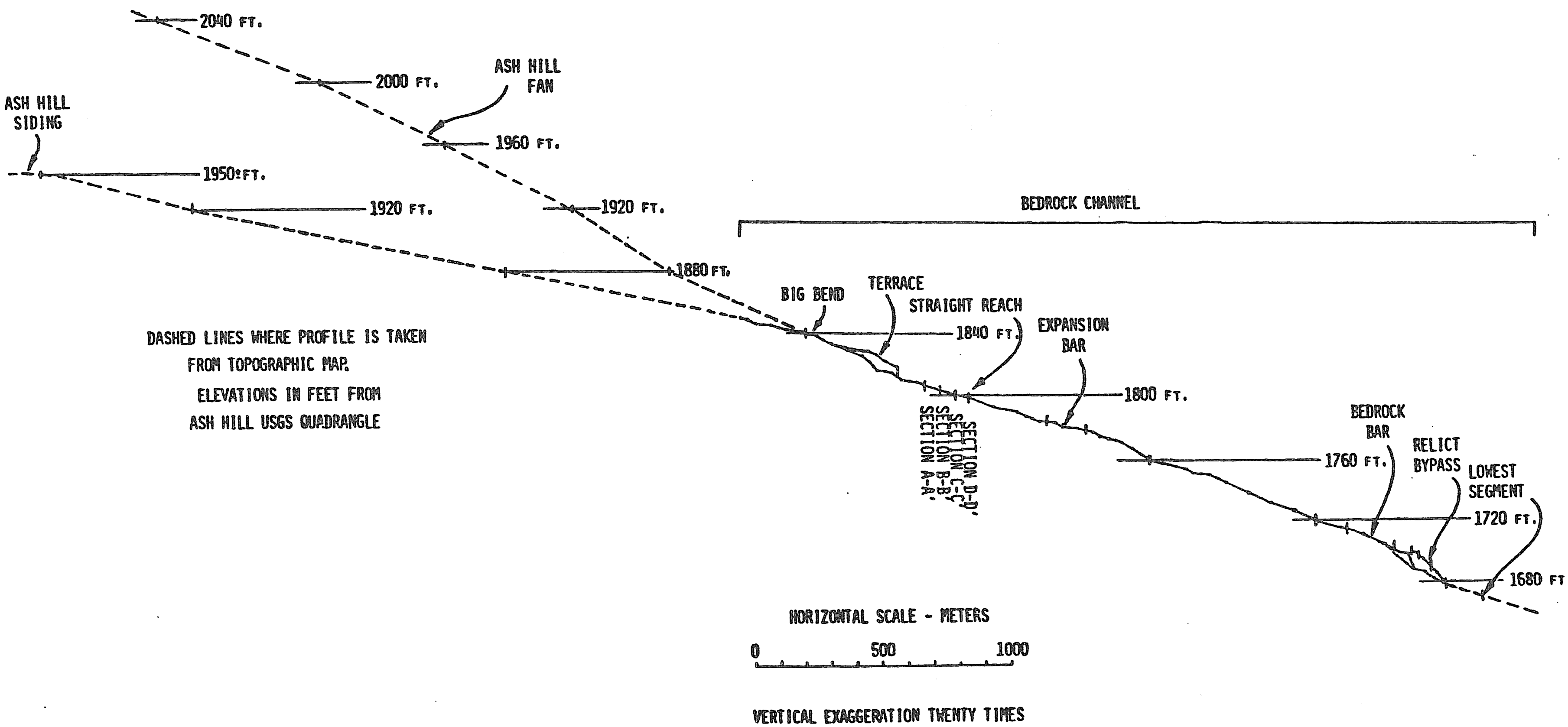


FIGURE 4. Longitudinal profiles from Ash Hill and Ash Hill Fan to lower end of Ash Hill Channel.

Channel Slope and Width

The longitudinal profile (Figure 4) of the bedrock channel steepens gradually downstream and so is gently convex upward. In a tectonically stable setting, this form is consistent with the presence of a resistant sill where incision has proceeded toward a graded profile but is not complete. Examination of the channel profile reveals two knickpoints--one in the hook-shaped bend and the other about 2 km downstream from the first. The presence of these knickpoints is taken as a sign that incision was in progress when the channel last carried large discharges.

Two longitudinal profiles are shown in Figure 4 upstream from the bedrock channel. Both are taken from the Ash Hill 7½-minute topographic map (USGS, 1955). The lower of these is the path water would follow from the watershed divide at Ash Hill. The upper is the path water follows as it courses down the most recently active channel on the beheaded Ash Hill Fan.

The gradient of the channel straight reach averages .015 (meters drop in elevation/meters horizontal distance). In the horizontally-measured kilometer upstream from the bedrock channel, the slope of the watercourse from the Ash Hill Divide is .010, while that of the fan channel is .028. This means that the bed slope of the channel straight reach is only half as steep as the fan channel that feeds it. At the same time the straight reach is considerably steeper than the watercourse from Ash Hill Divide.

The bedrock channel is also wider than its tributary fan channels. Channel width in the straight reach average 35 m. The width of the fan channel measured 80 m upstream from its confluence with the bedrock channel is 21 m. Sediment-bearing flows that would pass from the Ash Hill Fan channels into the wider, less-steeply sloping bedrock channel could be expected to deposit much of their load in the latter. The bedrock channel, however, is almost completely devoid of sediment.

Bedrock in the channel is masked by only a thin (10-20 cm) veneer of fine, clean sand--probably of aeolian origin. At Section C-C' clast sizes were measured at 115 sampling points on the channel floor. Only 16 particles having an intermediate diameter of 4 mm or more were encountered. The remaining 99 measurements were either bare bedrock or particles smaller than 4 mm. Plate 5 shows the generally debris-free character of the channel floor in the straight reach. Such a condition is consistent with an overflow of relatively sediment-free lake water from the Ash Hill Divide that passes over an increasingly steep basalt caprock surface.

Inspection of the longitudinal fan profile of Figure 4 also suggests that the channel incision would not be expected had most of the discharge come from the Ash Hill Fan. The longitudinal fan channel profile between 1960 and 2000 ft elevation may be extrapolated downstream in a linear fashion following generally-accepted alluvial fan form for tectonically stable settings (e.g., Bull, 1964). The extrapolated surface of the fan (and its channels) projects well above the channel banks. Any channel incision would, therefore, be anomalous. The upward convexity of the fan channel revealed by the longitudinal profile is consistent with subsequent and continuous removal of debris at the toe of the fan.

A third set of features that supports an exotic channel discharge source is as follows: The alluvial fan channels enter the bedrock channel perpendicularly, and one of them is the best developed of Ash Hill Fan channels and was probably the last of the fan channels that carried runoff from the Bristol Mountains prior to the fan's beheading. Together, these features suggest subordination of the fan discharge to the flow in the channel.

It might be argued that before entering the channel, a large hypothetical discharge first flowed down the Ash Hill Fan somewhere farther to the west. This hypothesis is rejected because there is no sign of the large fan channel that would necessarily have been involved. The longitudinal fan profile that radiates west toward Ash Hill Divide is upwardly convex. Any amount of water passing over the Ash Hill Fan surface great enough to incise a deep, wide channel in bedrock farther downstream should have left a large incised channel in the steeper alluvium of the Ash Hill Fan. It was pointed out earlier that one fan channel emptying into the hook-shaped bend appears to be the last fan channel to carry runoff from the apex of the fan. Obliteration of another fan channel draining to the Ash Hill Divide and subsequent incision of the channel emptying into the hook-shaped curve would have involved the excavation and transport of considerable debris, much of which should have lodged in the bedrock channel. Such deposits are not present. Also, the confluence at the hook-shaped curve would necessarily have been modified so as to make the fan channel into the dominant branch. The reverse is the case.

The lack of channel incision between Ash Hill and the Channel Entrance is consistent with other large watercourses where a reach of steep or precipitous gradient is approached by way of a volcanic tableland. The entrance to Fossil Falls on the Owens River near Little Lake, California, is an example. Although the magnitude and duration of channel activity in the latter instance were probably both much greater, the effect is the same as that observed at Ash Hill--signs of channelization tend to disappear as one moves upstream over the volcanic tableland from the incised reach.

It is for these several reasons that the large discharge estimated at Cross Section C-C' of the straight reach is interpreted as having come over the watershed divide at Ash Hill rather than from the nearby 18.4 km^2 catchment.

Although the evidence so far has been only circumstantial, the most likely source for such a discharge would be a large overflowing lake filling the Broadwell Basin and the large basin system connected to it, of which Death Valley is a part.

Shore Process Evidence Near Ash Hill

Direct evidence for such an overflowing lake has been found 3 km northwest of Ash Hill Siding on the smaller of two volcanic vents in the Ash Hill Basalt. The evidence is in the form of weathered rounded pebbles that are scattered along a horizontal contour at the overflow elevation. Rounding of pebbles at this site by any process other than wave action appears impossible. The site is uniquely suited to the development, preservation, and identification of shore process evidence of the hypothesized lake.

The larger of the two vents rises to 615 m elevation. This is 22 m higher than the overflow elevation. The smaller vent is a parasite upon the larger one. Maximum elevation of the smaller vent is 600 m. Both cones are centered on one of the several NNW-trending basalt ridges occupying the western half of the Ash Hill Trough.

If the Death Valley Watershed were filled today to the Ash Hill overflow level, these two vents would be small islands in the middle of the Ash Hill Trough very near the spillway. Wave action seems the only logical explanation for sound, rounded pebbles of basalt at or below overflow elevation on the upper parts of these cones. When it is not deeply buried, the Basalt of Ash Hill shows a strong tendency to weather by fracturing which produces angular fragments. Where deeply buried it is found as large joint blocks with well-developed calcium carbonate in the joints.

A test was conducted to see if a shoreline could be identified at the Ash Hill Vents on the basis of pebble rounding. The shapes of the best-rounded pebbles at the overflow elevation on the volcanic vents (594 m) were compared with those 20 m higher at the top of the larger vent. The higher elevation would, presumably, have been above water at all times unless a very large quantity of material had been removed from the spillway area at Ash Hill Siding. The procedure used was as follows:

(1) A small area chosen for sampling was traversed back and forth until a pebble showing unusual rounding was noted.

(2) A square meter grid was placed over the pebble so that any other similarly-rounded nearby pebbles were included. The objective was to take samples of the best-rounded pebbles the area has to offer.

(3) The five best-rounded pebbles were selected by eye (very coarse gravel and small cobble sizes were chosen).

(4) Each pebble was categorized according to an accepted criterion. This study used the Criteria for Evaluation of Sand-sized Grains established by Powers (in Compton, 1962, p. 215).

(5) The data were plotted by elevation categories. All samples taken at the suspected shore elevation are shown together in one plot, and all samples taken at the higher elevation are shown together on another plot. Both plots appear together in Figure 5 .

This rounding strongly supports the idea that the unusually great Ash Hill Channel discharges were caused by overflow from a large Death Valley lake.

ELEVATIONAL COINCIDENCE OF LACUSTRINE FEATURES ELSEWHERE IN BASIN

In addition to the rounded pebbles observed on the Ash Hill volcanic vents, a lake like the one hypothesized should have left shore process or depositional evidence of its existence in other parts of its basin. An extensive search for

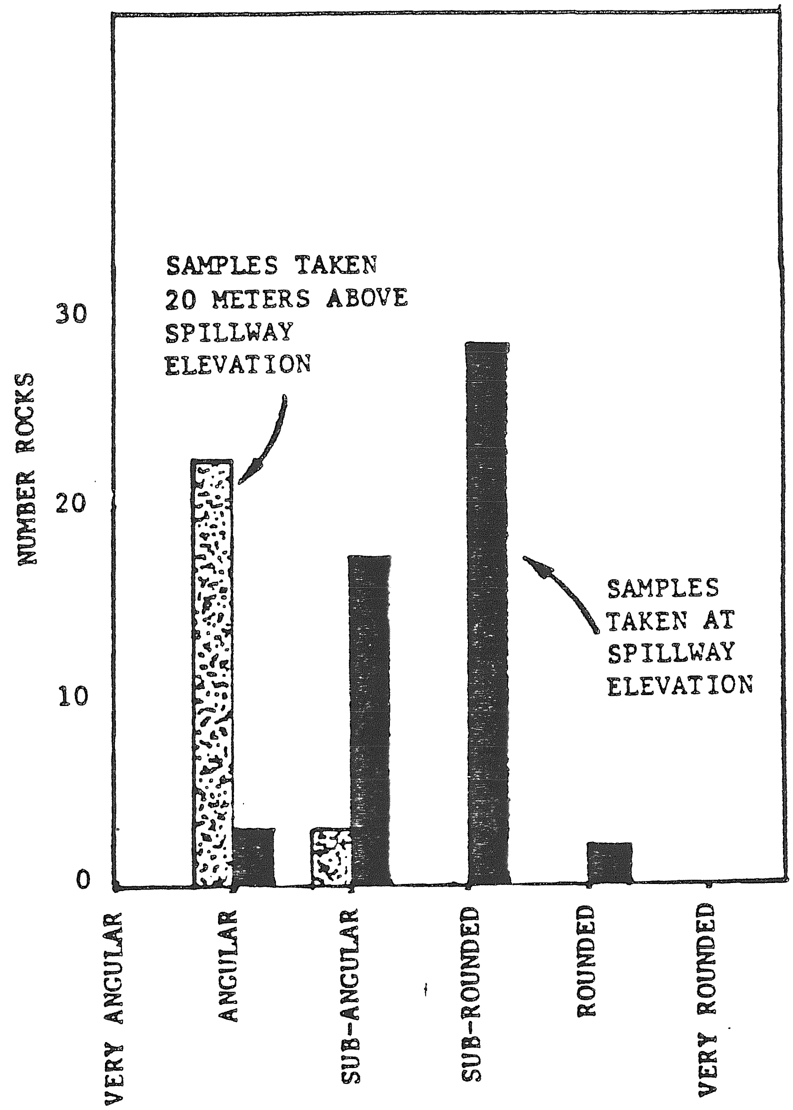


Figure 5. Degree of rounding exhibited by very coarse gravel and small cobbles in Death Valley Basin near Ash Hill Siding at shoreline or spillway elevation of 594 meters (dark bars) and 21 meters above shoreline elevation at 615 meters (stippled bars). Seventy-five rocks (15 samples) measured. Degree of rounding adapted from criteria for evaluating sand-sized grains after Powers in Compton, 1962, p. 215.

such evidence within the hypothesized lake basin has encountered a dozen sites where quiet water sediments crop out. None of these by itself provides irrefutable evidence of the overflow. Further examination will be needed before their actual relationships, if any, to the hypothesized lake are known.

Together, however, they are suggestive of an overflow-level lake in Death Valley. The suggestion comes from their almost coincident elevations. Although these sites are at widely-scattered locations in the hypothesized lake basin, nine of the twelve are close to the 594 m overflow elevation. The exceptions--one at the north end of Death Valley, one at Denning Spring in the Garlock Fault zone, and the third near the Helendale Fault southwest of the town of Barstow--are all tectonically-active areas. All twelve sites are characterized by depositional features.

Summary descriptions of these sites are given in Table 1 .

Five of these twelve sites (Red Pass Valley, Manix Basin, Tecopa Basin, North Death Valley, and Denning Spring) are in basins tributary to the main Death Valley Basin. In each of these examples, the smaller tributary basin containing lake sediments is presently connected to the main basin by a gorge or constriction. At least two of these constrictions (Tecopa and Manix) appear to have been closed for much of the Pleistocene, damming the smaller tributary basin. However, it is not known whether the other three constrictions (Red Pass, North Death Valley, and Denning Spring) were similarly dammed in the past. The significance of the possible damming of these three constrictions is that if they were dammed at the time that lacustrine sediments within them were deposited, they would add nothing to the overflow hypothesis. If, on the other hand, these constrictions were open to the Death Valley Basin and the topography was otherwise unchanged from the present, sediments found in their basins would indicate the presence of a lake like the one hypothesized. If each of these three basins were connected in estuary fashion to an overflowing Death Valley Lake, they would act much like settling basins permitting fine materials to settle close to the water's edge before passing into the main lake basin. Whether or not this mechanism was at work at the time the Ash Hill Channel was being incised is not known. Further investigation may provide an answer.

Another factor in the positioning of sediments from a large lake like the one hypothesized is isostatic deformation. Mifflin and Wheat (1971) and Crittenden (1963) have investigated isostatic deformation of shore features for Pleistocene Great Basin lakes. However, it is difficult to predict the extent to which isostatic deformation and rebound may have affected the elevations of shore features of the hypothesized lake, because it is not yet known whether the lake existed for long enough to fully depress its basin.

The elevational coincidence of apparent lacustrine features close to the overflow elevation of the hypothesized lake is remarkable. In spite of the possible alternate explanations for the sediments involved, the close elevational correspondence of so many sites tends to support the idea that a large lake filled Death Valley and overflowed toward the Colorado at sometime in the Pleistocene.

TABLE 1

DESCRIPTIVE SUMMARY OF POSSIBLE LACUSTRINE FEATURES
ELSEWHERE IN HYPOTHESIZED LAKE BASIN

Name Highest Elevation Meters (Feet) USGS Topographic Map (Calif. Geol. Map)	Location (Township and Range References are to San Bernardino Base and Meridian unless Specified	Descriptive Remarks
Red Pass Valley 585 m (1920 ft) Red Pass Lake (Trona)	Scattered between SE¼, Sec 12 T14N and NW¼, NW¼, Sec 7, T14N, R7E	Areally extensive lacustrine section with maximum vertical exposure of 10 m; no apparent deformation; well-developed dissected pavement at top lacustrine beds; weakly-developed K-horizon; constricted 500 m wide outlet to Silurian Basin; conceivably dammed at basin outlet; tributary watershed 650 km^2; no age control available.
Powerline Road 585 m (1920 ft) Red Pass Lake (Trona)	SW¼, SE¼, Sec 19, T14N, R7E	Small outcrop of moderately indurated mudstone; site open to Silurian Lake Basin; vertical section 3.5 m; unconformably overlain by 8 m overburden of cemented fanglomerate; mudstone section moderately deformed; no age control available.
Tecopa Basin ~610 m (~2000 ft) Shoshone, Tecopa, Eagle Mountain (Trona)	35°50'; N. Lat. to 36°15 N. Lat. and 116°5 W. Long. to 116°20'; W. Long.	Areally extensive sedimentary section representing an essentially continuous lacustrine history from Late Pliocene to Late Pleistocene described by Sheppard and Gude (1968); Amargosa River presently drains Tecopa Lake Basin to Death Valley via deep, narrow gorge at China Ranch; age control from volcanic ash correlations by Sarna-Wojcicki and others, 1980; most sediments undoubtedly deposited in lake basin when separate from Death Valley; basin would have become integrated with Death Valley to form hypothesized lake.

Name Highest Elevation Meters (Feet) USGS Topographic Map (Calif. Geol. Map)	Location (Township and Range References are to San Bernardino Base and Meridian unless Specified	Descriptive Remarks
Manix Basin 549 m (1800 ft) Cave Mtn., Cady Mtns., Alvord Mtn., Newberry Daggett, Barstow, (Trona, San Bernardino)	$34^{o}45'$ N. Lat. to $35^{o}5$ N. Lat. $116^{o}15'$ W. Long. to 117^{o} W. Long.	Lacustrine sediments representing Late Pliocene to Late Pleistocene lacustrine episodes are described variously by Buwalda (1914), Blackwelder and Ellsworth (1936), and Jefferson and others (1982); basin was dammed until opening of Afton (Cave) Canyon sometime after 20,000 B.P.; volcanic ashes associated with lacustrine sediments dated at $\sim$2 M.Y.B.P. (Sarna-Wojcicki and others, 1980) and .185 M.Y.B.P. (Izett in Jefferson and others, 1982); as with Tecopa, sediments undoubtedly deposited in lake basin that was separate from Death Valley; would have been integrated with Death Valley to form hypothesized large lake.
Soda Lake Basin 622 m (2040 ft) Old Dad Mountain (Kingman)	SW¼, SW¼, Sec 31, T12N, R11E	.6m thick claystone lens of unknown areal extent preserved beneath surface of alluvial fan at margin of Devil's Playground; site is open to Soda Lake Basin; stratum dips less than 4^{o} basinward and is truncated by steeper fan surface; no age control available; clay could conceivably have been deposited in a small pond formed by sand dune.
Cronese Basin Cronese Basin 549 m (1800 ft) Cave Mountain (Trona)	Sec 9, T12N, R5N	Outcrop of indurated, moderately deformed mudstone described on geologic map of California as upper Miocene non marine. Site is open to Cronese and Soda Lake Basins. no age control available; may have been deposited much earlier than time of hypothesized lake.

Name Highest Elevation Meters (Feet) USGS Topographic Map (Calif. Ceol. Map)	Location Township and Range References are to San Bernardino Base and Meridian unless Specified	Descriptive Remarks
Owl Hole Spring 488 m (1600 ft) Leach Lake (Trona)	N½, Sec 23, T18N, R3E	Areally extensive lacustrine deposit of non-indurated pink claystone; moderately to steeply tilted (15^{o} to 60^{o}); no age control available; site is immediately adjacent to Garlock Fault zone and is open to Death Valley; sediments may be part of a large Tertiary lake described by Bell (1971).
Helendale Quarry 799 m (2620 ft) Hawes (San Bernardino)	SW¼, SW¼, Sec 9, T8N, R4W	Areally extensive lacustrine section of buff to beige mudstone; vertical exposure 10 m; beds dip gently (10^{o} to 15^{o}) to north; open toward Harper Lake; sole suggestion of age is from opalization reported by Jefferson (personal communication, 1983); probably uplifted by displacement of Helendale Fault following deposition.
North Death Valley 830 m (2700 ft) Ubehebe Crater Last Chance Range (Mariposa)	T10S, R41E, Mt. Diablo base and meridian	Apparent lacustrine sediments mentioned by Blackwelder (1954) as possibly correlated with Sherwin glaciation based on degree of dissection; site is in tectonically active area and may have been elevated as a sliver attached to the Last Chance Range; the constriction separating the site from Death Valley is 1.3 km wide at 760 m elevation and 5.8 km wide at 820 m elevation; sediments conceivably could have been impounded behind a large dam at this wide constriction; the watershed feeding the site is approximately 1000 km^2 in area; no age control available.

Name Highest Elevation Meters (Feet) USGS Topographic Map (Calif. Geol. Map)	Location Township and Range References are to San Bernardino Base and Meridian unless Specified	Descriptive Remarks
Silurian Lake Basin Near Silurian Hills 561 m (1840 ft) Silurian Hills (Trona)	SW¼, NW¼, Sec 6, T16N, R9E	Small exposure of possibly lacustrine indurated silt overlain by pebbly alluvium; minimal clay content; open to Silurian Lake Basin; no age control available; sediment may be loess.
Silurian Lake Basin at Kingston Wash 570 m (1900 ft) Kingston Peak (Kingman)	SW¼, NW¼, Sec 26, T17N, R9E	Areally small lacustrine section of friable pink clay preserved beneath surface of alluvial fan; 4 m beneath thin pebbly alluvial overburden; no age control available; sediments may be part of Tertiary lake described by Bell (1971).
Denning Spring 671 m (2200 ft) Avawatz Pass (Trona)	$35^{\circ}35' 18''$ W. Lat. $116^{\circ}28'13''$ W. Long.	Areally small exposure of beige to pink horizontally bedded lacustrine mudstone; section is 5 m thick and overlies pebbly alluvium conformably; alluvium rests unconformably on a moderately ($30^{\circ}\pm$) dipping section that includes terrestrial and lacustrine sediments; no age control available; may have been deposited in ponds formed at several times on Garlock Fault in Late Pleistocene or Holocene; indurated, older, possibly lacustrine beds at highest elevation are buried by alluvial overburden .5 km west of location shown; site is connected to Death Valley by 100 m wide constriction; may have been dammed formerly.

TABLE 2

COMPARISONS OF MODERN AND PLEISTOCENE LAKE CLIMATES

LAKE NAME, VALLEY NAME	WHOLE PLUVIAL WATERSHED AREA $A_W = A_T + A_L$ km^2 (mi^2)	MODERN CONDITIONS: Lake Area A_L km^2 (mi^2)	Hydrologic Index $z = \frac{A_L}{A_T}$	P_L mm (in)	P mm (in)	E_L mm (in)	E mm (in)	Relief Factor $\frac{a}{E}$	Seasonal Factor	$\frac{P}{E}$	PLUVIAL CONDITIONS: Lake Area A_{L_2} km^2 (mi^2)	Pluv. Hydr. Index $z = \frac{A_L}{A_T}$	$(\frac{P}{E})'$	CHANGE $\Delta(\frac{P}{E})$
Death Valley, Calif. Hypothesized Lake System	65278 (25183)	520 (200)	.011[a]	118 (5)	202 (8)	1840 (72)	1592 (63)	.21	1.5	.13[a]	11949 (4610)	.22[b]	.57	.44
Death Valley, Calif. Lake Manly System	65806 (25387)	520 (200)	.010[a]	135 (5)	197 (8)	1920 (76)	1637 (64)	.21	1.5	.12[a]	6507 (2510)	.11[b]	.47	.35
Mono Basin, Calif. Lake Russell (Component of both Death Valley Systems)	2058 (794)[c]	223 (86)[m]	.12	203 (8)[m]	---	1070 (42)[m]	890 (35)[m]	.30[d]	1.5[d]	.49[b]	894 (345)[m]	.77[b]	.79	.30
Deep Spring, Calif. Lake Deep Spring	539 (208)[c]	7 (2.6)[de]	.012[d]	170 (6.7)[d]	280 (11.0)[d]	1100[d] (43)[d]	1030[d] (40)[d]	.20[d]	1.5[d]	.15[b]	44 (17)[c]	.09[h]	.43[h]	.28[h]
Spring Valley, Nev. Lake Spring Valley	4254 (1641)[c]	65 (25)[de]	.015[ad]	203 (8.0)[d]	305 (12.0)[d]	1168[d] (46)[d]	1118[d] (44)[d]	.14[d]	1.12[d]	.27[ad]	868 (335)[dg]	.26[d]	.65[d]	.38[d]
Ruby Valley, Nev. Lake Franklin	4917 (1897)[cl]	104 (40)[df]	.022[h]	230 (9.0)[d]	340 (13.5)[d]	1130[d] (44)[d]	1030[d] (40)[d]	.15[d]	1.15[d]	.32[hl]	1230 (475)[d]	.33[h]	.67[h]	.35[h]
Big Smoky Valley, Nev. Lake Toiyabe	3357 (1295)[c]	32 (12.5)[de]	.01[d]	127 (5.0)[d]	268 (10.5)[d]	1372[d] (54)[d]	1320[d] (52)[d]	.15[d]	1.2[d]	.20[ad]	583 (225)[dg]	.21[h]	.61[d]	.41[h]
Lahontan Basin, Nev. Lake Lahontan	112295 (42322)[i]	2074 (800)[df]	.02[d]	---	---	---	---	.20[d]	1.25[d]	.30[bd]	22460 (8665)[ig]	.25[h]	.62[h]	.32[h]
Great Salt Lake Basin, Utah Lake Bonneville	138224 (53325)[ij]	7776 (3000)[dj]	.06[h]	203 (8)[d]	---	1016[d] (40)[d]	889[d] (35)[d]	.25[d]	1.15[d]	.43[hj]	51686 (19940)[c]	.60[h]	.76[h]	.33[h]
Estancia Basin, N.M. Lake Estancia	5184 (2000)[cd]	39 (15)[dk]	.0076[d]	305 (12)[d]	381 (15)[d]	1473[d] (58)[d]	1397[d] (55)[d]	.11[d]	.85[d]	.26[d]	1166 (450)[dg]	.29[d]	.67[d]	.41[d]

a - Calculated from climatologic data
b - Calculated from value of z
c - Snyder, Hardman and Zdeneck, 1964
d - Snyder and Langbein, 1962
e - No perennial lake
f - Sum of present lake areas estimated
g - Did not overflow
h - Recalculation based on Snyder, Hardman, and Zdeneck, 1964.
i - After Russell (in Snyder and Langbein, 1962)
j - Includes Utah Lake; whole watershed (A_W) for Great Salt Lake and Utah Lake only is 65,000 km^2 (Snyder and Langbein, 1962); using this watershed to calculate modern $\frac{P}{E}$ gives .14
k - Estimated from map by Meinzer (in Snyder and Langbein, 1962)
l - Whole watershed (A_W) for Ruby Lake only is 3,370 km^2 (Snyder and Langbein, 1962); using this watershed to calculate modern $\frac{P}{E}$ gives .24
m - Stine, personal communication, 1983

TABLE 3

COMPARISONS OF LAKE SYSTEM MAGNITUDES

Age and Watershed	Watershed Area Including Lakes ($10^3 km^2$)	Lake Surface at Maximum Stand ($10^3 km^2$)	Ratio: Area Lakes to Area Watershed Including Lakes	Instantaneous Peak Discharge Toward Sea ($10^3 m^3$/sec)
Pleistocene Death Valley (Hypothesized)	65	12	.18	2
Pleistocene Death Valley (Manly)	66	7	.11	none
Late Pleistocene Lahontan	112	22	.20	none
Late Pleistocene Bonneville	138	52	.38	400*
Modern Sacramento River at City of Sacramento	69	8**	.12	17***

*Probable catastrophic removal of 30 m alluvial dam interpreted by Malde (1968)
**100-year flood inundation area (Kahrl, 1979)
***Computed 100-year recurrence-period flood (Calif. Dept. of Water Resources, 1983)

Paucity of Visible Shore Features

Extensive searching of most of the hypothesized lake's parimeter has failed to reveal any unmistakable shore traces. This contrasts sharply with the abundant and readily identifiable relect shore features of Late Pleistocene Lake Bonneville described by Gilbert and others (summarized in Crittenden, 1963). The absence of visible shorelines of horizontal lineation features for the hypothesized lake could be the result of some combination of the following: (1) a very short duration at the overflow elevation; (2) a relatively small fetch limiting wave energy (the power of the bench-cutting waves); or (3) a long interval of time since the shore features were formed. At present, little is known about the duration of the lake's overflow. The fetch of the hypothesized lake (were it in existence today) would nowhere be of the (same) magnitude attained over much of Lake Bonneville. The age of this lake (discussed elswhere in this paper) appears to be considerable.

DESCRIPTION OF HYPOTHESIZED LAKE SYSTEM

If it were to exist today, the lake that would spill toward the Colorado River through Ash Hill Channel would be more than 300 km in length, more than 25 km in average width, and--at its deepest place--more than 650 m in depth. Its surface would total more than 8000 km^2. If it is also assumed that the smaller lakes tributary to the large hypothesized lake would be of the extent shown by Snyder and others (1964), the total surface of this lake system would be about 12000 km^2. The watershed of the hypothesized lake system would be slightly larger than 65000 km^2 in area.

Computations earlier in this paper have suggested a peak discharge from the hypothesized lake of about 2000 m^3/sec. This means that in addition to supporting the lake system described above, there was a surplus runoff from the hypothesized lake watershed averaging .03 m^3/sec/km^2) for the duration of the estimated peak flow. This figure assumes that neither extraordinary snowmelt nor other catastrophic event--such as a lava flow, landslide, of natural dam failure--suddenly displaced a large volume of water from the lake.

Tables 2 and 3 show areas for the hypothesized lake system, its watershed, and similar values for other pertinent lake systems. The values for the hypothesized lake system and Manly Watershed were obtained by planimetric means from topographic maps of California at a scale of 1:500,000. Other values in these tables were taken from the sources cited. Values in Table 3 were taken from other parts of this paper, or from the sources cited, and rounded off to the nearest thousand km^2 or km^3.

A comparison of the relative size of watershed areas, lake surfaces, and peak discharge of overflows--if any--can be helpful in placing the hypothesized lake system into context with other systems of Pleistocene or modern lakes. Table 2 contains approximate values of these three parameters for several lake systems. It shows that while the catchment areas of the hypothesized and Manly Lake systems were about the same size, the former supported lake surfaces about twice as great as the latter and produced a surplus 2000 m^3/sec peak discharge as well.

In another comparison, the ratio of lake surface area to watershed area (Table 2) for the hypothesized lake is essentially the same as that for the Late Pleistocene Lake Lahontan system at its maximum (See Plate 1). However, the Lake Lahontan system did not experience overflow. This means that the hypothesized lake watershed was capable of sustaining a somewhat larger lake surface per watershed unit than was the Late Pleistocene Lake Lahontan system.

The Bonneville Lake system, on the other hand, was of much greater magnitude than the hypothesized lake system (See Plate 1). A watershed twice as large as that of the hypothesized lake system sustained lake surfaces more than four times as great as those of the hypothesized system. In addition, the peak overflow discharge of 400,000 m^3/sec from Lake Bonneville (reported by Malde, 1968) was roughly two orders of magnitude greater than that of the hypothesized lake system. Malde interprets this large figure for overflow discharge as being the probable result of a rapid lowering of the lake by about 30 m, perhaps because an alluvial overburden was stripped from the bedrock spillway area. No evidence has yet been found in the hypothesized lake watershed to suggest a similar explanation for the large peak discharge at Ash Hill.

In a fourth comparison, the Modern Sacramento River at Sacramento, California, is fed by a catchment area of about 69000 km^2 which is essentially the same size as the watershed of the hypothesized lake. The water surface, including lakes expected to be inundated by the 100-year recurrence interval flood on the Sacramento, is approximately 10000 km^2--or slightly less than the water surface of the hypothesized lake system in Death Valley. The peak outflow from this system during the same 100-year event is expected by the California Department of Water Resources to be about 17000 m^3/sec, or eight times that estimated for the hypothesized lake system at Ash Hill.

On a per-watershed basis, therefore, the hypothesized watershed was capable of sustaining a system of evaporative surfaces that was: (1) significantly larger than the Pleistocene Lake Manly system; (2) slightly larger than the Late Pleistocene Lake Lahontan system; (3) considerably smaller than the Late Pleistocene Lake Bonneville system; and (4) similar in magnitude to the present Sacramento Valley 100-year flood inundation area, but with a much smaller peak outflow than that calculated for the latter watershed.

The larger size of the hypothesized lake in the Death Valley Watershed when compared to the Pleistocene Lake Manly system is probably attributable, in part, to differences in the Transverse Ranges and Sierra Nevada rainshadow (discussed elswhere in this paper). The same (rainshadow) factor probably figured importantly (by offsetting latitudinal and elevational differences) in the close similarity of the late Pleistocene Lake Lahontan system and the lake system hypothesized for Death Valley.

There is general agreement that the Late Pleistocene Lake Bonneville Watershed was capable of sustaining the largest evaporative surface in the Great Basin. Its relatively high latitude and altitude, the high altitude of the mountains at its eastern margin, and its position removed from the rainshadow of the Sierra Nevada all contributed to make it the moistest sector of the Great Basin.

The greater runoff observed for the modern Sacramento Valley, when compared to the Death Valley Watershed at the time of the hypothesized lake is easily attributable to both latitude and the position of the mountains bordering the two watersheds.

Watershed Configuration

Investigations by several authors suggest that Pleistocene tectonism in the basin of the hypothesized overflowing lake has been at times severe in the central and western parts of the Mojave Tectonic Block, moderate to minor north and northeast of the Garlock Fault, and minor to non-existent east of the Ludlow Fault and south of the Kingston Range (Bull, personal communication, 1983; Dohrenwend and others, 1984; Weldon, personal communication, 1983; Sarna-Wojcicki and others, 1983; Sheppard and Gude, 1968; Jefferson and others, 1982; Jefferson, personal communication, 1983; Dibblee, 1967; Hoover and others, 1982; Smith and others, 1983; Wernicke and others, 1982; Christiansen and Blank, 1972). The suggestion taken from these interpretations is that it is reasonable to assume something close to modern topography when evaluating evaporative surfaces of the overflowing system of Death Valley lakes.

Tectonism in the Transverse Ranges

The mountains along the southwest border of the Death Valley Watershed appear to have had a somewhat different history from most of the hypothesized lake basin. Crowell (1982) reports correspondence of the basement terrane to the west of the San Andreas Fault in the Transverse ranges with that to the east of the San Andreas in the vicinity of the Orocopia Mountains in Southern California. The total strike-slip offset suggested by this correspondence amounts to about 250 km in 4 million years. The average displacement that is given by this offset is 6.25 cm/yr. If it is assumed that this displacement has occurred at a uniform rate, the total offset of the terrane on either side of the San Andreas Fault along the border of the Mojave Desert during the last million years can be interpolated to be about 60 km. Ehlig (1982) obtains a similar figure (240 km) for the same 4 M.Y. period based on correspondence of basement terrane in the Mint Canyon area with that found near Salton Wash on the opposite side of the San Andreas Fault. The same interpolation of about 60 km per million years can be made if the same assumption of uniform motion is made. The terrane that makes up the San Gabriel Mountains may, therefore, have been located due south of the San Bernardino Mountains in the mid-Pleistocene. Such lateral displacement would logically have placed western parts of the Transverse Ranges and possibly part of the Southern Coast Range along the border of the Mojave Desert in the mid-Pleistocene. The maximum elevation of the Mojave Desert border cannot be determined with certainty. Crowell points out that the Transverse Ranges apparently lack deep roots that might support them isostatically. This is suggested by the interpretation of earthquake waves and gravity studies (Hadley and Kanamori, 1977; Grannell, 1971). He states that the Transverse Ranges may therefore be held by plate-tectonic compression across the region (Crowell, 1982, p. 598). Thus, a significant fraction of the height presently observed in the Transverse Ranges along the southwestern margin of the Mojave Desert may have been attained within the last million years. This suggests that North Pacific extra-tropical cyclonic storms at one time had a relatively open corridor into the Mojave Desert and Southern Great Basin. It also means that given the same regional climates in mid- and Late Pleistocene, the earlier would produce greater runoff in the Death Valley Watershed.

Recent support for this idea has come from Winograd and others (1985). They report a 40 per mil reduction in the deuterium content of Southern Great Basin groundwater recharge during the Pleistocene as measured in fluid inclusions within dated calcitic veins. Presumably, as the Transverse Ranges and Sierra Nevada rose, they intercepted larger fractions of the heavier deuterium-containing water molecules. Also, Gable and Hatton (1983) summarize rates of uplift that could have pushed up the Transverse Ranges in the last million years.

Tectonism in the Sierra Nevada

The Sierra Nevada acts as an important rainshadow barrier for the northern parts of the Death Valley Watershed. Recent work on glaciation of Sherwin age in the Central Sierra Nevada by Curry (personal communication, 1983) suggests as much as 600 m of uplift in the vicinity of Tuolumne Meadows since the Sherwin Glaciation, which predates the 700,000-year-old Bishop tuff. Other recent work by Curry suggests that the total annual precipitation supporting the Sherwin Glaciation was much greater to the east of the present Sierran crest than in later Pleistocene glaciations, resulting in extensive glaciation on the eastern slope, with weaker glaciation to the west. This is consistent with the recently interpreted sediment record at Searles Lake which indicates a 300,000-year, continuous deep lake episode from 1.28 to 1.0 M.A. , followed by several deep lake episodes until about .6 M.A. (Smith and others, 1983).

By extrapolating old topographic surfaces, Huber (1981) arrives at a somewhat smaller figure for Central Sierra Nevada uplift, concluding that there has been 1000 m of uplift over the last 3 M.A. Assuming a uniform rate of tectonism, 300 m of uplift may have occurred in the Sierra Nevada near Deadman Pass in the last million years.

Huber also (1981) points out that higher rates of uplift are thought to have occurred in the Southern Sierra along the Western Margin of the Mojave Desert.

Regardless of the precise amount of Sierran uplift in the last half of the Pleistocene, the evidence from several sources suggests significantly greater mid-Pleistocene precipitation in the watersheds east of the range than that occurring in the Late Pleistocene and Holocene. This appears to support the possibility of a large mid-Pleistocene lake in the Death Valley hydrologic basin which could have spilled toward the Colorado River.

WATER BUDGET

Water budget computations have been carried out using the model of Snyder and Langbein (1962). The results suggest that the ratio between precipitation and evaporation $(\frac{P}{E})$ necessary to sustain an overflowing system of lakes like that hypothesized in this report is about .57. Similar computations suggest that the Lake Manly surface as described by Blackwelder (1933) would be characterized by a $(\frac{P}{E})$ ratio of .47.

TABLE 4

Estimated Precipitation and Evaporation Necessary
to Sustain Maximum-stand Pluvial Lake Systems
in the Death Valley Watershed

Postulated Seasonally Weighted Depression of Mean Annual Temperature	Watershed Mean Annual Precipitation/Evaporation to Sustain Hypothesized Mid-Pleistocene Lake System P/E = .57	Watershed Mean Annual Precipitation/Evaporation to Sustain Lake Manly at 75 m elevation P/E = .47
°C	mm precip/mm evap	mm precip/mm evap
3	660/1160	560/1190
5	490/860	430/910
7	330/580	290/620
8	230/400	210/450

The present mean annual precipitation for the Greater Death Valley Watershed is estimated to be 200 mm and the present mean annual evaporation for the watershed is estimated to be 1600 mm.

If temperature is assumed to have been related to evaporation over the whole basin as it is presently in the Town of Mojave, California, several postulated seasonally-weighted temperature depressions should produce the combinations of annual precipitation and evaporation shown in Table 4. By way of illustration, this means the following: (1) If mean annual temperatures are found to have been depressed by a seasonally-weighted 8^{o}C around the time of Death Valley overflow, then evaporation would have been one-quarter of what it is at present, and precipitation would have been only one-tenth greater. (2) If, instead, mean annual temperature is found to have been depressed by a seasonally-weighted 5^{o}C, then mean annual evaporation would have been half of what it is at present, while precipitation would have been two-and-one-half times as great as the present. Actual values of temperature depression for the several Pleistocene glaciations in the Mojave Desert and elswhere are the subject of wide-ranging debate and will not be discussed here (for example, see Dohrenwend, 1983; Galloway, 1970; Brackenridge, 1978, Mifflin and Wheat, 1979).

TIME SINCE OVERFLOW

The Ash Hill Fan was active and presumably capable of being graded during at least part, if not the whole, of the straight reach incision. Evidence for this is found in the fact that two of the three fan channels tributary to the hook-shaped bend are graded to the incised floor of the bedrock channel. The third--and smallest--hangs approximately 1.5 m above the bed. The additional fact that none of these fan channels has significantly incised or filled the fluvial channel at the hook-shaped bend suggests that both stopped about the same time. This can best be seen in Plate 4 . Cessation of grading activity on the fan was presumably the result of natural diversion or capture of the Bristol Mountains Watershed at the fan apex in the manner proposed by Denny in his discussion of steady-state alluvial fans (Denny, 1967). This diversion continues to bypass the fan today. It can be seen where Interstate Highway 40 crosses the fan apex. The final removal of material at the fan toe by bedrock channel discharge appears to have come at about the same time as the last grading of the fan. If such were not the case, the fan (which has an upwardly convex longitudinal profile) should have advanced its toe at least to the brink of the channel.

K-horizon in the Ash Hill Fan

The Ash Hill Fan exposes a strongly-developed calcrete or K-horizon. This K-horizon is important to the present study because the extent to which it is developed suggests a long period of inactivity of the fan. Long inactivity of the fan is, in turn, important because the last incision of the Ash Hill bedrock channel appears to have been contemporaneous with the most recent grading activity of the fan.

The calcrete in the Ash Hill Fan is exposed in virtually every site where the fan surface has been disturbed. In distant view undisturbed parts of the fan surface give no hint of the calcareous accumulation beneath. However, calcium carbonate is visible through the full 7 m depth of the active channel bank that has beheaded the fan on its east side.

Large areas of the fan surface have been scraped for road fill, exposing a smooth surface of well-developed Stage V calcrete. Elswhere on the fan this laminar horizon may be found lying 10-20 cm below the cobbly surface. A profile description of this calcrete may be found in Table 5.

The upper surface of the calcrete conforms to the contours of the fan. It is neither deeply buried by overburden nor truncated by the fan surface or its channels. No erosional overburden remnants are present on the fan to suggest that the K-horizon might be an exhumed, older, buried calcrete. These facts suggest that it has formed in place since the fan was beheaded.

The ages of seven Southwest U.S. Stage V calcretes are interpreted by Bachman and Machette (1977) to be between .4 MA and 2 MA. One of the seven is located at Vidal Junction, California, less than 150 km east of Ash Hill.

If their chronology is applicable to Ash Hill, it would suggest an Early or Mid-Pleistocene age for carbonate soils on the Ash Hill Fan and, thus, for the overflow of Death Valley toward the Colorado River. This interpretation is consistent with those of Butler (1982) and G.I. Smith (personal communication, 1985) for the last filling of Death Valley's Lake Manly. As was mentioned earlier, Lake Manly may have been a lower stand of the overflowing lake hypothesized in this paper.

Weathering Rinds

Surficial alteration by weathering can be observed on virtually all waterworn rocks lying more than half a meter above the Ash Hill channel bed. The same is true of the rounded pebbles observed at overflow elevation on the Ash Hill Volcanic Vent. Alteration of occasional pale green olivine phenocrysts to a rust red color is observable to a depth of 2 to 5 mm. The inner boundaries of true weathering rinds at these sites are difficult to distinguish with a hand lens from rock varnish and from the "inner rinds" described by Colman and Pierce (1981, p. 3). Weathering of vesicular samples of Ash Hill Basalt also occurs outward from individual vesicles a centimeter or more beneath the exposed surface. The weathering rinds of these rocks, therefore, are doubly difficult to identify and measure. The zone of major alteration, however, for more massive samples is clearly present. It is consistently between .3 and .7 mm in thickness. This magnitude figure is uniformly applicable to all relatively massive samples from both channel and shore sites at Ash Hill.

For purposes of comparison, weathering rinds of rounded basalt clasts from the shorelines of Death Valley Shoreline Butte were examined. Presumably, the same waves that carved the shorelines produced fresh surfaces on the cobbles making up the beaches there. The basalt of Shoreline Butte, like that of Ash Hill, varies from massive to vesicular. However, it is generally finer grained with fewer phenocrysts. Its vesicles tend to be smaller than those in the Ash Hill Basalt. In both vesicular and massive samples from Shoreline Butte, weathering rinds are similarly developed being clearly present, but consistently less than 1 mm thick on rounded surfaces.

The similarity in rind development between Ash Hill and Shoreline Butte suggests a possible contemporaneity between Lake Manly and the overflow at Ash Hill.

TABLE

PROFILE DESCRIPTION OF ASH HILL
LAMINAR CALCRETE

Overburden Unit (Thickness 100 - 200 mm)--This unit consists of poorly-sorted, cobbly fanglomerate and aeolian sand. The latter is abundant in all but the apical 10 percent of the fan surface. The larger cobbles protruding from the medial and distal fan surface bear the marks of long-term ventifaction. The exposed surfaces of these rocks are consistently more angular than their buried undersides. The upper surfaces of clasts not recently overturned are universally free of calcium carbonate. The clast undersurfaces that extend to the depth of the caliche are typically plastered with a laminar caliche a centimeter or more thick.

Unit A (Thickness .3 - 5 mm)--Where the loose fan material has been removed, buff to pinkish-white, smooth laminar calcrete is exposed. Plate 6 shows this upper stratum. In Plate 7 it can be seen on one of two large blocks quarried from the trench for a natural gas pipeline that crosses the fan. The laminar horizon is composed of many fine laminations in several distinguishable units. The uppermost (the top of the calcrete) is the lightest in color and is commonly less than 3 mm thick, although it may be 5 mm thick or more in places. Brecciation is uncommon in this unit. Its laminations are not as distinct as those beneath.

Unit B (Thickness 5 - 20 mm)--Immediately beneath Unit A is Unit B composed of many distinct, fine, undulating laminations. In the 5 to 20 mm thickness of this unit there are usually several dozen or more individual laminae. These vary in color from beige to white. Truncation, lensing, and variation in thickness are common within this zone. Brecciation occurs also, but is not characteristic. In several parts of the sample, the laminations change abruptly at the base of Unit B, becoming off-white to a thickness of 4 mm.

Unit C (Thickness 3 - 8 mm)--The laminations of Unit B grade downward into Unit C, which is a dense, dark golden brown, characterized by moderate brecciation which is clearly visible because fractures are recemented with lighter-colored material like that found above. Brecciated fragments show little or no sign of rounding prior to recementation. Truncation, lensing, and variation in thickness occur commonly in this unit.

Unit D (Maximum 16 mm thickness)--Betweem dark Unit C and another like it (Unit E) is Unit D which includes abundant small detrital pebbles. These pebbles are invariably separated from each other by the carbonate matrix. The calcium carbonate associated with the pebbles in this unit is generally porous and structurally weak.

Unit E (Thickness 5 - 10 mm)--The dense dark brown caliche encountered next in Unit E is much like Unit C above. However, brecciation and truncation are abundant. There is much wider variation in color and texture within the unit. Variation is from beige and somewhat porous (like the lighter-colored laminations above) to dark brown and dense (like the darker layers above).

Duricrust Unit (Thickness 300 - 1000 mm)--At the base of the laminated portion of the calcrete, the K-horizon becomes abruptly more porous and less dense. It loses most of its laminar quality, taking on instead a massive to nodular appearance. The few discernible laminations are coarse and poorly defined. The K-fabric extends to the base of this unit.

Particle sizes range from sand to small boulders. This cemented fanglomerate unit is the characteristic duricrust of the majority of the Ash Hill Fan. It varies between .3 m and 1 m in thickness, averaging about .6 m.

Cemented Clast Horizon (Maximum thickness 7000 mm)--Carbonate coatings on fan gravels are complete throughout this horizon. It extends to the full depth of all cross sectional exposures of the fan. The deepest exposure is 7 m on the fan's east side in the channel that has beheaded the fan.

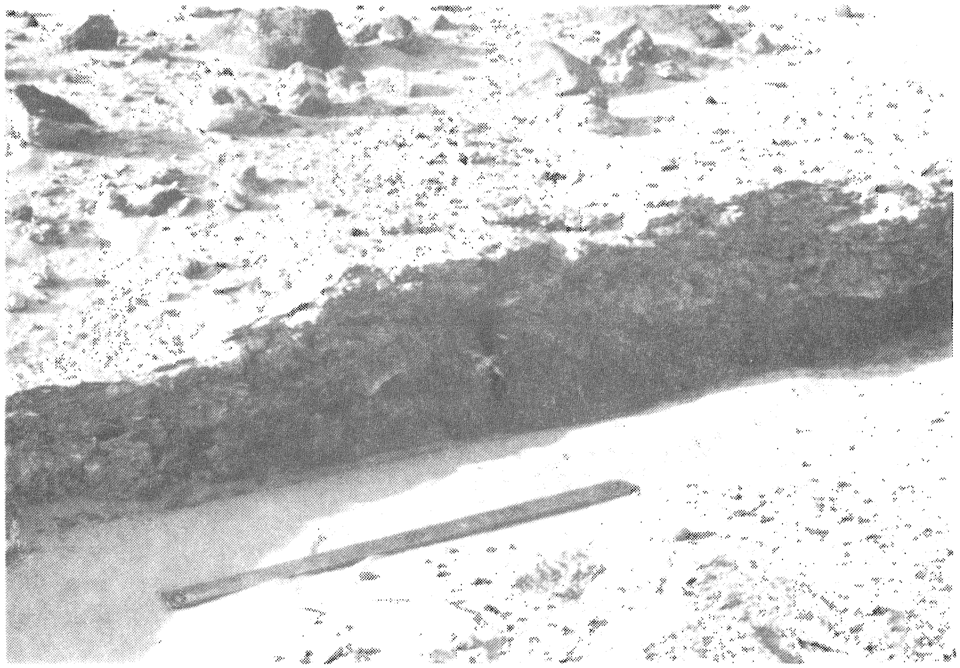


PLATE 6. ASH HILL FAN CALCRETE SHOWING PLUGGED HORIZON, TABULAR STRUCTURE, AND LAMINATIONS AT SURFACE.

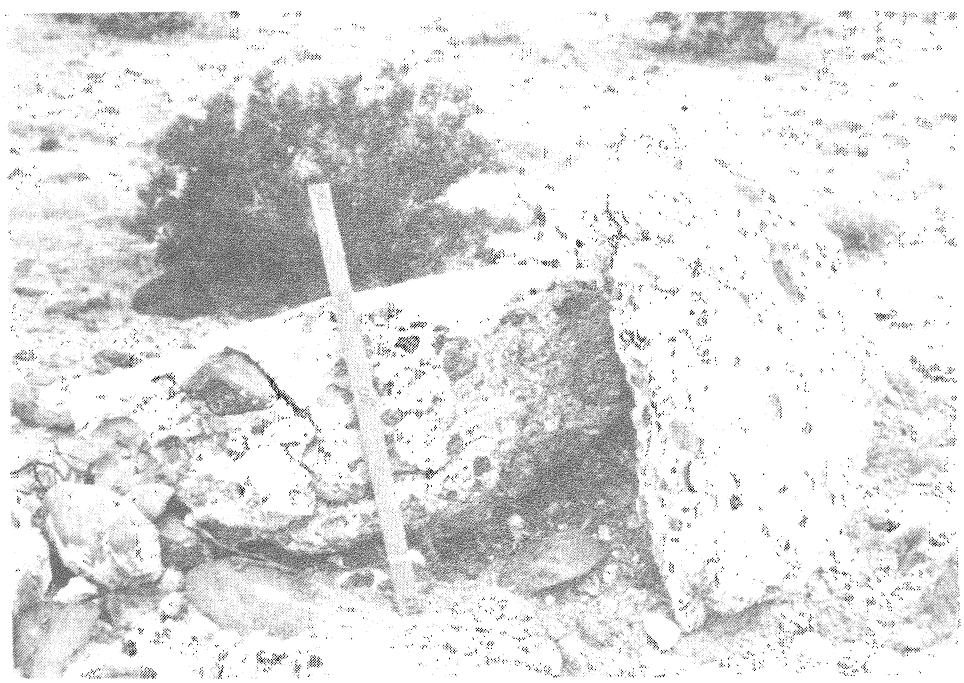


PLATE 7. ASH HILL FAN CALCRETE QUARRIED FOR PIPELINE SHOWING LAMINAR LAYER AND SEPARATION OF CLASTS.

CONCLUSIONS

The evidence presented in this paper and recent work by others together argue for the existence of a major lake filling Death Valley at some time in the Pleistocene. This is supported by the following:

(1) The existence of a large fluvial discharge channel at Ash Hill where the Death Valley Watershed overflowed toward the Colorado River.

(2) An estimated peak discharge in this channel that is too large to have been derived solely from the small watersheds near it.

(3) Evidence that shore processes were at work at the overflow elevation inside the hypothesized large lake basin.

(4) Evidence that this lake may have produced lacustrine features that can presently be seen near overflow elevation at several widely-scattered sites within the hypothesized lake basin.

(5) Evidence that considerable time has passed since both the large discharge and shore process activity occurred.

(6) Evidence that much of the lake bed in the Death Valley Watershed was either tectonically quiescent or inconsequentially active throughout the Pleistocene.

(7) Evidence that, on the other hand, the mountains on the southern and western watershed perimeter appear to have experienced great tectonic activity throughout the Pleistocene, progressively enhancing their rainshadow effect by both horizontal and vertical displacement.

The evidence presented in this paper tends to confirm Miller's (1946) and Hubbs and Miller's (1948) assertions that the icthyologic evidence demands a former connection between the Death Valley system of Lakes and the Colorado River and that the connection was made by an overflowing Death Valley Lake.

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LATE QUATERNARY GEOMORPHIC HISTORY OF THE SILVER LAKE AREA, EASTERN MOJAVE DESERT, CALIFORNIA: AN EXAMPLE OF THE INFLUENCE OF CLIMATIC CHANGE ON DESERT PIEDMONTS

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INTRODUCTION

The impact of Quaternary climatic fluctuations on sedimentation, surficial processes, and soil development on desert piedmonts of the southwestern United States has been the subject of intensive study over the past two decades (e.g., Melton, 1965; Lustig, 1965; Bull, 1974, in press; Wells, 1977, 1978a, b; Schenker, 1978; McFadden and Bull, 1981; Mayer and Bull, 1981; McFadden, 1982; Morrison and Menges, 1982; Christenson and Purcell, 1982). In most of these studies, however, temporal relations between changes in climate and responses in geomorphic systems have been proposed on the basis of relative-age dating methods such as soil-profile development or stone-pavement development. Independent age control on stratigraphic, geomorphic, and pedogenic features, as well as the climatic change itself, are difficult to establish in desert basins.

Radiocarbon dating of late Quaternary Lake Mojave deposits and shorelines along the north and west margins of Silver Lake playa (Fig. 1) allow correlation of alluvial and lacustrine sequences and provide age constraints on climatic fluctuations recorded by pluvial stands of Pleistocene Lake Mojave. The primary purposes of this paper are to (1) establish the late Quaternary geomorphic history of a distal piedmont area on the western margin of Silver Lake playa, and (2) compare the timing of geomorphic events at Lake Mojave with paleoenvironmental reconstructions based on published macrofossil data and pluvial histories of other closed basins in the southwestern United States.

GEOLOGIC SETTING OF THE SILVER LAKE AREA

The playa of Silver Lake covers an area of approximately 26 km^2 along the axis of an elongate north-south trending basin (Fig. 2). The playa surface lies at an elevation of 276.5 m (907 ft) (Ore and Warren, 1971) and is flanked on the west by the Soda Mountains (maximum elevation of approximately 1,040 m (3,410 ft)). This range is composed of Precambrian gneiss, Precambrian and Cambrian metasedimentary rocks, Paleozoic and Mesozoic sedimentary and volcanic rocks, and Mesozoic(?) plutonic rocks (Troxel, 1982). In addition, large areas of this range are overlain by Tertiary gravels. The northeast piedmont of this range drains into Silver Lake playa and is mantled by alluvium derived primarily from the metamorphic, plutonic, and Tertiary conglomeratic terranes.

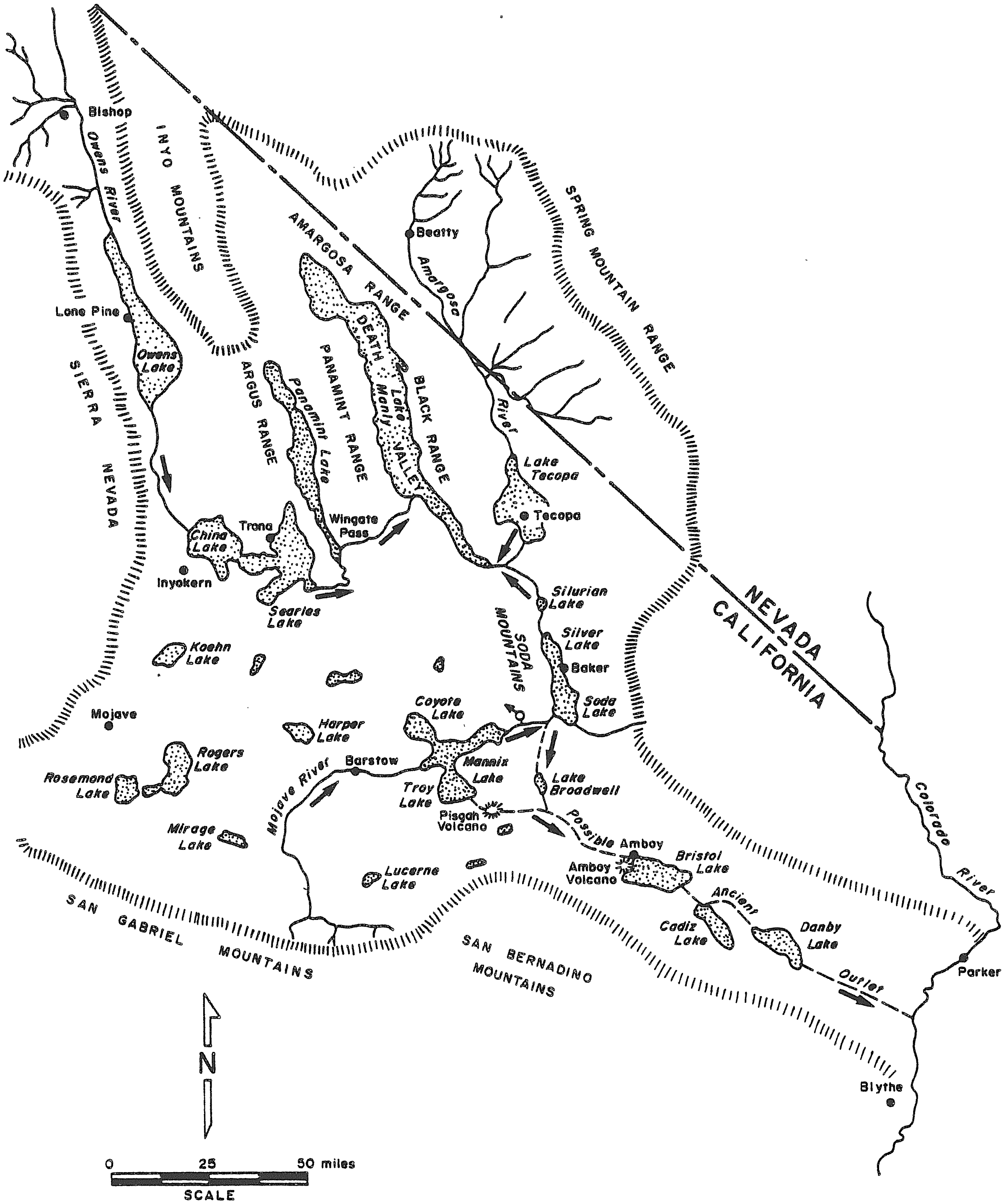

Figure 1. Pleistocene drainage paths in the Mojave Desert (modified from Blackwelder, 1954).

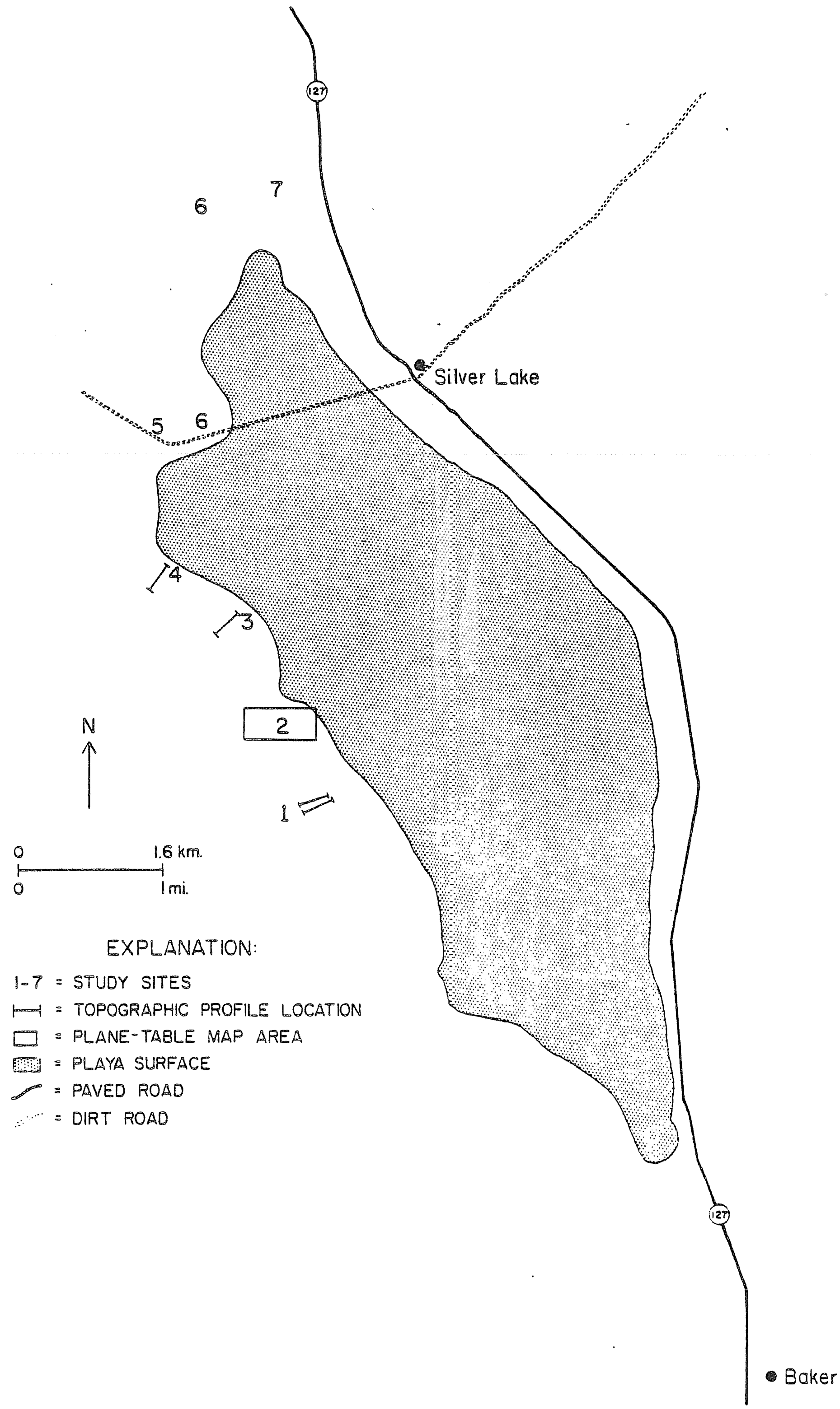

Figure 2. Index map of detailed study sites in the Silver Lake playa area. 1 = unit Qf3 topographic profile site; 2 = plane table map area site (Fig. 3); 3 = Marble promontory topographic profile site; 4 = Tufa bar topographic profile site; 5 = gravel quarry site; 6 = locations of radiocarbon samples dated during this study; 7 = overflow channel of Lake Mojave.

Silver Lake and Soda Lake playas comprise the terminus of the Mojave River which originates in the San Bernardino Mountains approximately 200 km to the southwest (Fig.1). In historic times, these playas have only contained water after extreme flooding on the Mojave River. In late Quaternary time, however, they were submerged beneath the waters of pluvial Lake Mojave, one of several pluvial lakes which formed a chain along the ancestral Mojave River. A narrow channel at the north end of Silver Lake playa was the outlet of Lake Mojave, and overflow from this channel drained northward into Lake Manly at the southern end of Death Valley (Blackwelder, 1954).

Prominent shoreline features, including beach ridges and wave-cut cliffs, occur at elevations between 285 and 288 m (935 and 945 ft) along the western and northern margins of Silver Lake playa. Several strandlines occur below these elevations and just above the playa floor. Radiocarbon dates on pelecypod shells (Anadonta californiensis) and tufa provide a detailed temporal framework for the late Quaternary chronology of this part of Lake Mojave (Ore and Warren, 1971).

METHODOLOGY

Field descriptions of soils and stratigraphic units were combined with plane table-alidade mapping, topographic profiling, and radiocarbon dating to interpret the late Quaternary geomorphic history of the Silver Lake area. Stratigraphic units were differentiated on the basis of field observations of grain size, sorting, surface morphology, and soil-profile development. Soil pits were excavated on what appear to be the most stable parts of fan surfaces or along recently exposed banks of active washes, and soil profiles were described according to methods and terminology outlined in the Soil Survey Manual (1951, 1981), The distribution and elevations of surficial deposits in a small area on the western margin of Silver Lake playa (Fig. 3) were mapped with plane table and alidade. This survey was not tied into a precise elevation; rather, elevations were defined from the general elevation of the playa floor (276.5 m, 907 ft). Topographic profiles of pluvial and historic shorelines were measured using a portable transit and stadia rod. Shorelines were identified on the basis of changes in slope and texture of the surficial deposits and were correlated on the basis of height above the playa floor and degree of preservation of geomorphic features. Ages of deposits and geomorphic features are based on the dates reported by Ore and Warren (1971); additional radiocarbon dating conducted for this study by Dicarb Radioisotope Company of Norman, Oklahoma; and cation-ratio dating of varnish on stone pavements (Dorn, this volume).

LATE QUATERNARY STRATIGRAPHY

Ten late Quaternary stratigraphic units, six alluvial fan units (Qf3, Qf4a, Qf4b, Qf5, Qf6, Qf7), three eolian units (Qel, Qe2, Qe3), and one undifferentiated lacustrine unit (Qlu), were defined for parts of the western and northern margins of Silver Lake playa (Figs. 3 and 4).

Alluvial Stratigraphy

Late Quaternary fan deposits consist of poorly to very poorly sorted bouldery sand derived from medium- to fine-grained plutonic rocks, metavolcanic and metasedimentary rocks, and Tertiary gravels. Younger fan deposits contain stream-rounded fragments of petrocalcic rubble and clasts partly coated with carbonate rinds (up to several millimeters thick) derived from older piedmont deposits or gully-bed cemented deposits. Younger deposits typically are inset into older fan deposits in the middle and upper piedmont areas and overlap older deposits in distal areas. All of these fan deposits are typically less than one to two meters thick, and locally some units are only 0.3 m thick. Preservation of soils beneath these thin deposits indicates little or no erosion prior to

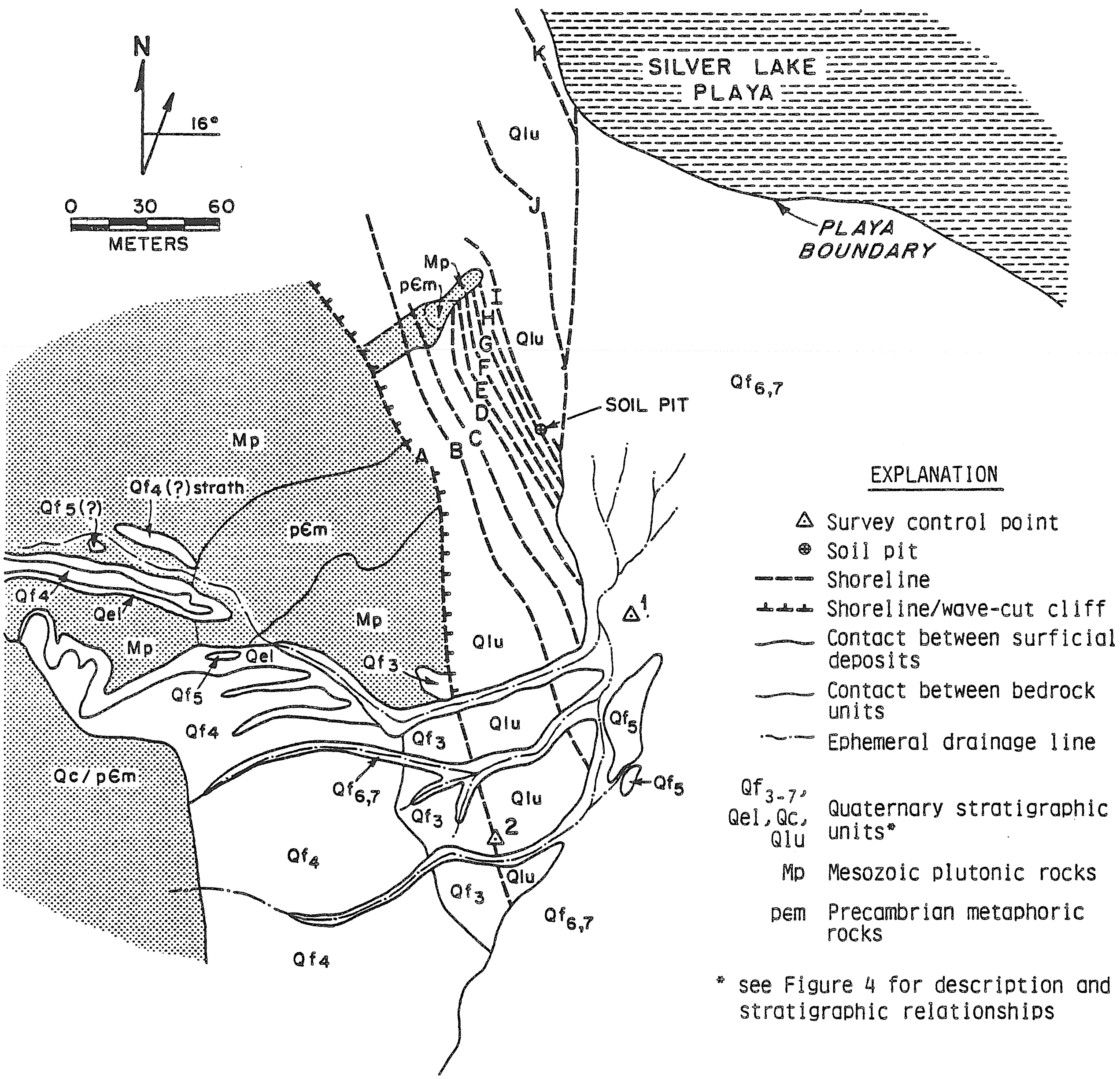

Figure 3. Detailed geomorphic map of a selected area on the western margin of Silver Lake playa. See Figure 2 for location.

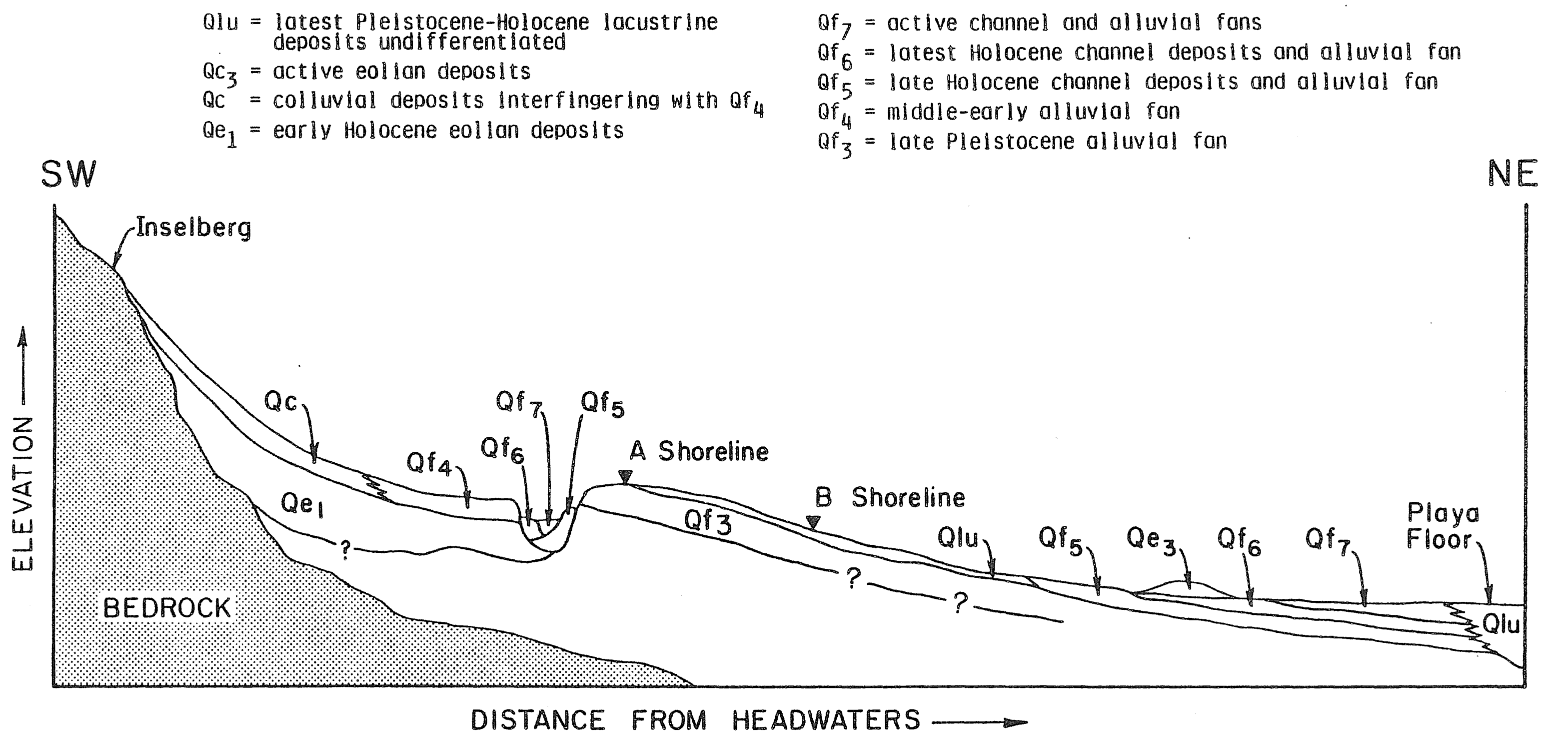

Figure 4. Schematic cross section showing topographic and stratigraphic relations on the western margin of Silver Lake playa (relations based on detailed mapping shown in Figure 3).

deposition. Thus, the distal piedmont consists of thin depositional veneers of younger over older fan deposits. Similar late Quaternary stratigraphic relations have been recognized on piedmonts in Arizona and California (Bull, 1974; McFadden, 1982; Schenker, 1978; McHargue, 1981; Wells, 1977). Holocene ages have been assigned to fan surfaces with depositional bar-and-swale topography on the basis of geomorphic and stratigraphic relations with Lake Mojave shorelines and pedologic evidence. Lithologic, topographic, and biotic similarity of the six alluvial fan units at Silver Lake permit the use of pedologic data for stratigraphic correlation and for determination of relative-age relations (Birkeland, 1974). Diagnostic morphologic features of soils and associated stone pavements are shown in Tables 1 and 2.

Units Qf7 and Qf6

These alluvial fan surfaces are common in middle and distal piedmont areas and are inset into or overlap all other deposits. Qf7 deposits occur in the active channels; whereas, unit Qf6 deposits occur adjacent to these channels in recently abandoned washes. Unit Qf6 surfaces are typically less than one meter above unit Qf7 surfaces. Both units display maximum depositional bar relief, no varnish on clast tops, and no reddening of clast undersides (Table 1). No soils occur in unit Qf7 and unit Qf6 exhibits only a weak, thin vesicular A horizon. Secondary calcium carbonate is not visible in the soil, and clasts are unweathered.

Unit Qf5

Unit Qf5 deposits are typically inset into units Qf4 and Qf3 and overlain or incised by units Qf6 and Qf7. Deposits of unit Qf5 are derived from bedrock watersheds and from older fan deposits on the piedmont. Qf5 surfaces are generally less than 1.5 m above active channels. The range in relief on depositional bars is similar to units Qf6 and Qf7; however, mean relief is slightly lower on unit Qf5. Unit Qf5 can be distinquished from younger units on the basis of incipient varnishing of clast surfaces and reddening of clast undersides (Table 1). Varnish typically covers less than 30 percent of exposed clast surfaces; thus, clast lithologies can be determined by visual inspection. Many clasts are slightly weathered, showing fracturing and partial mechanical decomposition. Unit Qf5 deposits possess weakly developed soils that possess stage I secondary carbonate morphology (carbonate morphologic classification after Gile, Peterson and Grossman, 1966; and Bachman and Machette, 1977). Powdery secondary carbonate coatings often cover indurated carbonate coatings on clasts that have been reworked from older fan deposits. Soils lack Bw horizons or secondary gypsum accumulations, typical properties of soils on older deposits.

Unit Qf4

Unit Qf4 both incises and overlaps unit Qf3. In proximal piedmont areas, Qf4 surfaces occur up to several meters above active channels; however, in distal piedmont areas, these surfaces are generally less than 1.5 m above active channels. Depositional bar relief is approximately 50 percent of the youngest fan surfaces (Table 1); however, bars are readily discernable. Varnishing on clasts is variable (depending on the clast lithology) and covers from 30 to almost 90 percent of the exposed parts of clasts in stone pavements. In many cases, resistant-clast lithologies can only be identified on freshly fractured surfaces.

Unit Qf4 can be locally subdivided on the basis of differences in soil profile development into two distinct subunits; Qf4a and Qf4b. Soils on the younger subunit, Qf4b, possess a thin (10 cm) reddish yellow Bw horizon and secondary gypsum. Granitic clasts in these soils are partly decomposed; and exposed medium-to-coarse grained plutonic rocks possess rough weathered surfaces with single-grain relief. Subunit Qf4a soils are characterized by a 35-cm thick Bw horizon, almost completely decomposed granitic clasts, stage II secondary carbonate and secondary gypsum. The presence of

Table 1.--Properties of desert pavements on late Quaternary alluvial fan surfaces along the western margin of Silver Lake playa

ALLUVIAL UNIT	INFERRED AGE	DEPOSITIONAL BAR RELIEF		VARNISH ON CLASTS[1]		REDDENING OF CLAST UNDERSIDES[1]
		Mean	Range	Color	Varnish Cover (%)	
Qf3	late Pleistocene	0.08	0.06-0.09	5YR 2.5/1	100	2.5YR 5/6
Qf4a	early Holocene	0.12	0.06-0.27	5YR 2.5/1	>80	2.5YR 5/6
Qf4b	middle to late Holocene	n.m.[2]	n.m.	7.5YR 2/1	>30 & <80	5YR 5/8
Qf5	late Holocene	0.22	0.11-0.37	7.5YR 2/1	<30	5YR 6/8
Qf6	latest Holocene	0.25	0.11-0.41	no varnish	0	no reddening

[1]Measured on fine-grained mafic plutonic rocks and metavolcanics
2Not measured

Table 2.--Summary of morphologic characteristics of soils on late Quaternary alluvial fan surfaces along the western margin of Silver Lake Playa

ALLUVIAL UNIT	B HORIZON THICKNESS (cm)	MAXIMUM RED[1] COLOR	MAXIMUM MOIST[2] CONSISTENCE	CLAY[2] FILMS	MAXIMUM[2] SOIL STRUCTURE	SECONDARY CARBONATE MORPHOLOGIC STAGE[3]
Qf3	42	7.5YR 6/6	ss, sp	3pf co	3sbk	2,3
Qf4a	39	7.5YR 6/6	ss, sp	co	3sbk	2
Qf4b	10	7.5YR 7/6	ss, np	co	m	1
Qf5	0	10YR/ 7/4	ns, np	n.o.[4]	m	1
Qf6	0	10YR/ 7/4	ns, np	n.o.	m	-

[1]Soil colors from Munsill Soil Color Chart

[2]Terminology follows Soil Conservation Service (1981)

[3]Morphology stage after Gile et al; 1966

[4]Not observed

secondary gypsum, carbonate, and much of the clay in these soils is interpreted to be the result of high rates of aerosolic dust influx. Probable local sources include upwind playas and distal piedmont areas (McFadden et al., this volume). High salt contents in aerosolic dust and/or precipitation may locally cause rapid mechanical weathering of Holocene depositional bars. These bars are typically subdued on the oldest Holocene and youngest Pleistocene surfaces. Fine-grained, resistant clasts are almost completely varnished (greater than 80 percent), and the varnish is more reddened (5YR 2.5/1) than on most clasts in the stone pavements of younger units. Locally, deeply pitted surfaces resembling tafoni have formed on mafic plutonic rocks on both Qf4a and Qf4b surfaces.

Unit Qf3

Unit Qf3 is the oldest subaerially exposed fan recognized on the piedmont flanking the western and northern margins of Silver Lake playa. All younger units either incise or overlap this unit. Unit Qf3 is best exposed in middle piedmont areas, although it occurs locally near the high shoreline of Lake Mojave. Unit Qf3 surfaces display the best developed stone pavements, and depositional bars are almost completely obliterated. Where bars are recognizable, depositional relief is due to single clasts and not to aggregates of clasts (Table 1). Stone pavements contain solution-fretted clasts of petrocalcic rubble, and varnish is almost continuous on resistant fine-to medium grained lithologies. Soils on unit Qf3 surfaces are moderately developed. B horizons are not substantially better developed than the B horizons of unit Qf4a soils; however, B horizons locally possess thin clay films that coat ped and grain surfaces and secondary carbonate locally attains a stage III morphology (Table 2). Many granitic clasts are completely decomposed in the soil and few granitic clasts are preserved at the surface.

Eolian Stratigraphy

Three eolian units are distinquished on the basis of stratigraphic position, soil development, and preservation of primary sedimentary structures. All deposits are composed of fine-to-medium grained, subrounded, well-sorted sand. These deposits range in thickness from a few centimeters to a few meters. The oldest eolian unit, Qel, underlies unit Qf4 in the detailed map area of Figure 3 and on parts of the piedmont to the north. Although there is no soil developed in unit Qel, secondary calcium carbonate forms cylindrical aggregates of sand that may be related to pedogenesis in unit Qf4 deposits. The lack of soil development suggests rapid deposition of Qf4 over Qel. Unit Qe2 has only been identified in a gravel quarry at the north end of Silver Lake playa (Fig.2). This unit overlaps the youngest dated lacustrine deposits (8,350 $\pm$ 300 yr B.P., Ore and Warren, 1971). It is approximately 30 cm thick and possesses a weakly developed soil with an 8 cm thick Bw horizon (7.5YR 6/6 dry) and stage II carbonate morphology. Unit Qe2 fills a shallow channel that cuts all prehistoric Lake Mojave shorelines; thus it postdates the final stand of Lake Mojave. The youngest eolian unit, Qe3, covers large areas of the modern land surface in the Silver Lake area. It is characterized by eolian depositional forms (dunes and sandsheets) and by the preservation of primary sedimentary structures. Qe3 is equivalent in age to units Qf6 and Qf7.

Lacustrine Deposits

Deposits of Lake Mojave include beach and offshore facies as well as interlacustrine playa-surface facies. These deposits were not differentiated because of limited exposures and lack of any significant break in the stratigraphy of the Lake Mojave deposits. Ore and Warren (1971, p. 2559) divide the lake deposits into 10 stratigraphic units on the basis of an exposed sequence in a gravel quarry at the north end of Silver Lake playa (Fig. 2). They interpret this sequence to represent lacustrine and interlacustrine depositional events. However, radiocarbon dates from this site and from other sites in the area (Table 3) indicate no major break in the lacustrine record from approximately 15,300 to at least 10,000 yr B.P. In addition, the only major

Table 3.--Readiometric and cation-ratio dates on lacustrine deposits and landforms of the Silver Lake playa area

AGE (yr B.P.)	MATERIAL/FEATURE DATED	SOURCE
8,350 ± 300	tufa in gravel quarry exposure[4]	1
9,160 ± 400	tufa in gravel quarry exposure	
9,340 ± 140	shells in beach deposits	1
10,000 ± 30	shells in gravel quarry exposure	1
10,260 ± 400	shells in gravel quarry exposure	1
10,270 ± 160	shells in beach deposits	1
10,580 ± 100	shells in gravel quarry exposure	1
10,850 ± 75	tufa encrusting boulder on shore	2
10,870 ± 450	tufa in gravel quarry exposure	1
11,320 ± 120	tufa in gravel quarry exposure	1
11,630 ± 500	tufa in gravel quarry exposure	1
11,860 ± 95	shells in beach deposits	2
12,450 ± 160	shells in beach deposits	1
13,040 ± 120	tufa in gravel quarry exposure	1
13,150 ± 350	shells in gravel quarry exposure	1
13,190 ± 500	tufa in gravel quarry exposure	1
13,290 ± 550	shells in gravel quarry exposure	1
13,620 ± 100	shells in beach deposits	1
13,670 ± 550	shells in gravel quarry exposure	1
14,550 ± 140	shells in beach ridge deposit	1
15,350 ± 240	shells in beach ridge deposit	1
36,600	cation-ratio date on unit Qf3	3
6,600	cation-ratio date on unit Qf5	3

[1] Ore and Warren (1971)

[2] Dicarb Radioisotope, this study

[3] Dorn, et al (this volume)

[4] see Figure 2 for quarry location

unconformity that occurs within this section is at the top of the lacustrine sequence where unit Qe2 overlies the Lake Mojave deposits. The radiocarbon dates and stratigraphic relations suggest that only one significant lowering of Lake Mojave (between approximately 10,000 and 9,500 yr B.P.) could have occurred between 15,350 ± 240 and 8,350 ± 300 yr B.P.

LAKE MOJAVE SHORELINES

Geomorphology

Eleven shorelines are preserved along the western and northern margins of Silver Lake playa (Fig. 5, Table 4). From the highest to the lowest these shorelines are designated A through K (intermediate shorelines which are only locally preserved are designated by a prime such as C'). These shorelines can be reliably correlated along the western margin of the playa on the basis of height above the playa floor. The highest shorelines, A and A', stand approximately 9 to 11 m above the playa floor. Several prominent wave-cut cliffs, beach ridges, and beaches are associated with these shorelines. Beaches are typically veneered with well-developed pavements and are locally dissected. Shorelines A and A' are best preserved on bedrock promontories and to a lesser extent on older alluvial fan deposits (unit Qf3). The next highest shorelines, B and B', stand approximately 7.0 m (± 0.3 m) above the playa floor. No prominent beach features are associated with the B and B' shorelines suggesting that the lake stand at this lower level was of short duration. Nine weakly expressed strandlines occur below shorelines B and B', from 6.8 to 0.5 m above the playa floor (Table 4).

Shoreline Ages

Shorelines A and A' range in age from approximately 15,500 to 10,500 yr B.P. on the basis of Ore and Warren's (1971) radiocarbon dates which were verified by additional radiocarbon dates obtained during this study (Table 3). Eight of the 18 radiocarbon dates which span this time period can be topographically related to the A shorelines. It was probably during the latter part of this period that the outlet channel was cut to the 285 m (936 ft) level. The prominence of shorelines A and A' indicates either multiple highstands as suggested by Ore and Warren (1971) or a single highstand of long duration.

Shorelines B and B' are correlated with the last and lowest stand of Lake Mojave. This stand (283 m, 930 ft) was estimated by Ore and Warren (1971) to have occurred between 8,500 to 7,500 yr B.P. However, Ore and Warren report dates of 10,270 ± 160 and 9,340 ± 140 yr B.P. on shell materials from beaches developed a few feet below the B level suggesting that by approximately 10,500 yr B.P Lake Mojave's shoreline was at the B level or lower. In addition, several cultural features (alignments of boulders oriented perpendicular to shore), occur below the level of shoreline B and above the level of shoreline C. These features were most likely constructed in shallow water during the lake stand at shoreline B and not in four to five meter deep water during the high stand at shoreline A. Therefore, the most likely age for shoreline B is between 10,500 and 8,000 yr B.P. However, the lack of prominent shoreline features at the B level suggests a short-lived lake stand lasting, perhaps, for only a few hundred years or less.

Shoreline C is the only shoreline between the B shorelines and the strandlines formed during historic flooding. Whether shoreline C represents a post-B regressive lake phase, a middle-to-late Holocene lake stand, or a prehistoric flood event is not known. Shoreline D occurs at the highest historic flood level; shorelines E through K occur below this level and, therefore, are correlated with very recent events (Table 4).

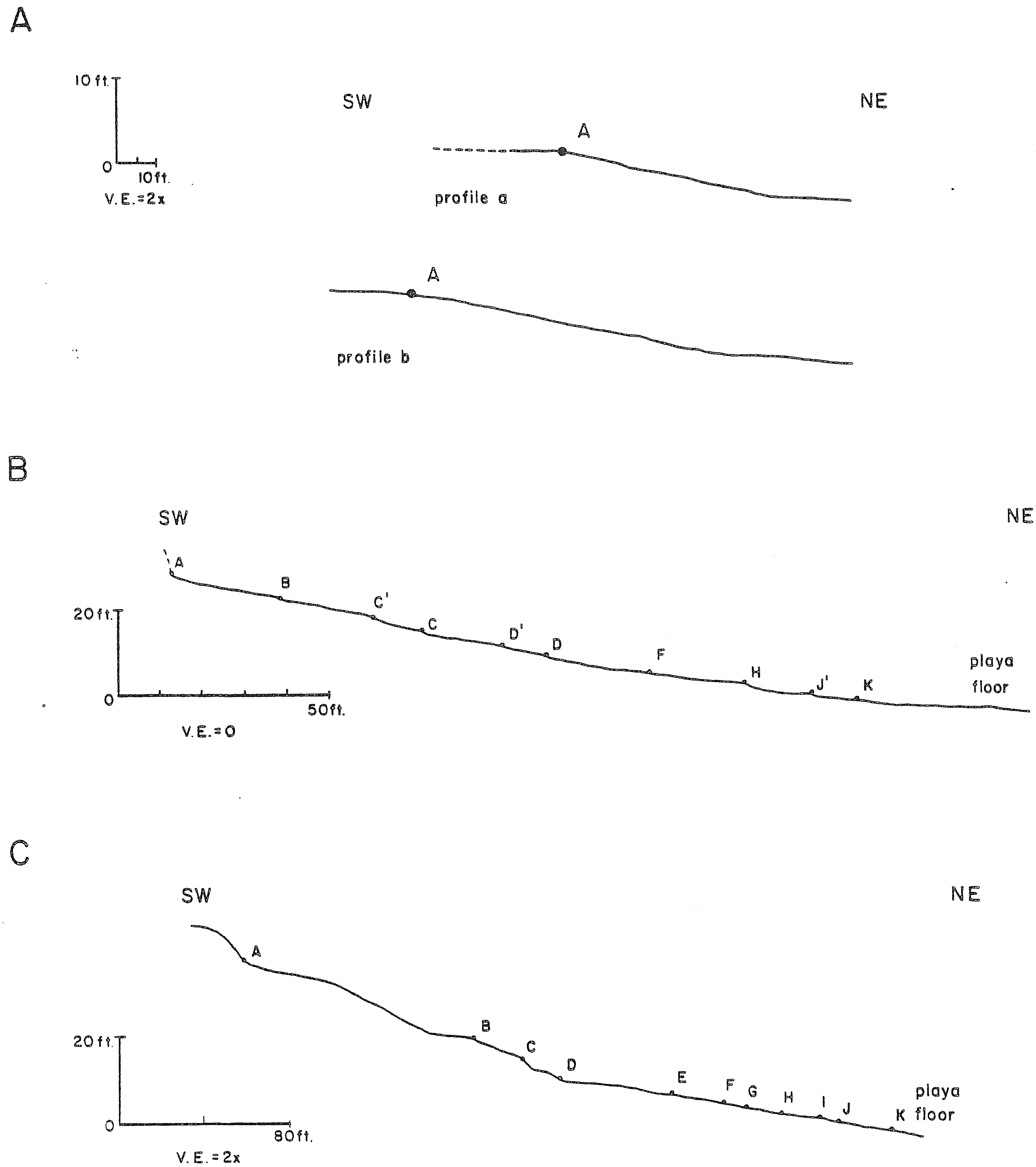

Figure 5. Topographic profiles of: (A) a shoreline and shoreface developed on unit Qf3, (B) shorelines developed on bedrock, and (C) shorelines developed on beach deposits. See Figure 2 for locations.

Table 4.--Height and ages of shorelines related to late Quaternary and historic events of the Silver Lake playa region

HEIGHT OF SHORELINE ABOVE PLAYA (m/ft)					
Shoreline	Map Area[1]	Marble Promontory[1]	Tufa Bar[1]	AGE OF SHORELINE	HEIGHT OF GEOMORPHIC FEATURES & HISTORIC SHORELINES (m)
A	10.1/33		11.3/37	latest Pleistocene	11.0-- overflow level
A'		8.8/29		(15,500-10,500 yr B.P.)	to prominent beaches, beach
B	7.3/23.9	7.0/23			8.5 ridges, wave cut cliffs
B'		6.8/22	6.7/22	early Holocene	7.3-- level of alluvial unit
				(10,500-8,000 yr B.P.)	Qf4a near playa margin
C	4.4/14.4	4.9/16	4.9/16	middle to	
C'		4.0/13		late Holocene(?)	
D	3.1/10.3	3.4/11	3.4/11	historic	3.1-- January, 1916 to
					July, 1917 level
E	2.4/8.0		2.6/8.5	historic(?)	
F	2.2/7.1	2.1/7.0	2.0/6.5	historic	2.1-- March, 1938 to
					September, 1939 level
G	1.8/5.8		1.8/6.0	historic	1.8-- June, 1969 level
H	1.5/5.0	1.5/5.0	1.5/5.0	historic(?)	
I	1.3/4.4		1.2/4.0	historic(?)	
J	0.8/2.6	0.9/3.0	0.9/3.0	historic(?)	
K	0.6/2.1	0.6/2.0	0.5/1.5	historic(?)	

[1]See Figure 2 for location

SHORELINE STRATIGRAPHIC RELATIONS: INFERRED AGES OF ALLUVIAL AND EOLIAN UNITS

The highest shorelines (A and A') truncate unit Qf3 alluvial fans, and berms and offshore slopes occur on Qf3 surfaces at the level of these shorelines. Thus, unit Qf3 is older than 15,500 yr B.P. Moreover, Dorn (this volume) obtained a cation-ratio age of approximately 36,600 yr B.P. on rock varnish from clasts in Qf3 stone pavements. This age is consistent with the shoreline-fan relations.

Along the western and northern margins of Silver Lake, the following relations exist between unit Qf4a and Lake Mojave shorelines: (1) A and A' shorelines do not truncate Qf4a; (2) unit Qf4a does not occur below shorelines B and B'; and (3) in two locations unit Qf4a appears to grade but not prograde beyond the B shoreline position. Therefore, unit Qf4a is interpreted to have been deposited after the high stand of Lake Mojave (at shorelines A and A'), probably during the final stand (at shorelines B and B'). The time of deposition is further constrained by the presence of unit Qel below Qf4a. Qel deposits were deposited a few meters below the B shoreline indicating that the lake level was below the level of the B shorelines during Qel deposition, then again rose to the B level before unit Qf4a was deposited. Based on these interpretations, units Qel and Qf4a were deposited sometime between 10,500 and 8,000 yr B.P. Significantly, the only apparent break in the lake record occurs between 10,000 and 9,500 yr B.P. which may represent the time of lake lowering and Qel deposition.

Units Qe2 and Qf4b cut shorelines B and C, and similar soil development on these units suggests approximately contemporaneous deposition. On the basis of these relations and relative soil development, units Qe2 and Qf4b are interpreted to be middle Holocene in age (less than 8,000 and greater than 3,000 yr B.P.).

Relative soil development indicates that unit Qf5 is much younger than unit Qf4b (Table 2). Also, a soil formed in noncalcareous bouldery piemont deposits in the Harquahala Valley of southwestern Arizona (dated at 750 $\pm$ 60 yr B.P.; S. G. Wells, unpublished data) is quite similar to the Qf5 soil (R. Ford, personal communication, 1984). A tentative cation-ratio date of approximately 6,600 yr B.P. has been obtained from varnish on surface clasts from the unit Qf5 surface (Dorn, this volume); however, this date is not consistent with the minimal soil development in this unit. The lack of weathering of subsurface clasts and the lack of varnish on surface clasts of unit Qf6 suggests that this unit is considerably younger than unit Qf5.

PALEOENVIRONMENTAL CONSIDERATIONS

Paleoenvironmental data from analysis of plant macrofossil remains in packrat (genus Neotoma) middens provide perhaps the most detailed information concerning the timing and nature of climatic changes during the late Quaternary. Several of these studies demonstrate that a juniper-pinyon-Joshua tree woodland was present throughout the Mojave and Sonoran deserts between 30,000 and 11,000 yr B.P. (Van Devender, 1973, 1977; King, 1976, Van Devender and Spaulding, 1979). This woodland was present at elevations as low as 320 m where a desert scrub community of xerophytic vegetation is now stable. After 11,000 to 10,000 yr B.P., pinyon and Joshua tree components of the woodland died out, but juniper persisted at elevations as low as 330 m until about 8,000 yr B.P. (Van Devender, 1973, 1977; King, 1976). Spaulding (1982) has recently shown that elements of the desert scrub were present in the Mojave Desert by 15,000 yr B.P. at elevations as high as 990 m. In any case, the complex mosaic of woodland and desert scrub communities was not completely eliminated from elevations below 800 to 900 m until the middle Holocene.

The plant macrofossil record suggests that the latest Pleistocene climate in desert areas south of latitude 36^{o} was characterized by milder, moister winters and cooler, dryer summers than the present climate (Spaulding, 1982; Van Devender and Spaulding, 1979). However, Galloway (1970, 1983) argues that these data do not rule out a significantly cooler and drier climate in this region. Whatever the case, two points appear to be certain: (1) vegetation change in the southwestern deserts was time transgressive, implying a transitional change in the climate from the latest Pleistocene to the Holocene; and (2) the effective moisture of late glacial and early Holocene climatic regimes was significantly greater than the effective moisture of the modern climate (McFadden and Tinsley, in press).

Greater effective moisture in latest Pleistocene and early Holocene time is supported by pluvial lake chronologies in the southwestern United States. Several brief latest Pleistocene (14,000 to 11,000 yr B.P.) high stands occurred in pluvial lakes fed from either glaciated or unglaciated source areas (Benson, 1978; Lajoie and Robinson, 1982; Smith and Anderson, 1982). These high stands lagged 4,000 to 7,000 yr behind the glacial maximum indicating that increased temperatures at the end of the Pleistocene were not accompanied by a decrease in effective moisture (Lajoie and Robinson, 1982), or that a significant increase in effective moisture occurred during this time (Smith and Anderson, 1982). Lake Mojave underwent a similar latest Pleistocene high stand, and an increase in effective moisture between approximately 15,500 and 10,500 yr B.P. was probably necessary to maintain this lake. Several pluvial lake basins fed by nonglaciated source areas also contained early to middle Holocene lakes (Bachhuber, 1982; Fleishhauer and Stone, 1982). The presence of a short-lived early Holocene lake at Silver Lake playa is inferred from the presence of shoreline B, and the relations of alluvial unit Qf4a and eolian unit Qel to this shoreline. These Holocene lakes suggest significant fluctuations in effective moisture from 11,000 to 7,000 yr B.P. and imply greater effective moisture during this period than in post-middle Holocene time. These conclusions are also consistent with the inferrences drawn from the paleobotanical evidence.

LATE QUATERNARY GEOMORPHIC HISTORY

The earliest late Quaternary event clearly recorded along the western margin of Silver Lake playa is the deposition of unit Qf3 alluvial fans. Deposition of these fans occurred prior to 15,500 yr B.P. and perhaps prior to 36,600 yr B.P. as suggested by the cation-ratio date on rock varnish. The next recorded event is the occurrence of a latest Pleistocene lake high stand (occurring as early as 15,350 $\pm$ 240 yr B.P.; Ore and Warren, 1971). Although Ore and Warren (1971) discuss lake levels higher than 11.5 m above the playa surface, no field evidence exists on unit Qf3 alluvial fan surfaces to suggest lake levels higher than shoreline A. Lajoie and Robinson (1982) documented episodic deep-water conditions in Mono Basin, California from 36,000 to 34,000, 28,000 to 24,000, and 14,000 to 12,000 yr B.P. Similar lacustral events may have occurred at Lake Mojave; however, they apparently did not exceed the level of shoreline A. The highest, latest Pleistocene stand in Lake Mojave occurred between 15,500 and 10,500 yr B.P. This event formed the prominent wave cut cliffs, beach ridges, and beaches that occur between 8.8 and 11.3 m above the playa surface. During the latter part of this time, the overflow channel was cut 1.5 m below the level of the highest stand.

Several hundred years of either low water or complete drying preceded the final Lake Mojave lacustrine event recorded by shoreline B. Field observations and detailed mapping show that a significant eolian event, deposition of unit Qel, occurred after abandonment of shoreline A and prior to the development of shoreline B. This indicates that unit Qel was deposited in a short time period during the early Holocene, perhaps between 10,000 and 9,500 yr B.P. A decrease in effective moisture at this time, resulting in decreased vegetation densities and lower lake levels or complete drying of the lake, would expose unconsolidated sands and silts in distal piedmont and axial basin areas thereby providing a readily available source of unit Qel sediments.

Following this brief interlacustrine episode, Lake Mojave filled to the level of shoreline B. Where alluvial source areas were close to the lake, unit Qf4a can be seen to grade to this shoreline; thus, it is inferred that this unit was deposited after 9,500 yr B.P. and before 8,000 yr B.P. This event marks the first major episode of alluvial fan deposition in this area since deposition of unit Qf3. In addition, unit Qf4a grades to and interfingers with colluvium on hillslopes developed on resistant metamorphic rock types. This relation records hillslope destabilization and increased sediment yield from bedrock source areas during the early Holocene.

After the early Holocene lake stand and fan deposition, Lake Mojave receded. Lake stands of 3.4 m and less appear to be related to historic stands. Two middle-to-late Holocene alluvial fan units (Qf5 and Qf6) were deposited on the distal piedmont regions, obliterating the B and C shorelines in most places along Silver Lake. Reworked clasts of pedogenic carbonate coatings, as well as the stratigraphic relation described above, suggest that the older piedmont fans (units Qf4 and Qf3) were a major source for these middle-to-late Holocene fan deposits.

GEOMORPHIC RESPONSES OF DESERT PIEDMONTS TO CLIMATIC FLUCTUATIONS

Comparison of the late Quaternary geomorphic history of the Silver Lake area with paleoclimatic reconstructions permit evaluation of the impact of climatic change during the latest Pleistocene and Holocene on fan deposition, hillslope stability, eolian activity and soil development. Changes in lake levels, eolian deposition, and alluvial fan aggradation appear to be in agreement with changes in climate inferred from plant macrofossil remains in packrat middens. Disappearance of pinyon and Joshua tree components from the plant record between 11,000 and 10,000 yr B.P. is coincident with post 15,500 to 10,500 yr B.P. high stand lowering of Lake Mojave. Following this lowering, rapid deposition of units Qel and Qf4a occurred at favorable positions on the piedmont near sediment source areas. Shortly after the early Holocene lake stand (sometime between approximately 9,500 and 8,000 yr B.P.), units Qe2 and Qf4b were deposited. These relations suggest systematic linkage among climate, lake levels, eolian deposition, and alluvial fan aggradation.

1) Reduction in effective moisture resulted in lower lake levels, exposing fine-grained deposits on distal piedmont and playa floors to active wind regimes. Changes in plant communities and vegetation densities on bedrock hillslopes or in piedmont source areas enhanced removal of stabilized colluvium on hillslopes or stabilized deposits on piedmonts resulting in an increased availability of sediment for transport out of source areas and onto the piedmont. Changes in hillslope or piedmont sediment supply and water discharge triggered piedmont aggradation. Such a model has been proposed for the arid watersheds in southern Israel (Bull and Schick, 1979) and in the eastern Mojave Desert (Bull, in press).

2) Reduction in effective soil moisture during the late Pleistocene-Holocene climatic transition probably also influenced depths of wetting and carbonate and clay translocation in soils. However, related landscape changes in the Silver Lake area probably caused even more profound impacts on pedogenesis. The increasing presence of extensive areas of deflatable deposits both locally and regionally presumably caused a sharp increase in the rate of aerosolic dust influx on favorably located piedmonts. Where a high percentage of salt-rich dust is available, secondary gypsum is eventually trapped in the decreasingly permeable soils. The presence of salts in soils accelerates mechanical decomposition of surface clasts and promotes the evolution of stone pavements. Such impacts of landscape change on soils genesis hold significant implications regarding the use of soils-geomorphic criteria for precise and detailed correlation of surficial deposits over large areas of the southwestern United States.

3) Complex response mechanisms in alluvial systems, such as those described by Schumm (1973), may occur in conjunction with climatic changes. An increased flux of eolian fines and associated aerosolic salts may accelerate hillslope instability. Eolian fines commonly mantle hillslopes (Smith, this volume). The effects of these fines and

salts on hillslopes are to promote debris flow activity (Wells, et al., 1982) and to accelerate mechanical weathering of bedrock. Such processes enhance sediment availability on hillslopes and the transport of this sediment onto the piedmont. Alluvial fan aggradation is largely controlled by sediment availability which may be linked to eolian processes as well as to climatically induced changes in vegetation.

4) Substantial changes in distal piedmont base level also occurred in response to lake-level fluctuations. These fluctuations controlled local base level for small fans with source areas close to the lake. Lowering of lake levels from the latest Pleistocene high stand resulted in minor incision of beach deposits, but little incision of distal piedmont deposits occurred with this lowering. Base-level stability at the early Holocene B shoreline restricted the vertical position of unit Qf4a where it was deposited close to the lake. Disappearance of Lake Mojave resulted in base-level lowering of distal piedmont areas to the modern playa surface. In distal areas, younger fans (Qf4b, Qf5, Qf6, Qf7) grade to this general level and occur vertically within one meter of each other near the lake margin. However, fans and washes with source areas on the middle and upper pediment probably did not respond to lake-level fluctuations. Deep piedmont dissection in upper and middle piedmont areas is recorded by large vertical separations between Qf4a and younger fan surfaces. This dissection, as well as most stream incision on the piedmont, is probably related to more complex responses of the fluvial system than climatically controlled base-level changes.

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LATE QUATERNARY PIEDMONT STRATIGRAPHY OF THE SALT SPRING HILLS AREA, EASTERN MOJAVE DESERT, CALIFORNIA

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INTRODUCTION

The objectives of this study are to 1) define the late Quaternary stratigraphy of alluvial fans flanking the Salt Spring Hills along the southwestern and northwestern margin of Pleistocene Dumont Lake, 2) correlate this stratigraphy with the piedmont stratigraphy of the Silver Lake area (Wells and others, 1984), and 3) compare the timing of geomorphic events between the Salt Spring Hills and Silver Lake areas.

GEOLOGIC SETTING OF THE SALT SPRING HILLS AREA

The Salt Spring Hills are composed of Precambrian and Cambrian metasedimentary rocks, Paleozoic metasedimentary and sedimentary rocks, and Mesozoic(?) plutonic rocks (Troxell, 1967). Drainage basins concentrated on in this study are essentially monolithologic, composed of quartzites from the Precambrian and Cambrian Stirling Quartzite and Wood Canyon Formations (Troxell, 1967; Erickson, pers. comm., 1985).

Wave-cut cliffs on colluvial and alluvial deposits along the northeast flank of the southern Salt Spring Hills indicate a high lake stand at approximately 176 m (577.5 ft). Whether this strandline was associated with a lake retained to the west and northwest by the Salt Spring Hills or a southern extension of Lake Manly is not clear. Throughout the remainder of this paper, the wave-cut cliffs and associated playa-lacustrine sediments east and southeast of the Salt Spring Hills will be referred to as Dumont Lake (Figure 1). If Dumont Lake was contained by the Salt Spring Hills, it would have comprised an area of approximately 6.6 sq km (2.5 sq mi). The playa surface lies at approximately 172.2 m (565 ft), but it is dissected along its northern and western margins by Salt Creek. Salt Creek was formerly part of the Pleistocene Mojave River system which integrated a number of lakes, including Dumont Lake and Silver Lake, along its course northward into Death Valley (Figure 2)(Blackwelder, 1954).

METHODOLOGY

Late Quaternary stratigraphic units for five alluvial fan complexes in the Salt Spring Hills area were delineated in the

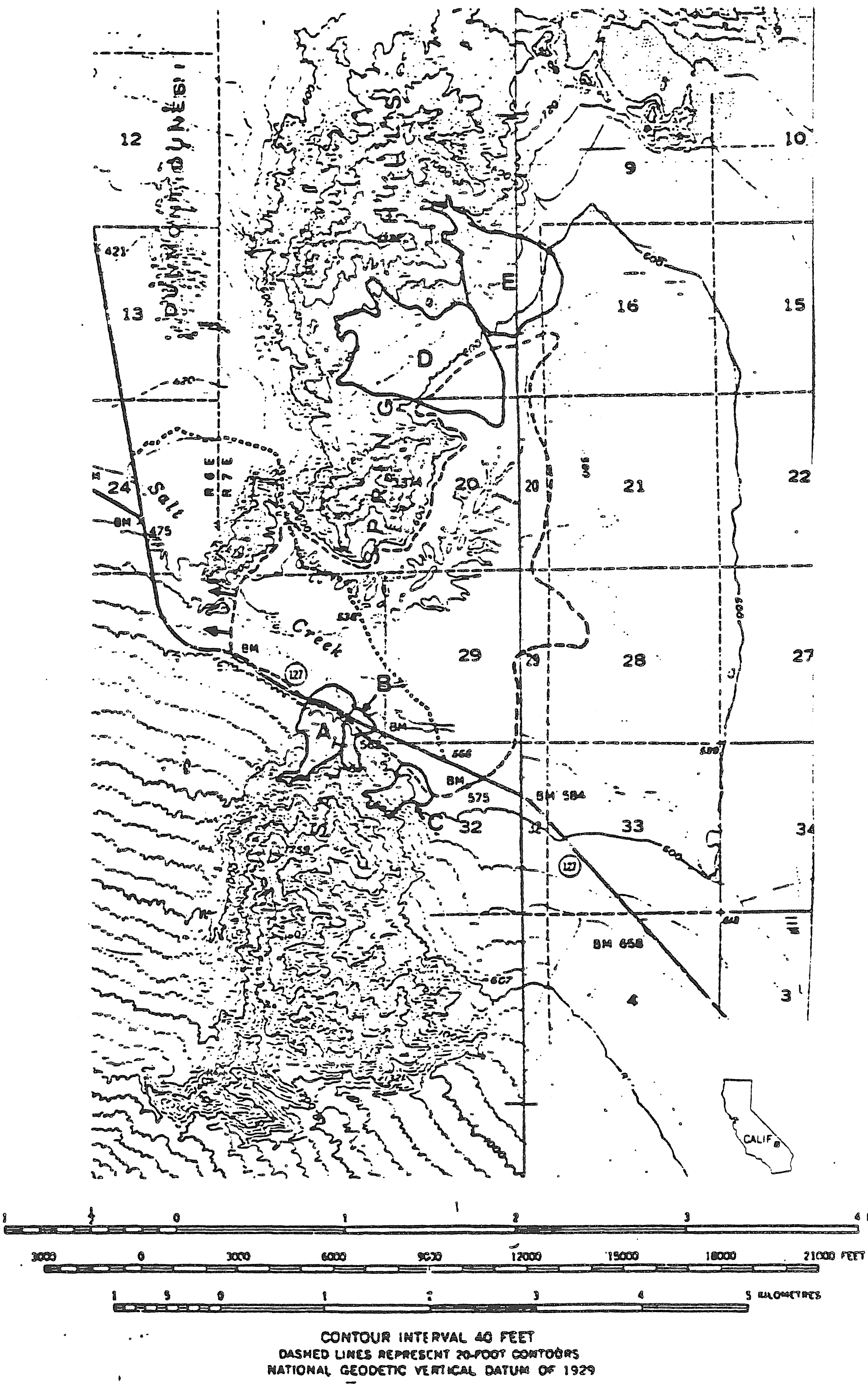

Figure 1. Salt Spring Hills study area. Alluvial fans A-E on which late Quaternary stratigraphy was established are outlined. Heavy dashed line approximates the extent of Dumont Lake.

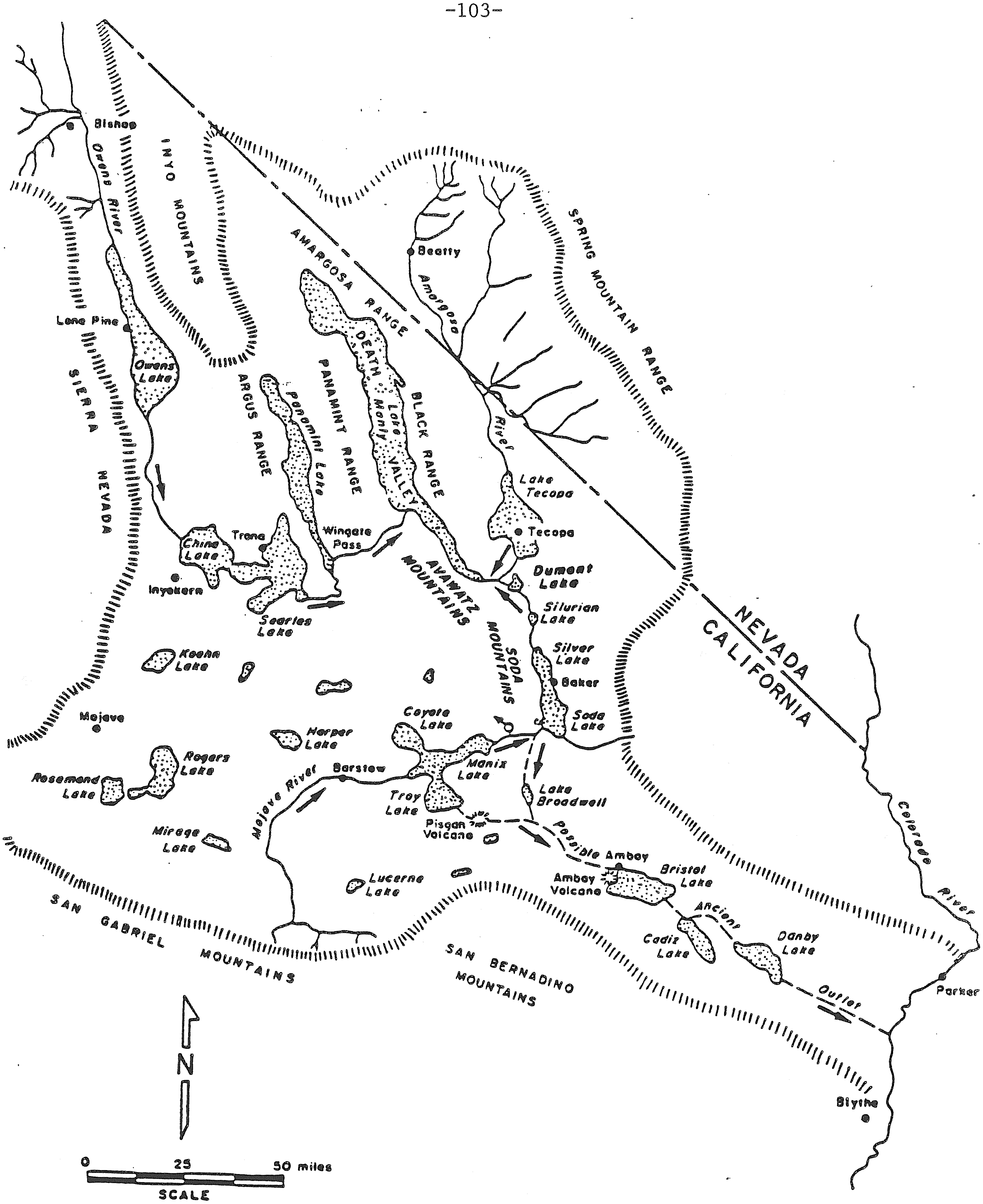

Figure 2. Pleistocene lakes and drainage systems in the Mojave Desert (modified from Wells and others, 1984).

field on the basis of soil profile development, desert pavement and varnish characteristics, surface morphology, and geomorphic expression. Topographic maps and profiles were made by plane table and alidade and laser theodolite to detail the distribution and elevation of surficial deposits on smaller fans (less than 1 sq km). Sedimentologic analysis of deposits consisted of measuring-tape transects, as described by Gerson and others (1978), in cut banks along which intermediate grain diameter, morphology, and lithology of clasts were described.

LATE QUATERNARY STRATIGRAPHY

The late Quaternary stratigraphy in the Salt Spring Hills area consists of seven alluvial units (Qf1, Qf2, Qf3, Qf4a, Qf4b, Qf5, Qf6), two eolian units (Qe1, Qe2) and one undifferentiated lacustrine-playa unit (Qlpu).

ALLUVIAL FAN UNITS

The late Quaternary alluvial fan deposits are very poorly sorted cobbly to bouldery sands. The gravel fraction, composed of predominantly quartzite clasts, is also poorly to very poorly sorted; mean grain diameters range from 3.8 cm to 11.7 cm for the intermediate axis. Quartzite clasts within the fan deposits are unweathered and angular and exhibit remnant varnish and hardened rinds of carbonate.

Younger fan deposits are typically inset into older fan deposits in proximal and medial fan areas. In distal fan areas, younger fan deposits are inset into, converge with, or overlap older fan deposits, depending on incision by Salt Creek. Thicknesses of alluvial fan deposits are greater than 2 m in proximal fan areas and decrease downfan.

The following alluvial stratigraphy has been compiled from five alluvial fans (Figure 1, A-E); the entire sequence may not be present on a given fan complex.

Unit Qf5 and Qf6

Unit Qf6 and Qf5 are defined as the active channel deposit and the most recently abandoned surface respectively. Qf6 deposits are typically inset into Qf5 deposits along major fan drainages; Qf5 deposits are inset into older fan deposits in upper, middle, and lower fan areas, but may also converge with or overlap older fan deposits in lower fan areas. Qf6 and Qf5 surfaces exhibit maximum depositional bar relief (Table 1). Abrasion of varnished and reddened surfaces on surface clasts suggests that varnish and reddening are predominantly inherited from older alluvial surfaces and hillslopes, however, Qf5 surface clasts do show evidence of incipient varnish development covering abraisons as well as in situ reddening of clast undersides (Table 1). Soils are not developed in Qf6 deposits and soil development

Table 1. Characteristics of surface pavements on late Quaternary alluvial fan deposits in the Salt Spring Hills area.

Alluvial Unit	Depositional Bar Relief		Varnish Cover (%)	Reddened Clast Undersides	
	Mean (cm)	Range (cm)		% of Clasts Reddened	Color[1]
Qf1	single clast relief n.m.[2]		95.8± 7.0	88	2.5YR5/8
Qf2	15.3	10.0-24.0	80.2±19.5	100	2.5YR4/8
Qf3	15.2	8.5-24.0	71.4±27.3	96	2.5YR5/8
Qf4a	20.6	10.0-39.0	38.0±29.7	100	2.5YR5/8
Qf4b	26.9	18.0-36.5	40.8±36.7	92	5YR5/8
Qf5	35.0	24.0-63.5	18.8±26.2	68	5YR5/8
Qf6	34.7	17.0-46.0	n.m.[2]	0	--

1 Colors from Munsell Soil Color Chart

2 Not measured

Table 2. Morphologic properties of soils developed in late Quaternary alluvial fan deposits in the Salt Spring Hills area.

Alluvial Unit	Thickness of B Horizon[1] (cm)	Maximum Red Color[2]	Maximum Moist Consistence[3]	Clay Films[3]	Maximum Structure[3]	Secondary $CaCO_3$ Morphologic Stage[4]
Qf1	50	7.5YR6/6	ss,sp	--	2 c sbk	3
Qf2	26 to 52	7.5YR5/6	ss,sp	co	2 f sbk	2
Qf3	7	7.5YR5/6	s,p	2 n pf	2 c-vc sbk	2
Qf4	0	10YR6/6	ns,np	n.o.[5]	1 m pl-sbk	1
Qf5	0	10YR6/4	ns,np	n.o.[5]	1 f-m gr	1

1 Argillic or cambic portion of B horizon

2 Soil colors from Munsell Soil Color Chart

3 Terminology follows Soil Conservation Service (1981)

4 Morphology stage after Gile and others (1966); Bachman and Machette (1977)

5 Not observed

in Qf5 deposits is limited to a thin vesicular A horizon.

Unit Qf4a

In proximal and medial fan areas, Qf4a deposits are inset into units Qf1 and Qf2; in distal areas, unit Qf4a is inset into, converges with, or overlaps older fan deposits (Qf1, Qf2, Qf3). Height of unit Qf4a above the active wash ranges from 1 to 2.5 m in upper fan areas and decreases to 0 to 1 m in the distal fan areas. Relief on remnant bar and swale topography on the Qf4a surface averages approximately 20.6 cm, considerably less than younger surfaces. Incipient desert pavement is characteristic of the fine gravel fraction in swales. Varnish covers an average of 38 percent of exposed clast surfaces. Most clast exhibit reddening (2.5YR5/8) on undersides (Table 1). Soils developed in Qf4 deposits are characterized by a vesicular A horizon (1-3 cm thick) that forms polygonal peds and secondary carbonate morphology of stage 1 (Table 2)(stage classification after Gile and others, 1966; Bachman and Machette, 1977). Qf4a deposits are present on all fans in the study area.

On large fan complexes, Qf4a can be differentiated from a younger fan deposit, designated Qf4b because soil and surface characteristics more closely resemble those of Qf4a than Qf5. Qf4b is typically inset inset in Qf4a, the two units being delineated only where this relationship is clear. Relief between Qf4a and Qf4b is approximately 0.5 m in upper and middle fan areas and converges in the downfan direction.

Unit Qf3

Unit Qf3 deposits have been identified on one fan (Fan B, Figure 1) in the study area. Qf3 deposits are not clearly inset into Qf2 deposits, however, the apex of Qf3 deposition is downfan from that of Qf2 and the slope of the Qf3 surface is significantly less than Qf2, suggesting separate fan segments (Bull, 1964). Depositional bar relief is less than half the relief of Qf5 and Qf6 surfaces. Mean percent varnish cover on exposed clast surfaces is 71.4 percent (Table 1). Soils exhibit a 7-cm thick Bw horizon with a maximum red color of 7.5YR5/6 with discontinuous clay films on ped faces. Secondary carbonate morphology is stage 2 (Table 2). Although the Qf3 surface crosses a known shoreline elevation, the surface is not truncated by the shoreline. Zones of bleached and deeply pitted (up to 10 cm) quartzite clasts are present on the Qf3 surface.

Unit Qf2

In basins where Qf2 deposits do not form the oldest subaerially exposed fan segment, unit Qf2 is inset into unit Qf1. The Qf2 surface lies from 2 to 6 m above the active wash in proximal fan areas; the Qf2 surface and active wash converge toward the distal fan area. All younger units either incise or overlap this surface. Mean depositional bar relief is indistinguishable from the Qf3 surface, and although the mean percent of varnish cover on surface clasts is greater on the Qf2 surface (80.2 percent), this difference is not apparent in the field (Table 1). Soils developed in Qf2 deposits have Bw horizons 26- to 52-cm thick with colloidal stains on sand grains and granules and stage 2 carbonate morphology. Unit Qf2 is truncated by one apparent shoreline at approximately 176 m (577.5 ft). Surface quartzite clasts both above and below the shoreline are bleached and pitted, similar to clasts on the Qf3 surface.

Unit Qf1

Unit Qf1 deposits are the oldest subaerially exposed alluvial fan deposit in the study area (Fans D and E, Figure 1). In proximal fan areas, Qf1 surfaces grade into hillslopes and are more than 7 m above the active wash. All younger fan deposits are inset into the Qf1 deposit in upper and middle fan areas; in lower fan areas, younger are inset into or overlap the Qf1 deposit. Depositional bars are unrecognizable in middle and lower fan areas with surface relief due strictly to single clast relief (Table 1). Desert pavement is strongly interlocking with varnish typically covering greater than 90 percent of exposed clast surfaces (Table 1). Soils are characterized by a 50-cm thick Bt horizon with stage 3 secondary carbonate morphology. The surface of unit Qf1 does not extend below an elevation of approximately 177 m (580 ft) in the study area and shows no evidence of truncation at its lowermost extent.

EOLIAN UNITS

Two eolian units (Qe1, Qe2) have been identified in the Salt Spring Hills area. Unit Qe1 is stratigraphically beneath Qf2. In distal fan areas, where Qf2 is less than 15 cm thick, soils formed in the Qf2 deposit are superimposed on the Qe1 deposit with original eolian depositional structures being preserved at depth. In middle and upper fan areas, where Qf2 deposits thicken to as much as 2 m, no soils are developed in Qe1, suggesting deposition of Qf1 immediately following Qe1 deposition. Qe1 has been identified only along the northern and northwest margin of Dumont Lake. Qe2 is an active eolian deposit which forms dunes and sandsheets in the study area.

LACUSTRINE-PLAYA UNITS

Lacustrine-playa sediments consist of fine sands, silts, and clays and are typically light yellowish brown (10YR6/4). Extensive dissection of Qlpu sediments along the western margin of Dumont Lake expose up to 4 m of lacustrine-playa deposition. Buried channels with gravel and sand lenses suggest periods of subaerial fluvial erosion and deposition, followed by lacustrine-

Table 3. Correlation of late Quaternary Piedmont stratigraphy between the Salt Spring Hills and Silver Lake areas.

Silver Lake Stratigraphy (Wells and others, 1984)	Inferred Age[1]	Salt Spring Hills Stratigraphy
Qf3	late Pleistocene (36,600 - 15,500 yr B.P.)	Qf1
Qe1		Qe1
Qf4a	early Holocene (10,500 - 8,000 yr B.P.)	Qf2
Qe2		
Qf4b	middle to late Holocene (8,000 - 3,000 yr B.P.)	Qf3
Qf5	late Holocene (3,000 yr B.P.)	Qf4a
		Qf4b
Qf6	latest Holocene	Qf5

1 Ages reported in Wells and others (1984)

toward the distal fan area. All younger units either incise or overlap this surface. Mean depositional bar relief is indistinguishable from the Qf3 surface, and although the mean percent of varnish cover on surface clasts is greater on the Qf2 surface (80.2 percent), this difference is not apparent in the field (Table 1). Soils developed in Qf2 deposits have Bw horizons 26- to 52-cm thick with colloidal stains on sand grains and granules and stage 2 carbonate morphology. Unit Qf2 is truncated by one apparent shoreline at approximately 176 m (577.5 ft). Surface quartzite clasts both above and below the shoreline are bleached and pitted, similar to clasts on the Qf3 surface.

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Qf4a	early Holocene (10,500 - 8,000 yr B.P.)	Qf2
Qe2		
Qf4b	middle to late Holocene (8,000 - 3,000 yr B.P.)	Qf3
Qf5	late Holocene (3,000 yr B.P.)	Qf4a
		Qf4b
Qf6	latest Holocene	Qf5

1 Ages reported in Wells and others (1984)

playa deposition. Differentiation of Qlpu sediments is beyond the scope of this study.

CORRELATION OF SALT SPRING HILLS AND SILVER LAKE STRATIGRAPHIES

Jenny (1940) has shown that by holding the soil forming factors of climate, organisms, topography, and parent material constant, pedogenic trends can be used for determination of relative-age relations and for stratigraphic correlation. Climate, organisms, and topography are assumed constant between Salt Spring Hills and Silver Lake. Parent material of the alluvial units is not the same between study areas; quartzitic rocks in the Salt Spring Hills area are more resistant than the fine- to coarse-grained plutonic rocks in the Silver Lake Area (Hooke and Rorher, 1977). Qualitative observation of alluvial deposits in the two areas suggests, however, that in situ weathering of the deposits is minimal in Holocene time and that soil plasma (clay and carbonate) is externally derived from aerosolic dust influx, as suggested by McFadden and others (1984) for soils southeast of the Silver Lake area. Assuming rates of dust influx have been equal between the Salt Spring Hills and Silver Lake areas, B horizon development (Gile and Grossman, 1968; 1979) and secondary carbonate morphology (Gile, Peterson, and Grossman, 1966; Bachman and Machette, 1977) can be used to correlate stratigraphic units between the two areas. Trends in depositional bar denudation, varnish cover on surface clasts, and development of stone pavement were also used. The correlation of alluvial and eolian stratigraphic units between the Salt Spring Hills area and the Silver Lake area is shown in Table 3. (It may benefit the reader to compare Tables 1 and 2 of this report with Tables 1 and 2 of Wells and others, 1984; this volume.)

COMPARISON OF TIMING OF GEOMORPHIC EVENTS BETWEEN THE SALT SPRING HILLS AND SILVER LAKE AREAS

Correlation of stratigraphic units between the study area and the Silver Lake area suggests synchronization of major geomorphic events. Qf1 deposits of Salt Spring Hills and Qf3 deposits of the Silver Lake area, which exhibit similar soil development and surface morphology, are the oldest subaerially exposed deposits; all younger deposits are either inset into or overlap these units. Contemporaneous deposition of units Qf2 and Qf4a in the Salt Spring Hills area and the Silver Lake area respectively is indicated by similarities of soils and surface characteristics and the presence of an eolian deposit stratigraphically beneath both alluvial deposits. In the Silver Lake area, shoreline stratigraphic relations of unit Qf4a suggest that alluvial deposits were graded to an early Holocene lake stand (10,500 - 8,000 yr B.P.)(Ore and Warren, 1971; Wells and others, 1984). In the Salt Spring Hills area, Qf2 alluvial deposits along the southwestern margin of Dumont Lake (Fan 1, Figure 1) are truncated by a high lake stand indicating Qf2 deposition predated the high stand of Dumont Lake. Absence of the strandline on Qf2 surfaces along the northern margin may be due to the embayed nature of the lake in this area. If Qf2 and Qf4a deposits are indeed time equivalent, the delayed filling of Dumont Lake may be related to the timing of damming of the Dumont Lake outlet or backfilling of the much larger Lake Manly to the north. Similarity of soil development and surface morphology of units Qf2 and Qf3 in the Salt Spring Hills area suggest that Qf3 deposition closely followed Qf2 deposition, but that unit Qf3 was not associated with a known lake stand; these relations are consistant with deposits Qf4a and Qf4b in the Silver Lake area (Wells and others, 1984). Eolian deposits contemporaneous with alluvial unit Qf3 have not been identified in the Salt Spring Hills study area. Unit Qf4a deposits in the Salt Spring Hills area are correlative with Qf5 deposits along Silver Lake on the basis of A horizon development and percent varnish cover. In both areas, these deposits are inset into older alluvial units in upper and middle piedmont areas and are inset into or overlap older deposits in distal fan areas. In the Salt Spring Hills area, Qf4b deposits are very similar to but inset into unit Qf4a deposits on larger fan complexes; this relationship is not observed in the Silver Lake area. Younger alluvial and eolian stratigraphic relationships are similar between both areas.

The synchronous nature of geomorphic events between the Salt Spring Hills and Silver Lake areas is apparent for large piedmont systems, suggesting a linkage between climate and piedmont evolution as discussed by Wells and others (1984). However, examination of individual fans in the Salt Spring Hills area suggests both spatial and temporal variations in fan evolution. For example, whereas unit Qf1 deposits have been identified only on Fans D and E (Figure 1), the two largest fans examined, Qf2 deposits have been found on all five fans in the study area. Such variation may represent the complexity of alluvial response to external system changes such as climatic change, as described by Schumm (1973) for semiarid drainage systems. Variation in fan complexes in the Salt Spring Hills area are probably the result of drainage area and resultant water and sediment yields, hillslope orientation and slope, drainage network morphology, and fan and active wash gradients, among other factors.

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PATTERNS AND PROCESSES OF MIDDLE AND LATE QUATERNARY DISSECTION IN THE TECOPA BASIN, CALIFORNIA

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ABSTRACT

Breaching and subsequent draining of the Tecopa basin during late middle(?) Pleistocene time resulted in the general dissection of a 70 m section of lacustrine deposits over an area of approximately 240 km^2. Dated tephra deposits within this lacustrine section and relations between the high lake shoreline, erosion surfaces cut across the lake beds, and alluvial surfaces graded to and inset within these erosion surfaces enable the reconstruction of an approximate middle and late Quaternary erosional history: (1) existance of a lake until late middle Pleistocene time; the highest shoreline of this lake attained an average elevation of approximately 545m; (2) breaching of the basin to a temporary outlet elevation of 440 m and draining of the lake between approximately 0.3 to 0.5 m.y. ago; (3) formation of a gently sloping, nearly featureless erosion surface that cut across the deposits of the lake and graded to the 440 m level; (4) deepening of the outlet canyon to an approximate elevation of 410 m followed by renewed dissection of the basin; (5) formation of low straths and local pediments veneered by late Pleistocene and Holocene alluvial gravels during the late stages of this second pulse of dissection. During the two major pulses of erosion, dissection of the Tecopa basin followed a general pattern of relatively deep and complete dissection near the center of the basin accompanied by a conspicuous and systematic decrease in both the depth and completeness of dissection with increasing distance from the basin axis. As much as 50 m of lacustrine deposits have been completely eroded from axial areas; however, less than 5 m of net dissection has occurred beyond the limits of the Pleistocene lake beds. Indeed, several alluvial systems have aggraded slightly in middle and upper piedmont areas in response to late Quaternary climatic changes (both above and below the level of the Pleistocene lake).

Subsurface erosion has been a significant factor in the post-lake dissection of the Tecopa basin. Areas sculpted by piping processes are a common element of the Tecopa landscape, and landforms produced by 'megapiping' or tunneling (including sinkholes up to 30 m wide and 10 m deep, vertical pipes as much as 15 m deep, and blind valleys with disappearing washes up to several 100 m long) dominate the landscape in several areas. The large scale piping that has produced these landforms is the result of a combination of several factors. The lacustrine deposits of the Tecopa basin possess: high silt-clay contents with high percentages of expansive clays, alternating subhorizontal horizons of relative permeability and impermeability, and abundant deep and continuous cracks. In addition, rapid dissection of the lake beds has produced an abundance of free faces which in turn have generated steep hydraulic gradients within the intermittent surface and subsurface drainage of the area. These conditions and the piping landforms which they generate are characteristic of dissected lacustrine deposits in the Mojave Desert.

INTRODUCTION

The Tecopa basin is a small intermontane basin located between Death Valley and Pahrump Valley in the California section of the northern Mojave Desert. From late Pliocene until middle Pleistocene time, the basin was closed and, during most of this time, was partly filled by Lake Tecopa. Because of the relatively large area of the Amargosa drainage upstream from Tecopa (approximately 6000 km^2) and the relatively small size of the lake (approximately 240 km^2), Lake Tecopa served as an efficient trap for late Pliocene and Pleistocene air fall tuffs deposited within the Amargosa drainage. A number of volcanic ash beds have been identified within the lake deposits and provide the basis for reliable time calibration of basin history (Sheppard and Gude, 1968; Sarna-Wojcicki et al., 1984; Hillhouse, in press). During late middle Pleistocene time, the southern margin of the basin was breached, Lake Tecopa was drained, and the Amargosa River became a through-flowing stream to Death Valley. Subsequent dissection of the deposits of Lake Tecopa has exposed an nearly continuous latest Pliocene to middle Pleistocene lacustrine stratigraphy and has preserved a middle to late Quaternary erosional history of basin adjustment to a discontiunously lowered baselevel. In this regard, the Tecopa basin is an exceptional natural archive of Quaternary lacustrine and geomorphic history. Various aspects of the lacustrine stratigraphy have been documented by several workers: distribution and genesis of authigenic silicate minerals (Sheppard and Gude, 1968); clay mineralogy (Starkey and Blackmon, 1979); tephrachronology (Izett et al., 1970; Izett and Wilcox, 1982; Sarma-Wojcicki et al., 1984); paleomagnetic stratigraphy (Hillhouse and Cox, 1976); and general stratigraphy and areal geology (Hillhouse, in press). The post-lake erosional history of the Tecopa basin is the subject of this preliminary report.

Physiography

The Tecopa basin is a roughly triangular intermontane lowland surrounded by varied and irregular uplands. The southern parts of the Nopah and Resting Spring ranges form the eastern margin, the Dublin and Ibex hills flank the basin on the west, and the Sperry and Alexander Hills lie to the south. The basin is small, with an area of approximately 500 km^2, and relatively shallow, with a maximum relief of approximately 1000 m between the basin axis (approximate elevation 405 m) and the highest peaks in the surrounding uplands. Below 545 m (the average elevation of the highest stand of Lake Tecopa) the basin is principally underlain by lacustrine deposits that have been extensively pedimented and deeply dissected by post-lake erosion. In contrast, piedmont areas above the high shoreline are underlain by alluvial fans that are only slightly dissected and locally are actively aggrading.

The present climate of the Tecopa basin is typical of the northern Mojave Desert (Rowlands et al., 1982). Precipitation is low, potential evapotranspiration high, and temperatures extremely variable. Mean annual precipitation is less than 10 cm in lower parts of the basin and averages only 15 cm along the crests of the surrounding uplands (Winograd and Thordarson, 1975). On average, most of this precipitation falls during the winter months. However, precipitation can vary widely from year to year, and an occasional severe storm can produce the equivalent of the mean annual precipitation in less than a single 24-hour period (Rowlands et al, 1982). Temperatures are also extreme and variable. In the northern Mojave, mean minimum January temperatures are commonly less than 0° C, mean maximum July temperatures commonly exceed 35° C, and diurnal temperature variations can exceed 25° C (Rowlands et al., 1982).

Bedrock Geology

The uplands east and west of the Tecopa basin are principally underlain by late Precambrian and Cambrian sedimentary strata and by Tertiary volcanic and sedimentary rocks (Mason, 1948, Chesterman, 1973, Burchfiel et al., 1982). A nearly complete section of uppermost Proterozoic to middle Cambrian marine sedimentary strata (Noonday Dolomite, Johnnie Formation, Stirling Quartzite, Wood Canyon Formation, Zabriskie Quartzite, Carrara Formation, and Bonanza King Formation) is exposed in the prominently banded cliffs of the southern Resting Spring Range and along the eastern flank of the Dublin Hills. These strata are unconformably overlain by Tertiary volcanic rocks, predominantly rhyolitic ash flow tuff, and Tertiary continental clastic rocks. The volcanic rocks and probable correlatives in Greenwater Valley (approximately 40 km to the northwest) have yielded potassium-argon ages ranging from 5.3 to 8.7 m.y. (Fleck, 1970; Hillhouse, in press) indicating that the uplands surrounding the Tecopa basin are geomorphically younger than latest Miocene.

Along the southern margin of the basin, the Sperry Hills and Alexander Hills are composed of Tertiary (?) granite and Precambrian gneiss overlain by the Tertiary China Ranch Beds, a pervasively faulted and folded association of gypsiferous mudstone, heterogenous fanglomerate, and monolithologic megabreccia (Hillhouse, in press). These deposits are extensively exposed along the canyons of the Amargosa River and its tributaries, all of which were cut after breaching of the Sperry Hills and draining of Lake Tecopa. The non-resistant nature of the China Ranch Beds was a major contributing factor to the rapid formation of these deep canyons.

SURFICIAL GEOLOGY

Lacustrine Stratigraphy

The lacustrine stratigraphy of Pleistocene Lake Tecopa has been well documented by Sheppard and Gude (1968) and Hillhouse (in press), and the following stratigraphic summary is based principally on these sources. The deposits of Lake Tecopa underlie an area of approximately 240 km^2 and attain a stratigraphic thickness of at least 70 m in the center of the basin. These lake beds occur as high as 560 m on basin margins and slope gently basinward with dips of generally less than one degree (the combined effect of orginal depositional attitude and post-depositional compaction); however, minor warps, folds, and slumps occur locally. In the central part of the basin, lake beds are predominantly well-bedded mudstone with lesser amounts of claystone and volcanic ash. These deposits are for the most part moderately indurated; however, they disaggragate easily upon wetting and, where not protected by resistant caprocks, commonly degrade into a distinctive lacustrine badlands of low rounded hills and shallow, widely spaced drainageways (Fig. 1). Calcareous strata occur locally and include several thin beds of dolomite. These well-indurated beds form resistant ledges and caprocks that are particularly prominent along the lake's northern margins. Around the margins of the lake, beds and lenses of sandstone and conglomerate interfinger with the finer grained deposits. These coarser grained deposits include deltaic sandstones and pebble conglomerates of the ancestral Amargosa River that flank the present course of this river in the north-central part of the basin.

At least 12 volcanic ash beds, ranging in thickness from one cm to 4 m, occur within the lacustrine deposits (Sarna-Wojcicki, 1984; Hillhouse, in press). These tuffs provide the basis for reliable time control of the Quaternary stratigraphy of the basin. They are composed of vitreous rhyolitic shards with minor crystal and rock fragments. Unaltered ash is white to pale grey and extremely friable. However, most of the ash beds have been altered to zeolites, potassium feldspar, and searlesite. This authigenesis follows a clearly defined areal pattern: unaltered ash along the relatively shallow lake margins, zeolites in a zone of intermediate depth, and searlesite and potassium feldspar in the deep central part of the basin (Sheppard and Gude, 1968). Altered ash is typically pale yellow to pale green in color and porous in texture with a blocky fracture (Hillhouse, in press). Bedding within these tuffs is generally prominent and even; however, channels, ripples and soft sediment deformation structures are common (Fig. 2). Three of the ash beds are sufficiently thick and continuous to be mappable throughout the basin. These three ashes have been comprehensively described by Sheppard and Gude (1968), carefuly mapped by Hillhouse (in press), and reliably correlated with isotopically dated volcanic sources by Izett et al. (1970), Izett and Wilcox (1982), and Sarna-Wojcicki et al. (1984). From youngest to oldest these three ash beds are designated: Tuff A which correlates with member B of the Lava Creek Tuff of the Yellowstone caldera (K-Ar age of 0.62 m.y.); Tuff B which correlates with the Bishop Tuff of the Long Valley caldera (K-Ar age of 0.73 m.y.); and Tuff C which tentatively correlates with the Huckleberry Ridge Tuff of the Yellowstone caldera (K-Ar age of 2.02 m.y.). The salient characteristics of these tuffs and their isotopic ages are summarized in Table 1.

Lake Shorelines

Four types of shoreline indicators are preserved within the Tecopa basin: maximum elevation of the lacustrine deposits, truncated pediment and alluvial fan surfaces, carbonate-cemented beach-rock deposits, and alignments of tufa mounds. Although none of these indicators is by itself sufficient for precise location of a lake shoreline, the systematic association of two or more of these indicators at similar elevations around the perimeter of the basin provides a reasonably conclusive indication of the general extent of the ancient lake. The approximate locations and elevations of these indicators are mapped in Figures 3, 4, and 5. As these figures show, remnants of as many as four probable shorelines are locally preserved within the basin, most notably along the basin's east and southeast flanks. However, the highest shoreline is generally the best developed and most widely preserved. This high shoreline is probably the most relevant to the arguments of this preliminary paper; therefore, evidence of its location is summarized in the following paragraph.

Along the west side of the basin, scattered outcrops of fine-grained lacustrine deposits, mainly mudstone and sandy mudstone, occur as high as 510 to 520 m on the piedmonts of the Dublin, Ibex and Sperry Hills. Along the east side of the basin, similar outcrops occur as high as 540 to 545 m on the west piedmont of the Resting Spring Range and as high as 550 to 560 m on the west piedmont of the Nopah Range. Truncated remnants of the Sperry erosion surface and/or alluvial fans coeval with this surface show a similar eastward rise in elevation. These remnants occur at 510 to 520 m on the east piedmont of the Ibex Hills and along the northwest flank of the Sperry Hills on the west side of the basin (Fig. 3), at 530 to 535 m on the northeast flank of the Sperry Hills (Figs. 3 and 4), at approximately 540 m on the west piedmont of the Resting Spring Range (Fig. 5), and at 550 to 560 m along the west flank of the Nopah Range in the southeast corner of the basin (Fig. 4). Caprock-protected tufa mounds form two discontinuous, subparallel trends at 525 to 530 and at 540 m along the west piedmont of the Resting Spring Range (Figs 5 and 6). The 2- to 4-m-thick caprock on these mounds varies in composition from dense, nearly pure tufa to carbonate-cemented, cross-bedded littoral sands and gravels. The underlying beds, typically highly calcareous, fossiliferous and friable, grade downward into non-calcareous mudstone. As Hillhouse (in press) points out, the lithologic character of these mounds and their association with truncated alluvial fan remnants indicates that they are ancient shoreline features and not the product of fault-controlled spring activity. Narrow shelves of carbonate-cemented, cross-bedded, moderately to well-sorted sands and gravels are preserved on bedrock slopes in at least two locations: at an elevation of about 525 m along the southeast tip of the Dublin Hills approximately 5 km south-southwest of Shoshone (Figs. 3 and 7) and at an elevation of about 520 m on an isolated bedrock knob approximately 3 km southeast of Tecopa and 1.5 km east of the Amargosa canyon (Fig. 4).

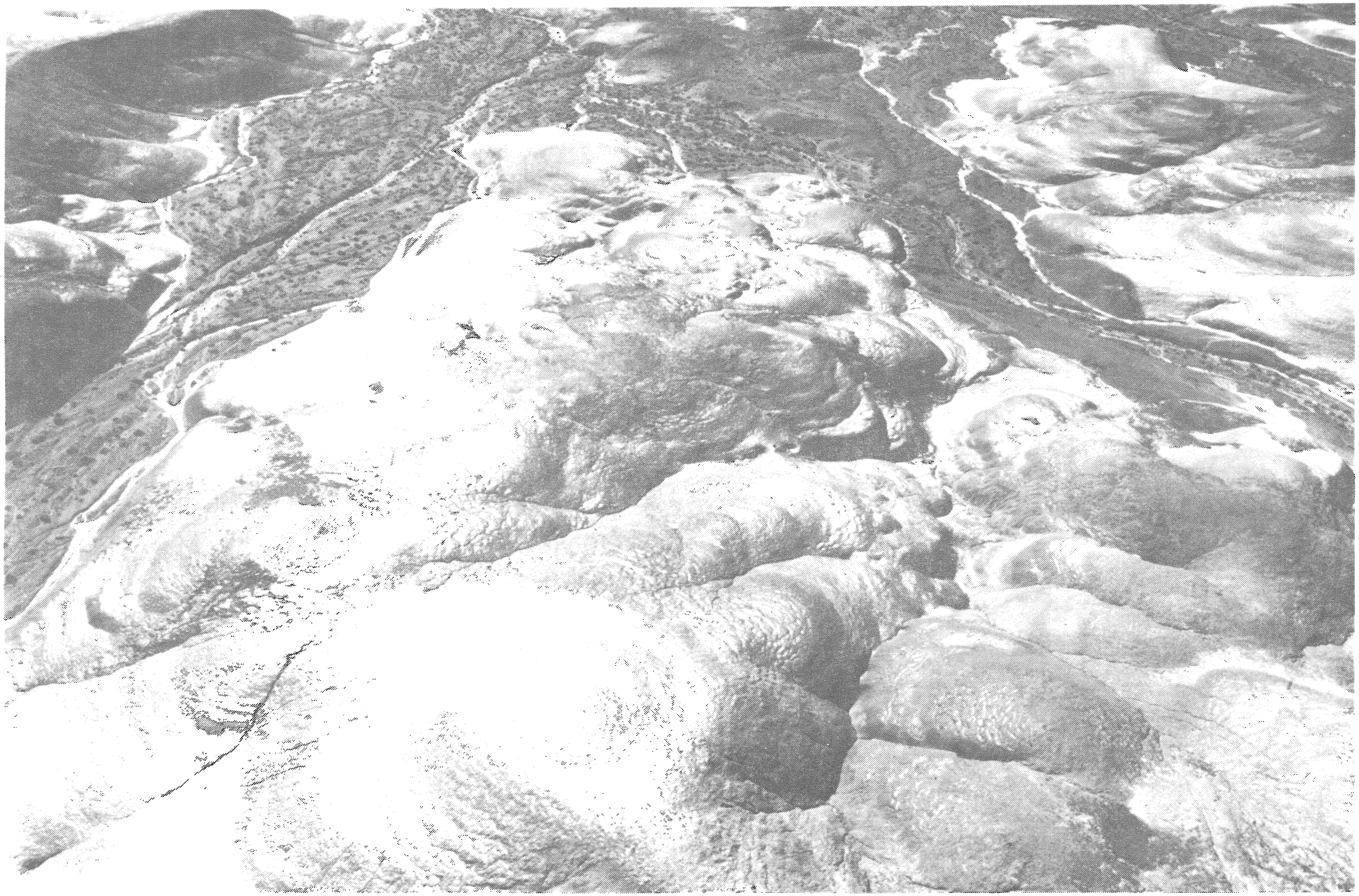

Figure 1. Low-altitude oblique aerial view of dissected mudstone and claystone beds in the central part of the Tecopa basin. Low, rounded hills covered with a 'popcorn' veneer (formed by the weathering of expandable clays; Sheppard and Gude, 1968, Hillhouse, in press) and limited surface drainage are characteristic features of these lacustrine badlands.

Figure 2. Fine bedding and soft sediment deformation structures within the Lava Creek ash (Tuff A). Detailed view of a quarry exposure at the south end of the Dublin Hills.

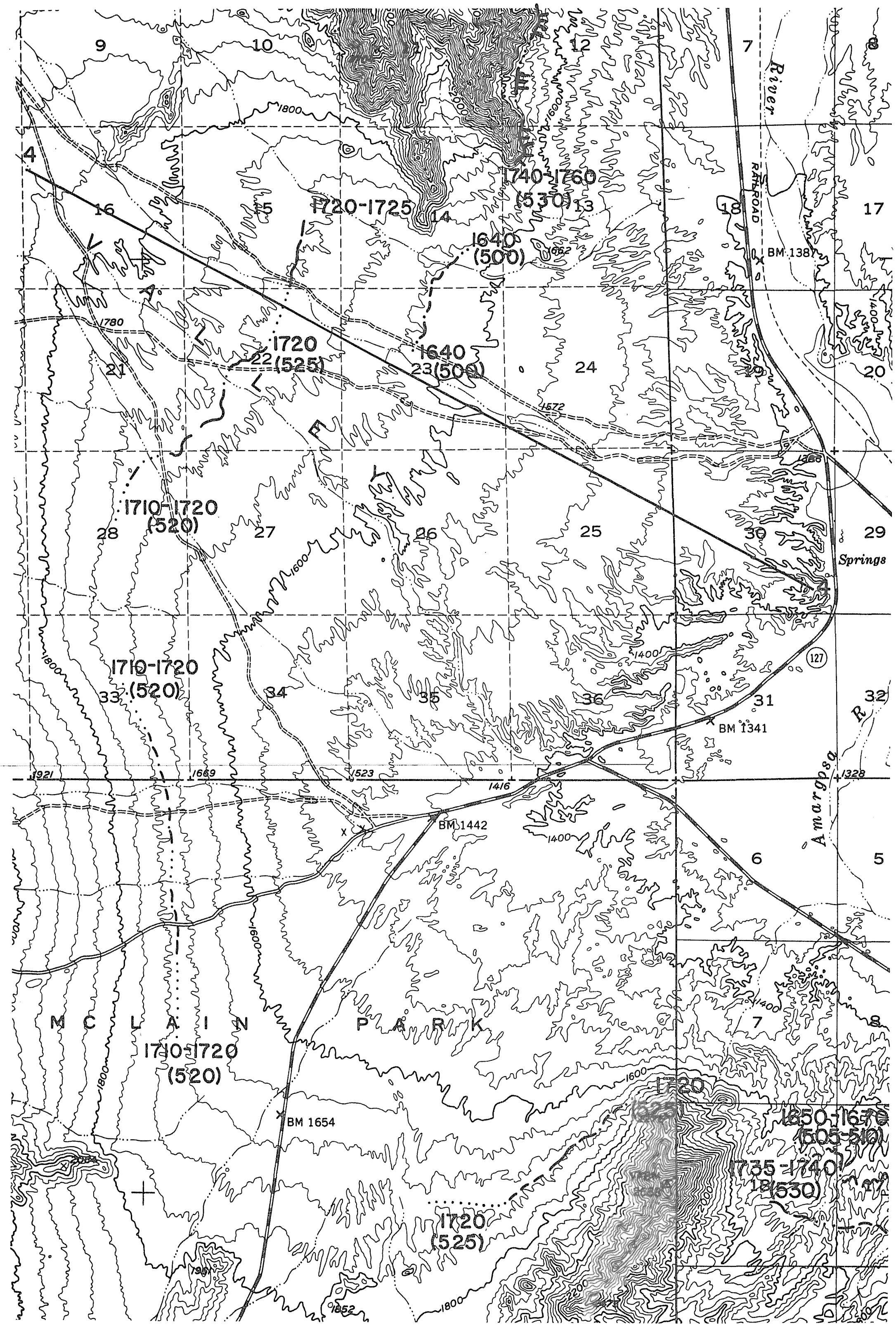

Figure 3. Approximate shoreline locations and elevations in the southwest part the Tecopa basin. Dashed lines indicate truncated alluvial and/or pediment surfaces. Hatchured line along the southeast tip of the Dublin Hills indicates carbonate cemented beachrock benches. Numbered line indicates topographic profile location (see Fig. 9). Contour interval = 40 ft; shoreline elevations in meters.

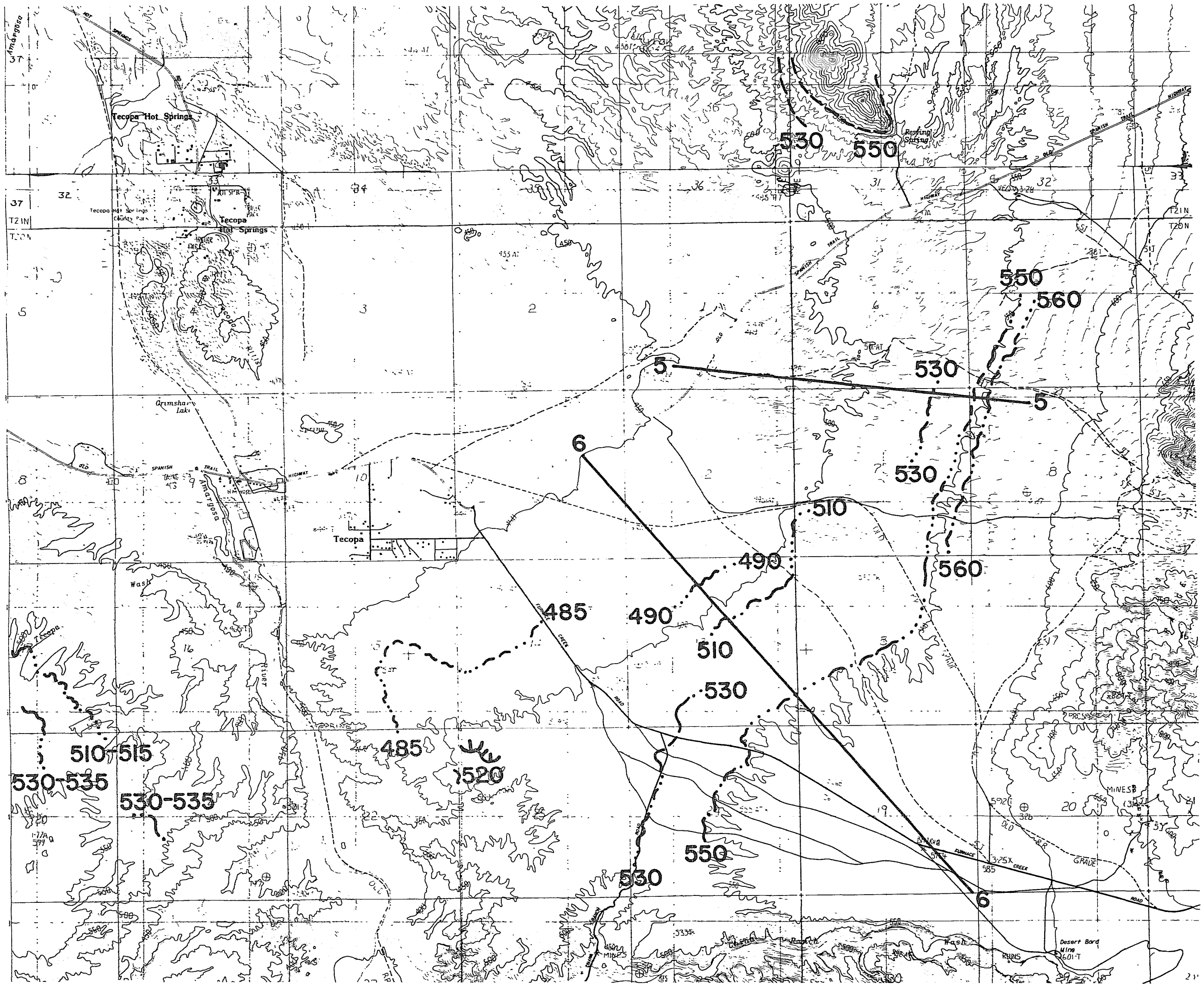

Figure 4. Approximate shoreline locations and elevations in the southeast part of the Tecopa basin. Dashed lines indicate truncated alluvial and/or pediment surfaces. Short hatchured line around an isolated high about 1.5 km east of the Amargosa River canyon indicates a carbonate cemented beachrock bench. Numbered lines indicate topographic profile locations (see Fig 9). Contour interval = 10 m; shoreline elevations in meters.

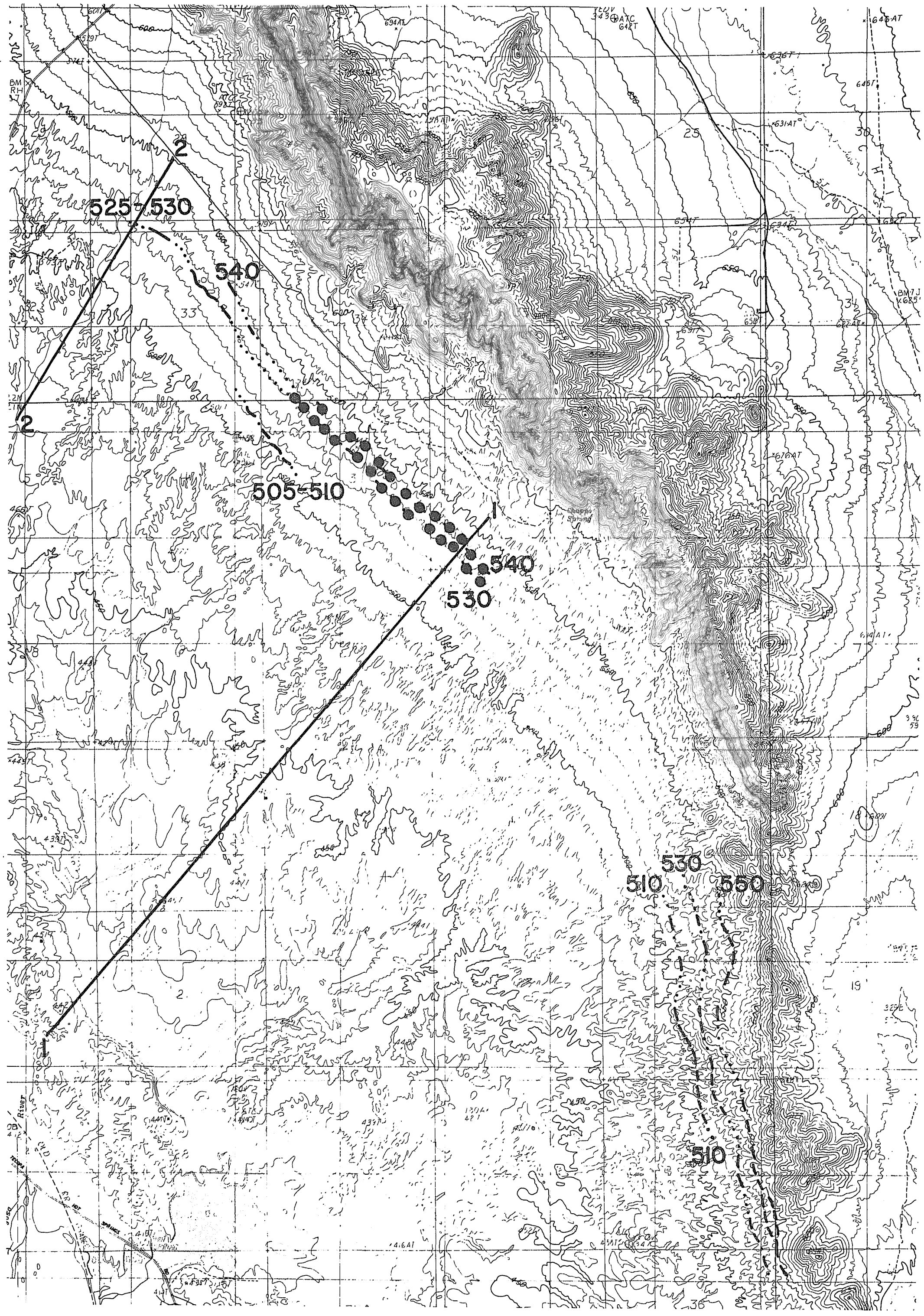

Figure 5. Approximate shoreline locations and elevations in the northeast of the Tecopa basin. Dashed lines indicate truncated alluvial and/or pediment surfaces. Large dots indicate alignments of tufa mounds. Numbered lines indicate topographic profile locations (see Fig. 9). Contour interval = 10 m; shoreline elevations in meters.

Post-Lake Tilting

Three lines of evidence indicate 30 to 40 m of east-to-west tilting across the 19 km width of the lacustrine deposits in the southern part of the basin: (1) systematic increase in the highest elevations of shoreline indicators around the periphery of the basin (from approximately 520 m along the west shore to 550 to 560 m at the southern end of Chicago Valley on the easternmost shore); (2) a general west to east increase in the elevation of structural contours of the lacustrine deposits as drawn on the base of Tuff A (from approximately 475 m along the west side to 505 m along the east side of the lake; Sheppard and Gude, 1968); and (3) a corresponding east to west tilt in the elevation ranges of the diagenetic facies of Tuffs A, B, and C as determined by Sheppard and Gude (1968). In combination, the consistency of these three lines of evidence provides a strong indication that the Tecopa basin has experienced approximately 0.1° of east to west tilt sometime during the past 0.5 m.y. This very gentle tilt is only apparent when the distribution and elevation of the tuff beds and shoreline features are viewed from a synoptic perspective. Therefore, it can be safely concluded that differential tectonic movement has played a very small role in the post-lake dissection of the Tecopa basin.

Alluvial Stratigraphy

The alluvial stratigraphy of the Tecopa basin has been generally described and mapped by Hillhouse (in press), and the pre-Holocene alluvial stratigraphy of the Nopah and Resting Spring piedmonts along the east flank of the basin has been described and mapped in detail by McKittrick (in press). At least three Holocene alluvial surfaces (mapped as Holocene alluvium by Hillhouse and as unit Qf4 by McKittrick) and four pre-Holocene surfaces (mapped as Quaternary older alluvium by Hillhouse and as units Qf3, Qf2, QF1, and QTf by McKittrick) cover the piedmonts of the Tecopa basin above the level of the highest Lake Tecopa shoreline.

The alluvial fan and pediment deposits of the Tecopa basin consist of poorly to very poorly sorted bouldery sand and gravel derived from the carbonate, clastic, metasedimentary, volcanic, plutonic, and metaplutonic terranes that surround the basin. The younger deposits also contain partly rounded fragments of secondary calcite (calcrete rubble and disassociated carbonate rinds) derived from the older deposits. In upper piedmont aresas, younger alluvial deposits are typically inset into older deposits; however in middle and lower piedmont areas, younger deposits commonly overlap older surfaces above the high lake shoreline but are inset into the lake beds below this shoreline (Fig. 8). Thus the several piedmont surfaces of the Tecopa basin generally converge at the approximate elevation of the highest Lake Tecopa shoreline.

The Holocene alluvial surfaces of the Tecopa basin are morphostratigraphically similar to the Holocene surfaces along the west margin of Silver Lake approximately 70 km to the south (described by Wells et al., 1984). These deposits are mainly aggradational and at most only slightly dissected. They occur within or adjacent to (and generally less than one to two meters above) active drainage channels, and they display prominent bar and swale relief of as much as 0.4 m. Stone pavements and rock varnish are absent on the youngest surface. However the development of these features progressively increases on intermediate and oldest Holocene surfaces; and remnants of the older Holocene surfaces generally support substantial areas of moderately to darkly varnished, interlocking stone pavements.

The Pleistocene alluvial surfaces of the Tecopa basin are morphologically distinct from the Holocene surfaces. All of the Pleistocene surfaces are characterized by broad, nearly flat interfluves covered with interlocking stone pavements. Non-carbonate clasts on these surfaces are typically darkly varnished. As is typical in most areas of the Mojave Desert, the oldest Pleistocene piedmont surfaces are preserved as high, well-dissected fan relicts along range fronts, and the

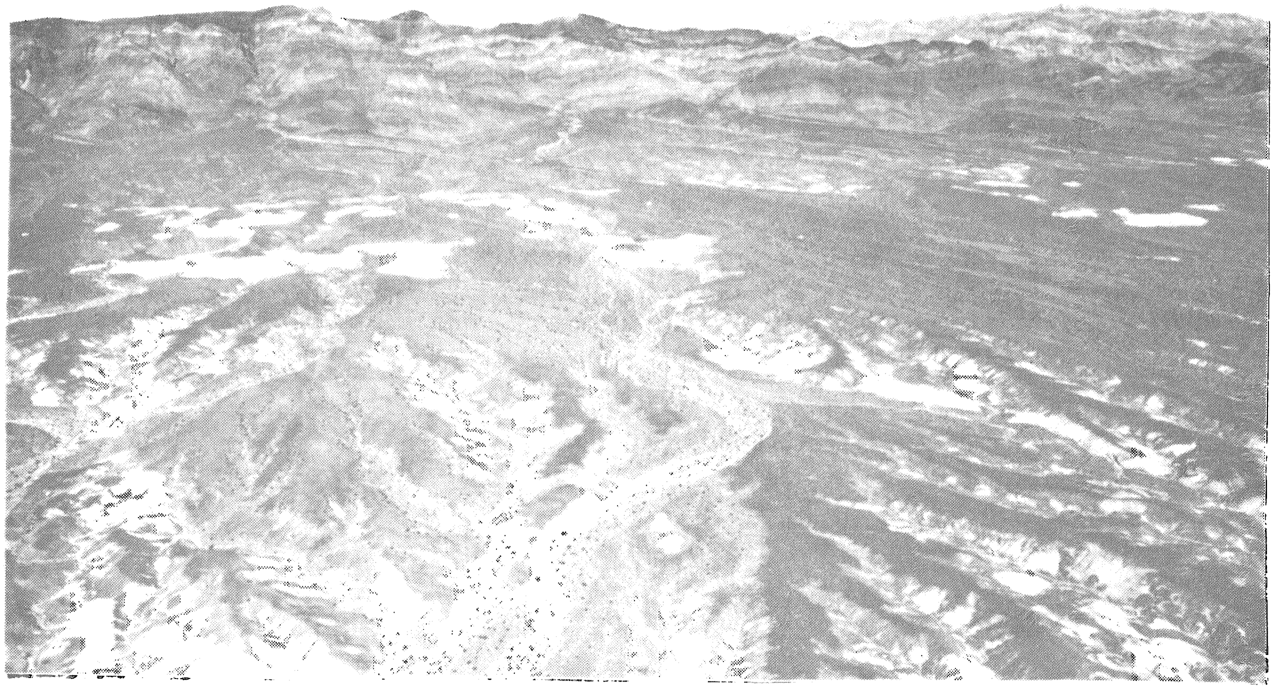

Figure 6. Dissected remnants of the Greenwater erosion surface along the eastern margin of the Tecopa basin. Approximate high shoreline of Lake Tecopa is delineated by the prominent tonal and textural lineament running subparallel to the range front in the mid-distance. Oblique aerial view east towards the Resting Spring Range.

Figure 7. Bench of carbonate cemented beachrock at approximately 525 m elevation on the southeast tip of the Dublin Hills. This deposit, approximately 2 to 10 m wide and 2 to 5 m thick, is composed of moderately-sorted, cross-bedded pebble to cobble gravel. Similar deposits occur discontinuously for about 500 m along the range front.

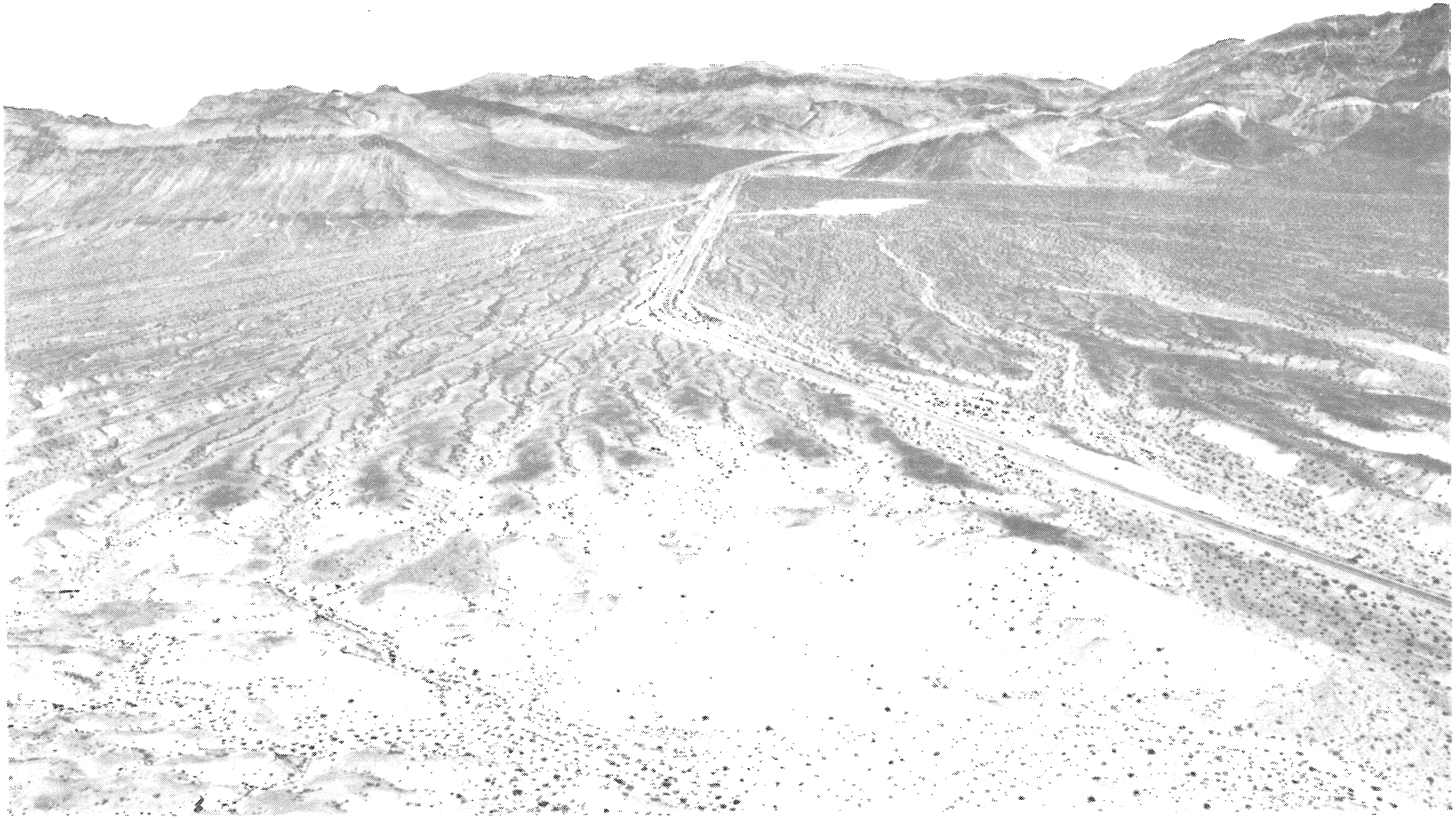

Figure 8. Oblique aerial view of the Greenwater erosion surface and overlapping alluvial deposits in the vicinity of the high lake shoreline (the prominent tonal and textural discontinuity that runs straight across the lower part of the photograph). Above the high lake shoreline, Holocene and late Pleistocene alluvial deposits overlap the pediment surface; below the high shoreline, Holocene and late Pleistocene erosional terraces are inset as much as 10 m into the pediment surface (lower right corner of the photograph). View northeast along the Charles Brown highway (California 178) towards the Resting Spring Range.

younger surfaces typically overlap these oldest surfaces in more basinward areas. Interfluve rounding, piedmont-sourced drainage development, piedmont-sourced drainage dissection, and incorporation of secondary carbonate fragments in surface pavements generally increase with surface age on these Pleistocene alluvial surfaces.

The oldest subaerially exposed alluvial surface preserved in the Tecopa basin provides an instructive morphologic contrast to the younger surfaces of the area. This surface is best preserved on the southeast flank of the Ibex Hills west of Ibex Pass and on the west piedmont of the Nopah Range. Wide v-shaped valleys, from 10 to as much as 30 m deep, separate high, well rounded interfluves. Well-indurated, matrix-supported calcrete occurs on or just below interfluve surfaces, and calcrete rubble locally makes up the majority of surface clasts. The extreme morphological and pedological differences between this surface and the Pleistocene surfaces that either grade to or are truncated by the high Lake Tecopa shoreline suggest considerable antiquity, possibly a pre-Quaternary age, for this oldest surface.

Pediment Surfaces

Three general pediment surfaces cut across the Tecopa lake beds, and the highest of these surfaces also bevels the China Ranch Beds along the southern margin of the basin. These three erosion surfaces are informally designated (from youngest to oldest) the Amargosa surface, the Greenwater surface, and the Sperry surface. Representative longitudinal profiles of these surfaces are presented in Figure 9.

The Amargosa surface is defined as the modern erosion surface of the Tecopa basin. This surface includes the broad fluvial plain of the Amargosa River in the central part of the basin and the multitude of washes and small canyons that incise the Greenwater surface below the high lake shoreline. Thus, this surface is the most complex of the three general erosion surfaces in both aeral extent and surface character. However, the primary significance of this surface lies in its comparison with the two older erosion surfaces; other aspects of the surface are not particularly important to the present discussion.

The Greenwater surface is the most extensive and prominent pediment surface in the Tecopa basin. Broad, nearly continuous remnants of this surface bevel lacustrine deposits on the piedmonts of the Resting Spring Range, Dublin Hills and Sperry Hills (Figs. 6 and 8); however, this surface is best preserved immediately south of the Dublin Hills as a broad, gently sloping, fan shaped area of approximately 25 km^2. Basinward margins of this surface are dissected by steep-sided arroyos and small box canyons as much as 20 m deep. Capping gravels vary in thickness from several meters in the vicinity of the high lake shoreline to as little as a single layer of gravel clasts on degraded surface remnants near the basin axis. Surfaces are generally flat and are almost everywhere covered by darkly varnished, interlocking stone pavements.

Dissected remnants of the Sperry surface grade to the approximate level of the high Lake Tecopa shoreline but are truncated by that shoreline. Isolated remnants of the Sperry surface and of coeval alluvial fans are preserved on the upper to middle parts of all the piedmonts flanking the Tecopa basin. However, the surface is most extensively preserved on the northeast flank of the Sperry Hills where pediment gravels, as much as 10 m thick, cap the eroded surface of the China Ranch Beds. This gravel cap can be distinguished from the underlying conglomerate of the China Ranch Beds by its lack of deformation, consistent basinward slope, generally smaller clast sizes, and absence of large boulders (Hillhouse, in press). Although dissected by steep-sided canyons as much as 40 m deep, remnants of this surface are generally only slightly rounded and are typically covered by darkly varnished, interlocking stone pavements.

In the general vicinity of the high lake shoreline, relations between the Sperry surface, the Greenwater surface and the Pleistocene alluvial surfaces are locally complex. All of these surfaces are morphologically similar, and at the approximate level of the high lake shoreline, they all lie

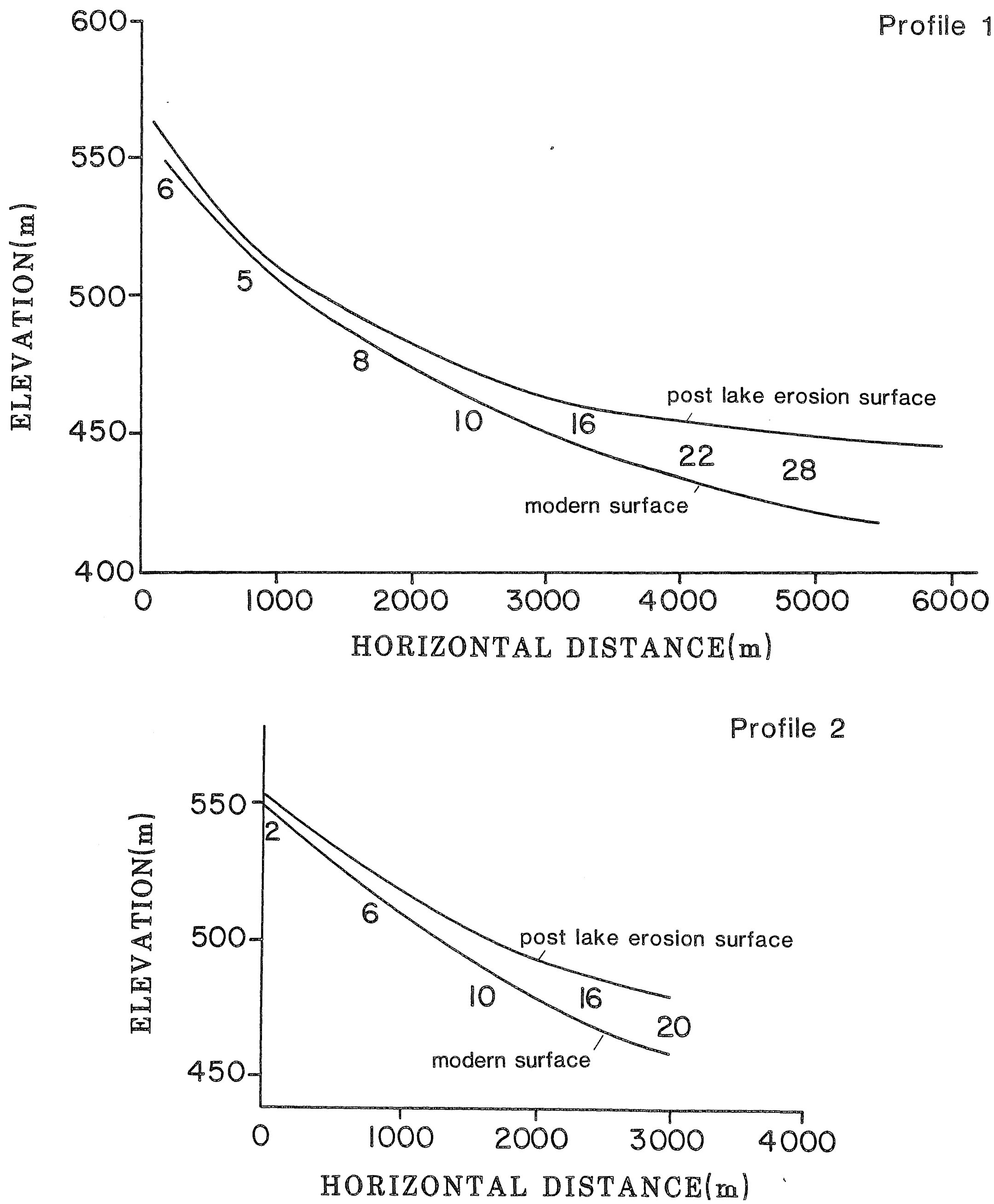

Figure 9. Representative longitudinal profiles along remnants of the major erosion surfaces of the Tecopa basin. Vertical exaggeration: 20X. The modern surface is equivalent to the Amargosa surface and the post lake erosion surface is equivalent to the Greenwater surface. Numbers along the profiles indicate height differences (in meters) between the surfaces. Profile locations are given in Figure 5.

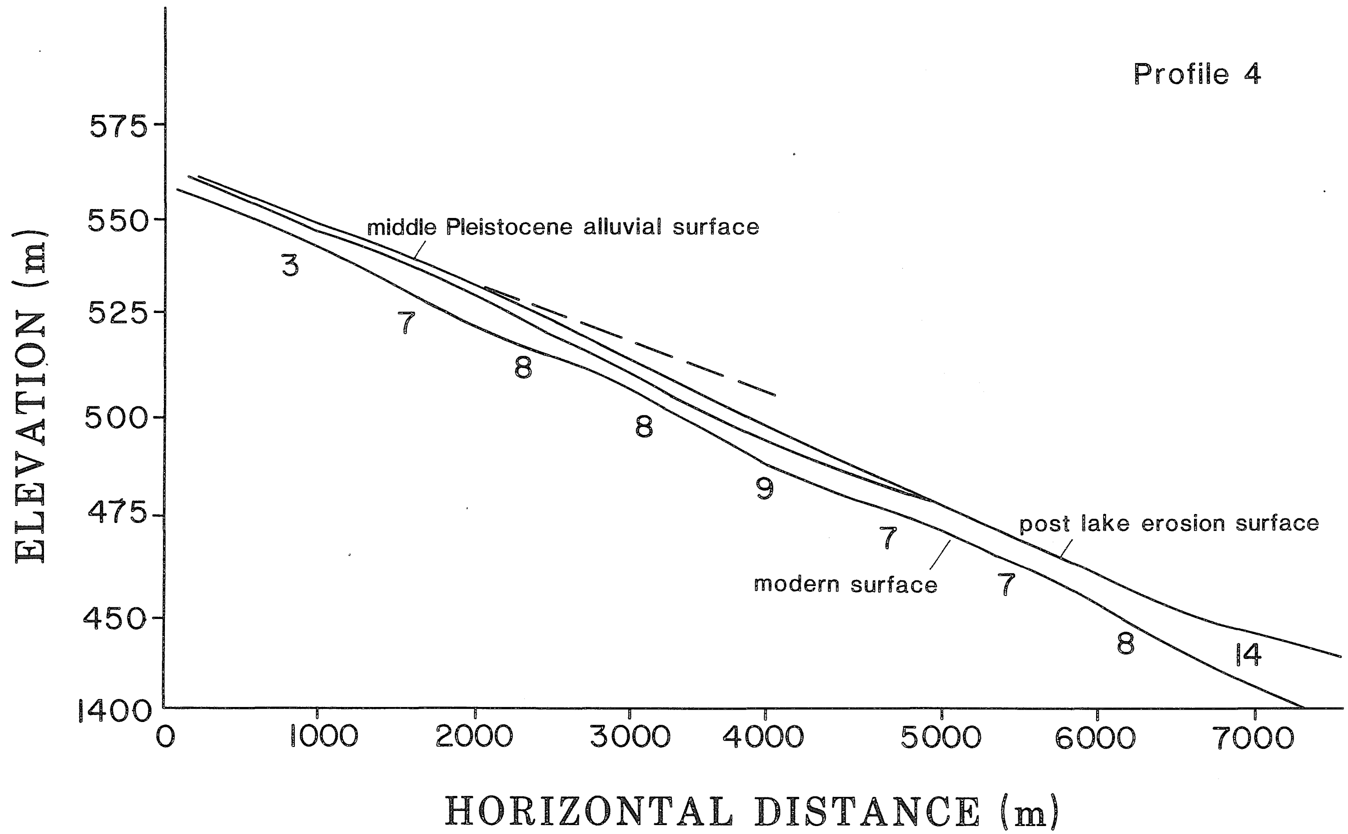

Figure 9 (con't). Representative longitudinal profile along remnants of the major erosion surfaces of the Tecopa basin. Vertical exaggeration: 20X. The modern surface is equivalent to the Amargosa surface, the post lake erosion surface is equivalent to the Greenwater surface, and the middle Pleistocene alluvial surface is equivalent to the Sperry surface. Numbers along the profile indicate height differences (in meters)between the surfaces. Profile location is given in Figure 3.

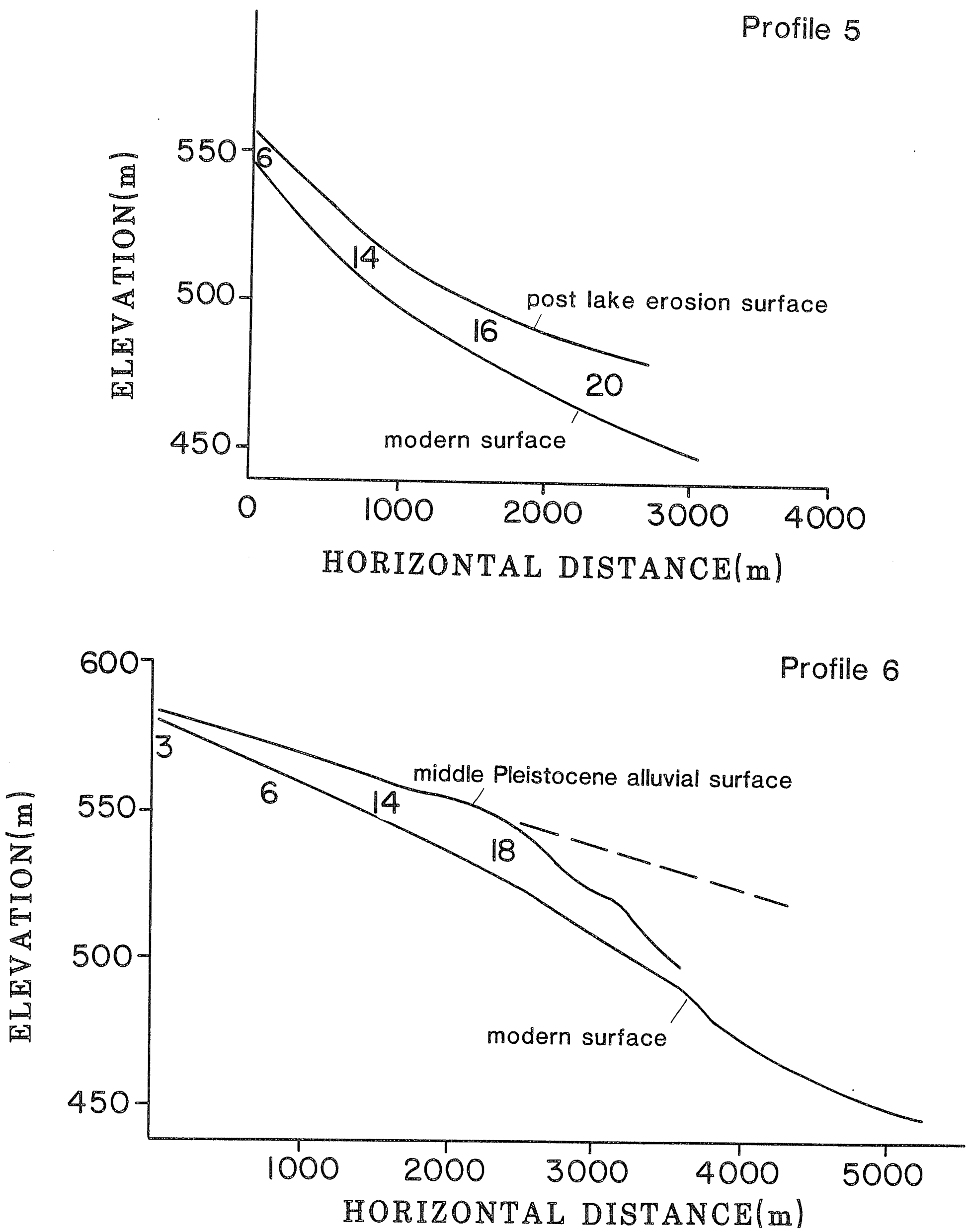

Figure 9 (con't). Representative longitudinal profiles along remnants of the major erosion surfaces of the Tecopa basin. Vertical exaggeration: 20X. The modern surface is equivalent to the Amargosa surface, the post lake erosion surface is equivalent to the Greenwater surface, and the middle Pleistocene alluvial surface is equivalent to the Sperry surface. Numbers along the profiles indicate height differences between the surfaces. Profile locations are given in Figure 4.

within a few meters of the modern washes. These surfaces are commonly overlapped in an up-piedmont direction and incised in a down-piedmont direction by younger alluvial deposits of the same age (Fig. 8). Because the soils of the Tecopa basin have not been systematically described and mapped, these several surfaces are best differentiated in the area of the high lake shoreline by tracing divergent surfaces from areas either higher or lower on the piedmont.

Approximate Ages of the Pleistocene Erosion Surfaces

Several lines of reasoning can be employed to estimate the approximate ages of the Pleistocene erosion surfaces in the Tecopa basin. Tuff A (0.62 m.y., Table 1) lies 4 to 8 m stratigraphically above Tuff B (0.73 m.y., Table 1); thus the average rate of deposition bewteen these two tuff beds was approximately 0.062 m/10^3yr. Extrapolating this rate to the aproximately 20 m of sandy mudstone and silty conglomerate that stratigraphically overlies Tuff A provides the basis of a probable minimum age (approximately 0.3 m.y.) for the youngest lacustrine deposits and for breaching of the basin and draining of the lake. Recognizing that the marginal lake facies deposits overlying Tuff A were probably deposited at a higher average rate than that of the mudstone between Tuffs A and B, a conservative age range for the youngest lacustrine deposits would be 0.3 to 0.5 m.y. The Sperry surface, truncated by the high lake shoreline, must be somewhat older than this approximate age and the Greenwater surface, cut across the lake beds, must be somewhat younger. Additional constraints on the age of the Greenwater surface are provided by the late Pleistocene alluvial fills and straths that are inset 10 to 20 m below the level of the Greenwater surface (along its deeply dissected basinward margin) but stand only 3 to 4 m above the modern washes. This relation suggests that the Greenwater surface is significantly older than latest Pleistocene. An uranium-trend age of 0.16 ± 0.02 m.y. from a soil sample collected in a gravel pit on the Greenwater surface approximately 2.5 km east of Shoshone (Hillhouse, in press) is consistent with these general age constraints and indicates a likely age of approximately 0.15 to 0.25 m.y. for the Greenwater surface. Although these imprecise relations can only provide general age estimates for the dominent geomorphic features of the Tecopa basin, it can still be concluded that: (1)the most likely age of the Sperry surface is somewhat older than 0.5 m.y., (2)the most likely age for the youngest deposits of Lake Tecopa lies between 0.3 to 0.5 m.y.; and (3) the most likely age for the Greenwater surface lies between 0.15 and 0.25 m.y.

BASIN DISSECTION

Temporal Patterns of Dissection

The presence of several widespread and approximately dated geomorphic surfaces permits reconstruction of a preliminary middle to late Quaternary erosional history for the Tecopa basin. Comparison of the relative positions of the original lake bed surface, the post-lake erosion surface (Greenwater surface), and the modern surface (Amargosa surface) provides a general picture of post lake dissection (Figs. 10 and 11). Two general erosional pulses, separated by a period of erosional stability, can be inferred: (1) rapid vertical dissection of the original lake bed surface to a baselevel of approximately 440 m (the level of the Greenwater surface); (2) formation of the Greenwater surface; and (3) rapid vertical dissection of the Greenwater surface to the present baselevel of 410 m. Initial dissection to the 440 m baselevel and formation of the Greenwater surface probably took place between 0.5 and 0.25 m.y.B.P. The actual duration of these two events may have been as long as 0.3 m.y. or as short as 0.1 m.y. During this time, formation of the Greenwater surface required the complete removal of 20 to 30 m of lacustrine deposits from the central part of the basin (Fig. 10D) at an average erosion rate of somewhere between 8 and 25 cm/10^3 yr. Assuming that erosion rates would have slowed substantially as the final broad and

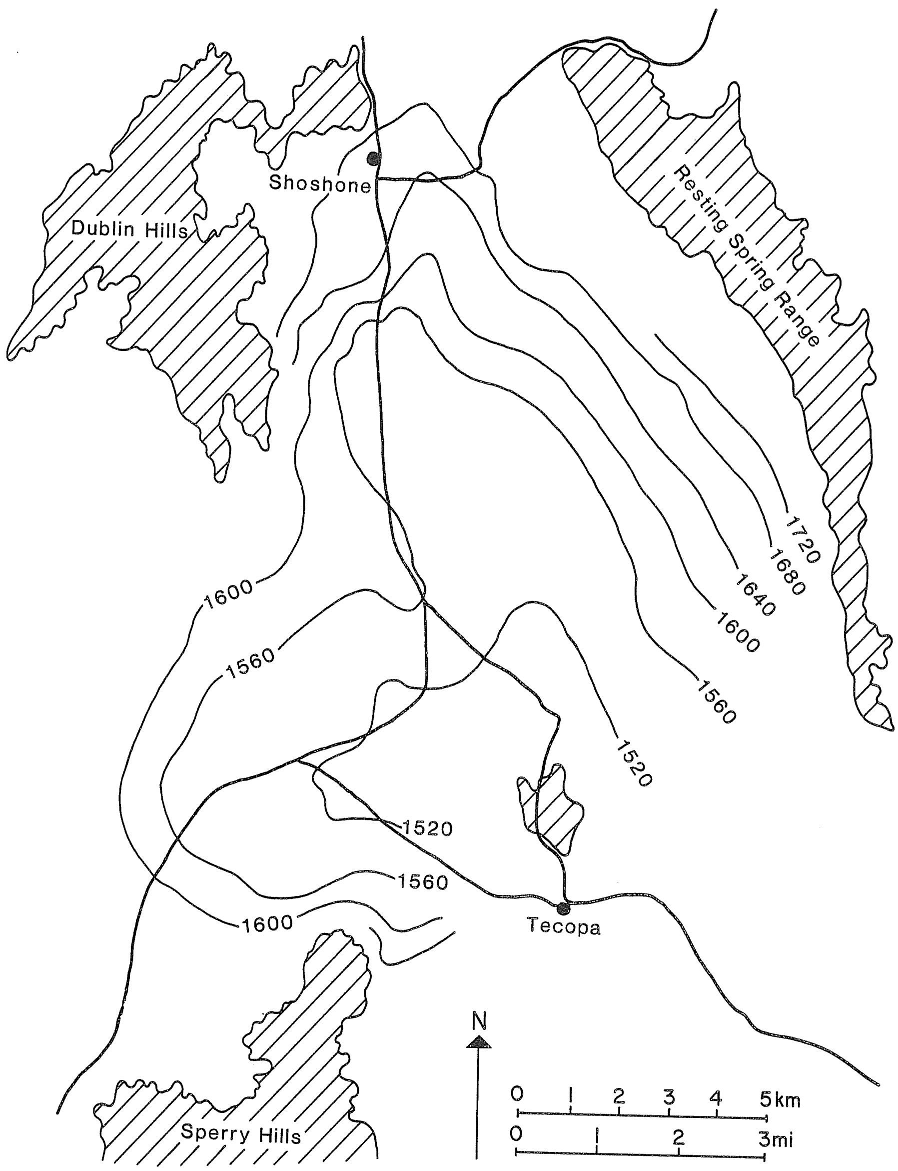

Figure 10A. Post-lake dissection of the Tecopa basin. Generalized contour map of the approximate original surface of the lacustrine deposits. Contour interval = 40 ft.

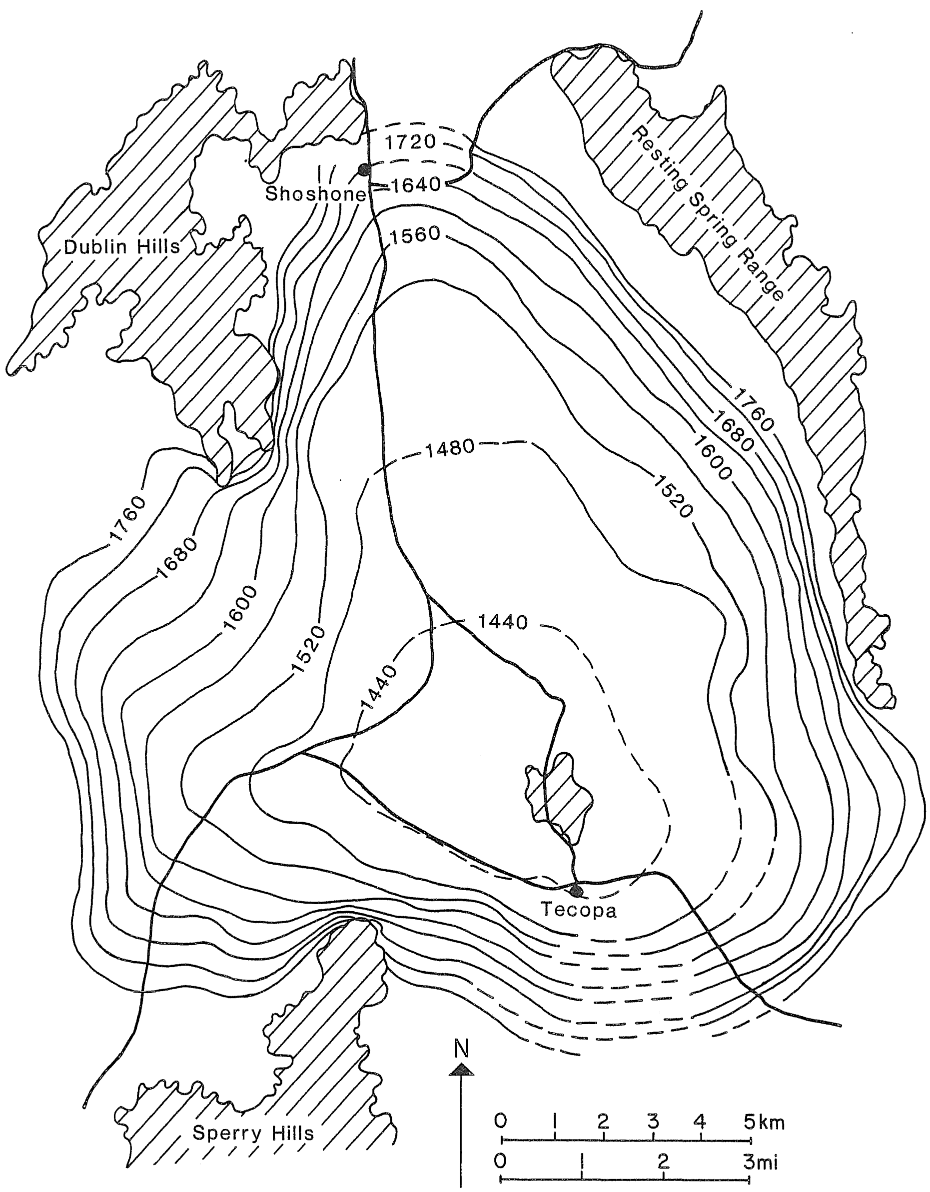

Figure 10B. Post-lake dissection of the Tecopa basin. Generalized contour map of the Greenwater erosion surface. Contour interval = 40 ft.

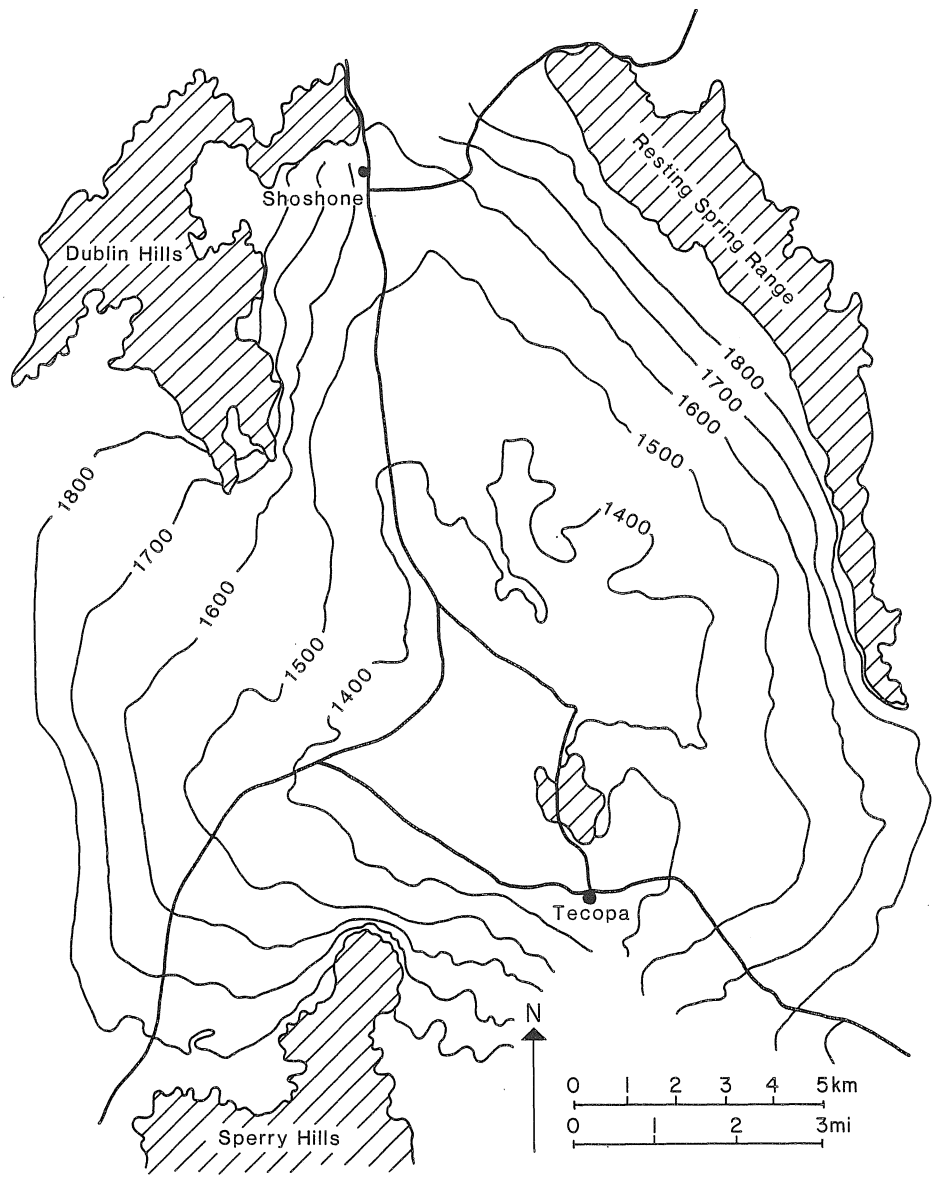

Figure 10C. Post-lake dissection of the Tecopa basin. Generalized contour map of the Amargosa erosion surface. Contour interval = 100 ft.

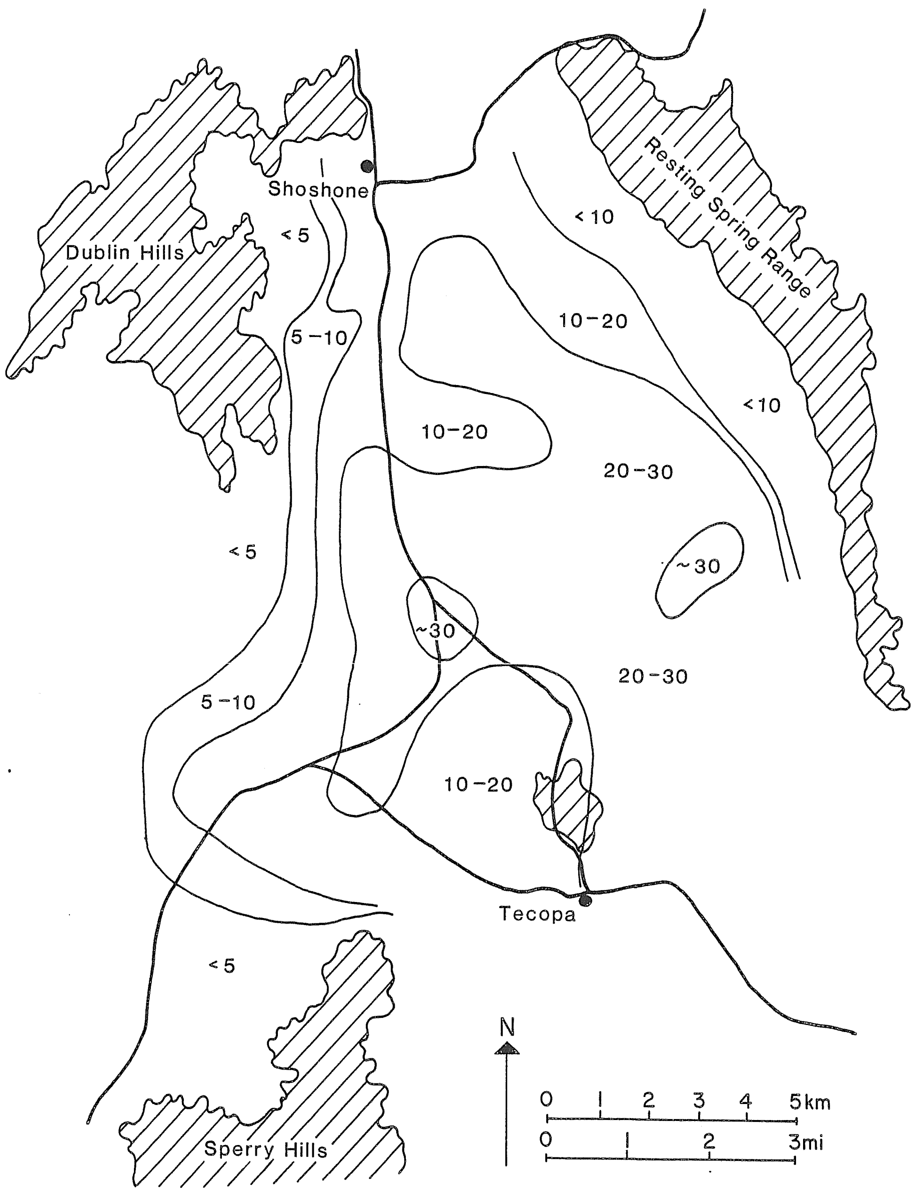

Figure 10D. Post-lake dissection of the Tecopa basin. Generalized contour map showing the approximate amount of vertical dissection between draining of the lake and formation of the Greenwater surface. Units in meters.

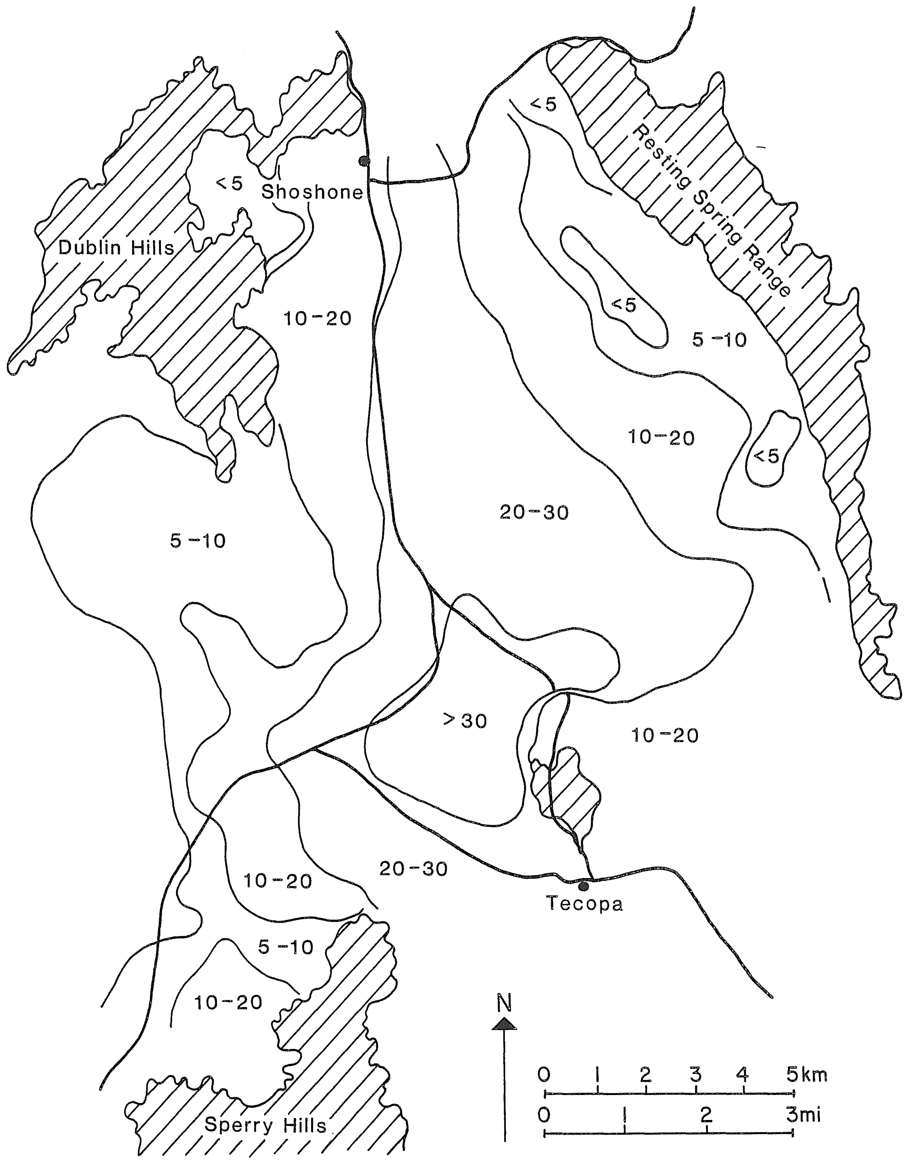

Figure 10E. Post-lake dissection of the Tecopa basin. Generalized contour map showing the approximate amount of vertical dissection of the Greenwater erosion surface. Units in meters.

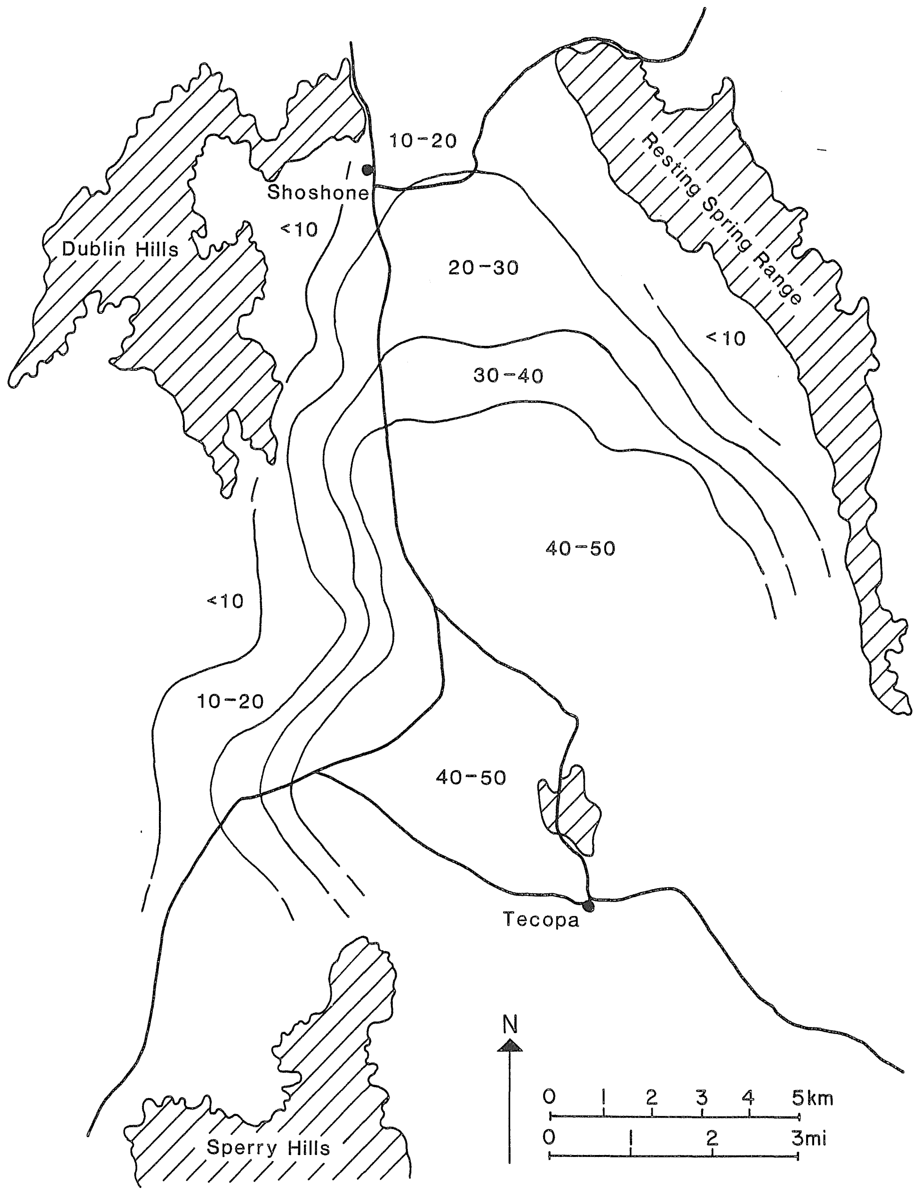

Figure 10 F. Post-lake dissection of the Tecopa basin. Generalized contour map showing the approximate amount of total post-lake dissection. Units in meters.

nearly featureless form of the Greenwater surface was approached, erosion rates during the early stages of lake bed dissection must have been considerably higher. Following the formation of the Greenwater surface and the inferred period of relative erosional stability which it represents, renewed erosion to the present 410 m baselevel commenced. The onset of this second erosional pulse probably took place after 0.25 m.y.B.P. and perhaps as late as 0.15 m.y.B.P. Formation of the Amargosa surface required the local removal of as much as 30 m and the general removal of 10 to 20 m of lacustrine deposits from the center of the basin (Fig. 10E) at an average rate of somewhere between 6 and 10 cm/10^3yr. Given the irregular configuration of the modern Amargosa surface, it can be inferred that this second pulse of basin dissection is probably continuing. However, the presence of numerous Holocene and probable late Pleistocene straths and alluvial fills within a few meters of the modern surface indicates that this dissection is proceeding slowly and somewhat discontinuously.

Spatial Patterns of Dissection

During the two major post-lake erosional pulses, dissection of the Tecopa basin has followed a general pattern of relatively deep and complete dissection in the central part of the basin accompanied by a dramatic and systematic decrease in both the depth and completeness of dissection with increasing distance from the basin axis (Figs. 10 and 11). As much as 50 m of lacustrine deposits have been completely eroded from axial areas; however, less that 5 m of net dissection has occurred beyond the limits of the lake beds (Figs. 10F and 11). Significant dissection ends near the level of the high lake shoreline; and in middle piedmont areas above the high shoreline, many alluvial systems have aggraded slightly, probably in response to late Quaternary climatic change (Fig. 8). Therefore, it would appear that the alluvial systems in most upper and middle piedmont areas of the basin have not been significantly affected by draining of the lake and dissection of the lacustrine deposits. The alluvial systems of the Tecopa basin are largely controlled by their own geometry and sediment delivery. Sediment yields from the ranges surrounding the basin have obviously been sufficient to induce mid-piedmont deposition in the face of significant baselevel lowering in axial areas on the basin. Moreover, although dissection of the lacustrine deposits has produced 50 m of baselevel lowering accompanied by the formation of spectacular erosional scarps and badlands in the central part of the basin, overall piedmont geometry has been little affected. The 0.7 to 1.9° slopes of the modern surface in lower piedmont areas, although significantly steeper than the older geomorphic surfaces (Fig. 11), remain significantly less than the 2.5 to 4° middle piedmont slopes of the active alluvial systems.

Dissection Processes

Post-lake dissection of the lacustrine deposits of the Tecopa basin has been accomplished by a combination of concentrated subsurface flow and overland flow processes. Undoubtedly, many degradational aspects of the Tecopa landscape have been produced by concentrated overland flow. The Greenwater erosion surface was formed primarily by a combination of vertical stream incision and lateral stream corrasion. Moreover, dissection of this pediment has been strongly influenced by fluvial action. A well integrated subparallel to subdendritic network of closely-spaced, steep-sided washes radiates outward from the central part of the basin; and in areas of resistant caprock, these washes have carved small canyons up to 20 m deep into the Greenwater surface. However, indications of long-continuing concentrated subsurface flow are also widely and abundantly distributed across large areas of the basin. Irregular to crudely polygonal patterned ground, indicative of general gradual surface collapse, cover many degraded remnants of the Greenwater surface in the east-central part of the basin. Small sinkholes, 1 to 3 m wide and deep, and small pipes, as large as 0.4 m in diameter, are common along the steep-sided margins of many of the

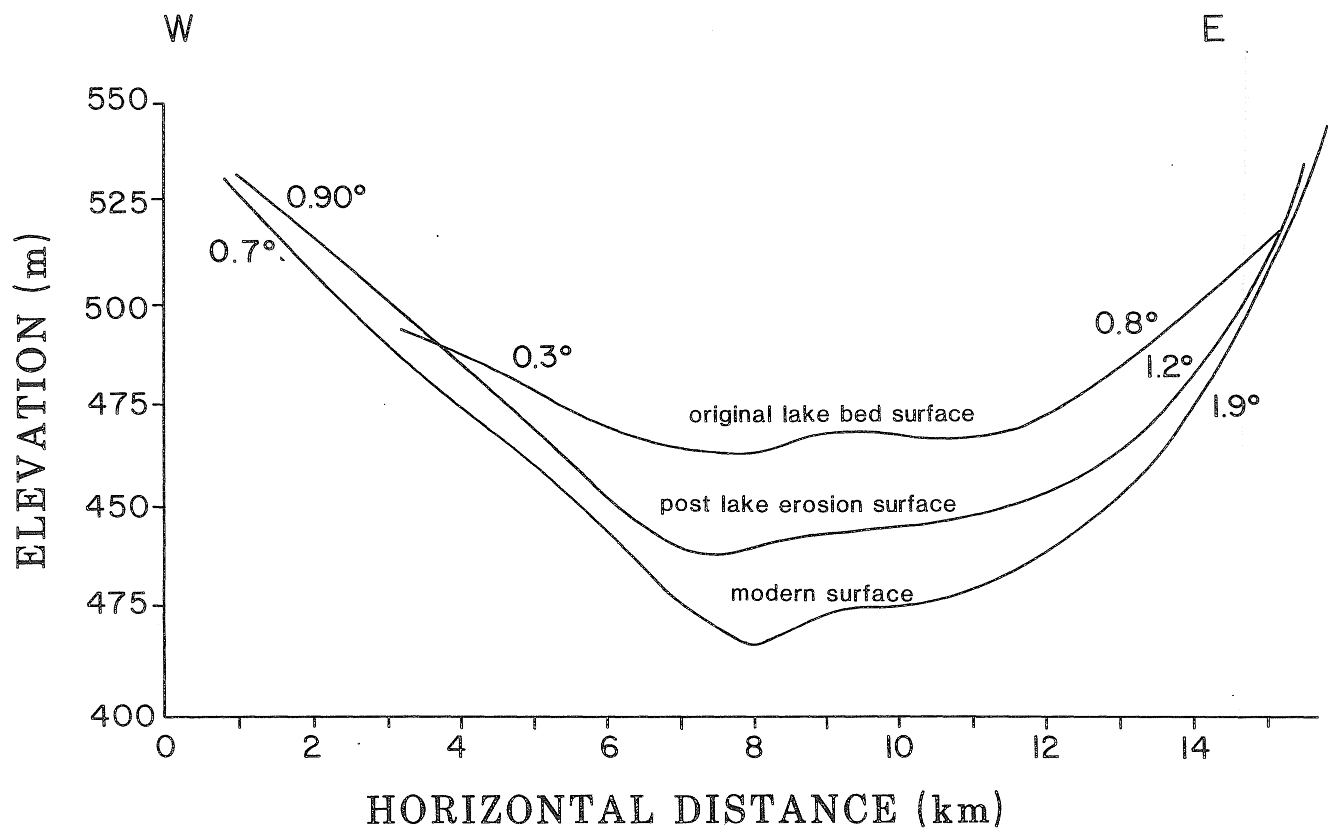

Figure 11. Generalized east-west cross-section across the Tecopa basin comparing the approximate positions of the original surface of the lake beds, the post-lake erosion surface (Greenwater surface), and the modern surface (Amargosa surface).

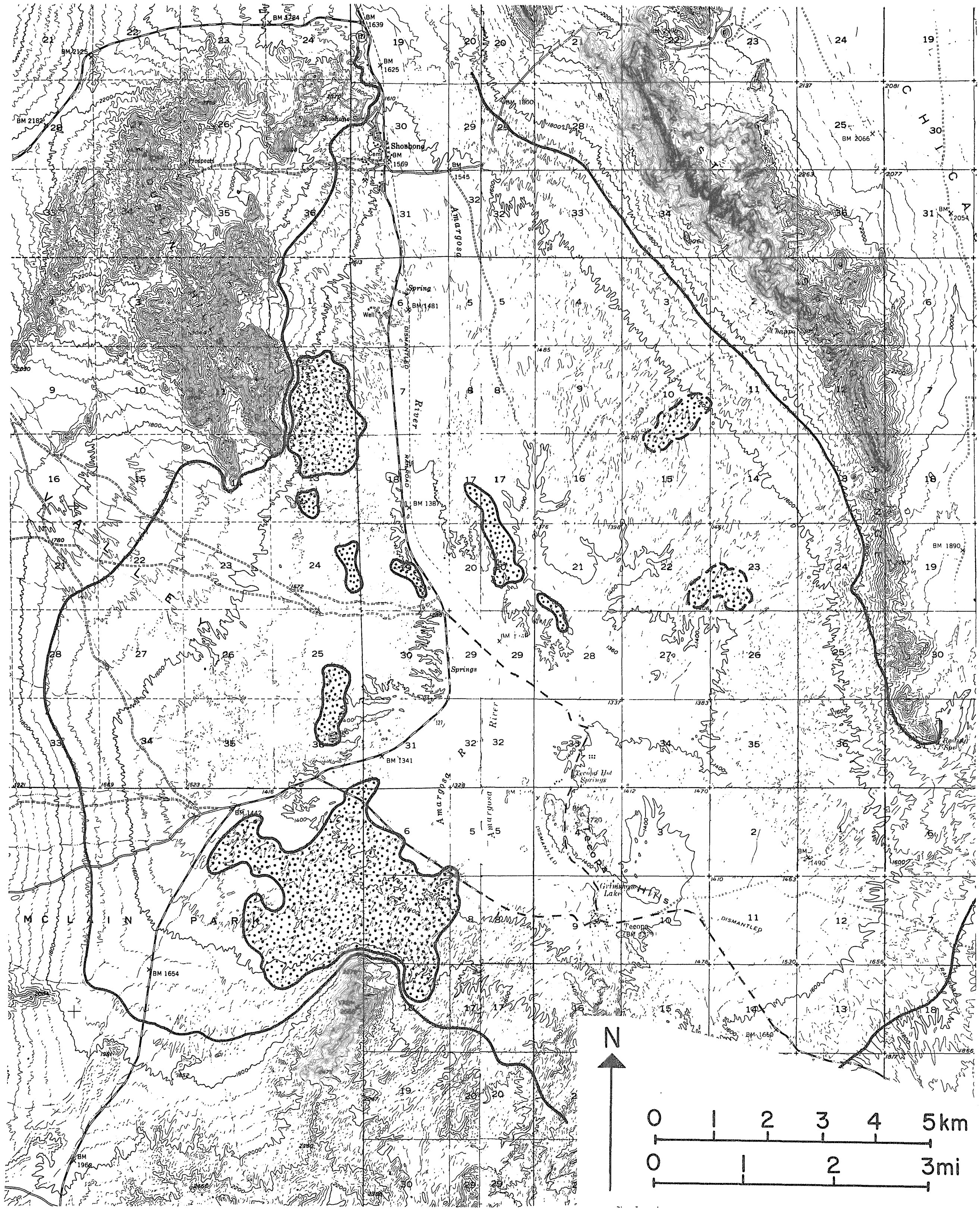

Figure 12. Map showing the locations of areas dominated by megapiping in the Tecopa basin.

larger washes. On a larger scale, sinkholes (as much as 30 m wide and 10 m deep), vertical pipes (1 to 2 m in diameter and at least 15 m deep), and blind valleys (up to 10 m deep and several 100 m long) dominate the landscape of large areas in the west-central and southwest parts of the basin (Figs. 12, 13, 14, and 15). In addition, 'megapiping' tunnels as much as 100 m long with rooms up to 8 m wide and 5 m high have been located in the west-central part of the basin near the south end of the Dublin Hills (Rogers, 1981). In several areas, these large piping landforms have coalesced into chaotic landscapes of irregular low ridges and hills without integrated surface drainage.

Although the exact geomorphic and hydrologic role of pipeflow in arid lands is still relatively unknown, this process has been demonstrated to play a significant if not dominant role in the initiation and continued erosion of some arid badlands (Bryan and Yair, 1982; Harvey, 1982), and large-scale piping landforms have been reported in dissected basin deposits from several areas of the Mojave Desert (Hunt et al., 1966; Rogers, 1981; Rogers et al., 1983). Piping can develop in many different types of materials and under a variety of geologic conditions, "the only essential requirement being juxtaposition of a steep hydraulic gradient and a steep free face" (Bryan and Yair, 1982, p.8). However, large scale piping appears to develop most readily where high infiltration capacities occur in subhorizontally bedded, poorly indurated materials of low intrinsic permeability. Moreover, the largest pipes or tunnels are typically associated with strongly indurated layers that provide protection against roof collapse (Bryan and Yair, 1982).

In the specific case of the Tecopa basin, several geologic, geomorphic, and climatic factors combine to produce ideal conditions for large-scale piping processes. (1) The subhorizontally bedded, poorly to moderately indurated, lacustrine deposits of the basin generally possess low intrinsic permeabilities. (2) High infiltration capacities within these deposits are produced by extensive cracking. (3) High silt-clay contents with high percentages of expandable clays provide the necessary shrink-swell potential for this cracking. Throughout the basin, the lacustrine deposits include significant amounts of illite and montmorillonite (Starkey and Blackmon, 1979), and within the area of large scale tunneling in the west-central part of the basin, the silt-clay size fraction of these deposits contains about 60% illite and 9% smectite (Rogers, 1981). (4) High potential evapotranspiration coupled with low mean annual rainfall and the occasional occurance of high intensity, short duration storms provides the appropriate climatic conditions for shrink-swell cracking and for the generation of steep, albeit short-lived, hydraulic gradients. (5) Deep dissection of the lacustrine deposits provides abundant free faces and a high geomorphic potential for steep hydraulic gradients adjacent to these free faces. (6) Strongly indurated layers of carbonate-cemented mudstone or marl occur within extensively piped areas and typically form the roofs of the larger pipes. Where these factors all come together, concentrated subsurface flow processes locally outstrip overland flow processes and large-scale piping landforms dominate the landscape.

CONCLUSIONS

(1) Four types of shoreline indicators are preserved within the Tecopa basin: (a) maximum elevation of lacustrine deposits; (b) truncated pediment and alluvial fan surfaces; (c) carbonate-cemented beach-rock deposits; and (d) alignments of tufa mounds. The systematic association of these indicators at similar elevations around the periphery of the basin provides a reasonably conclusive indication of the general extent of the ancient lake and indicates a 0.09° east-to-west, post-lake tilt across the southern part of the basin. The elevation of the highest shoreline increases from approximately 520 m along the west shore to as high as 560 m along the southeast shore.

(2) Three general erosion surfaces occur within the Tecopa basin: (a) the Amargosa surface which includes the modern fluvial plain of the Amargosa River and the numerous lesser

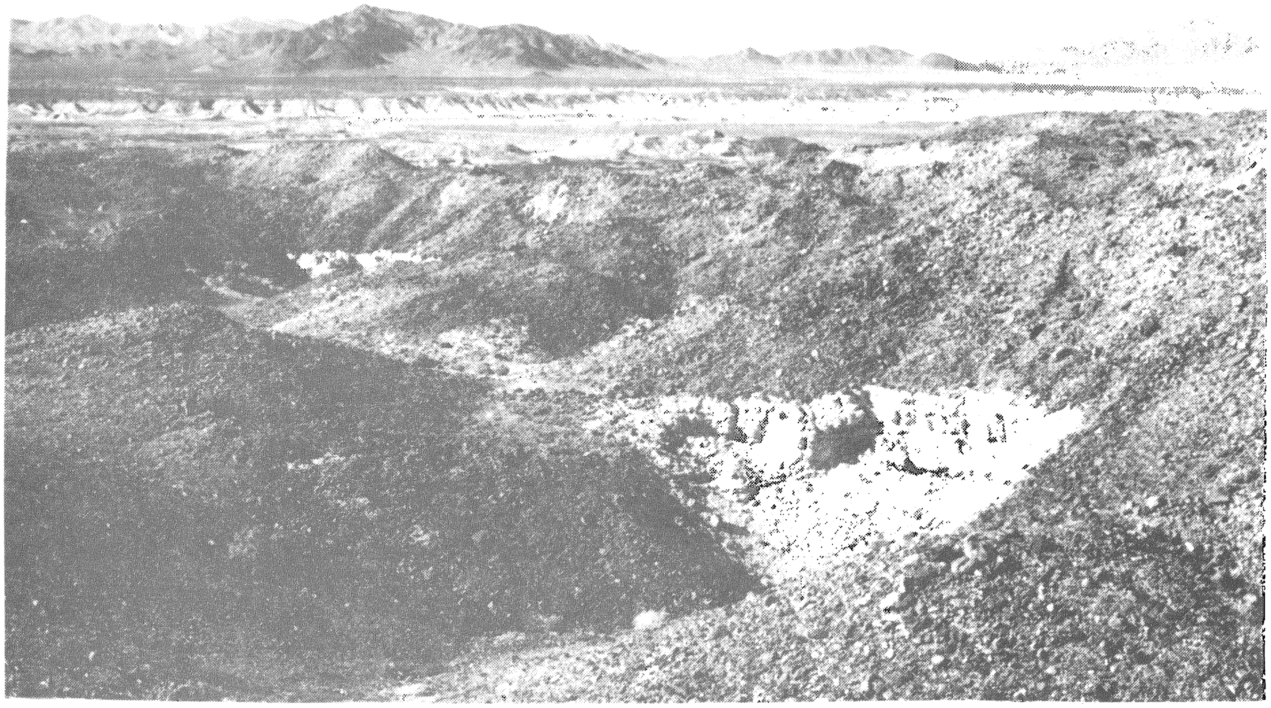

Figure 13. Landscape formed by 'megapiping' in the lacustrine deposits of the Tecopa basin. Piping is a significant erosion process in these lake beds, and piping landforms dominate large areas of the basin. The sinkhole in the foreground is approximately 20 m wide and 10 m deep. The high erosional scarp in the far distance marks the dissected basinward margin of the Greenwater erosion surface.

Figure 14. Large sinkhole, approximately 15 m wide and at least 8 m deep, in the lacustrine deposits of the Tecopa basin. Hundreds of these depressions disfigure large areas of the dissected lake beds along the western and southern margins of the basin.

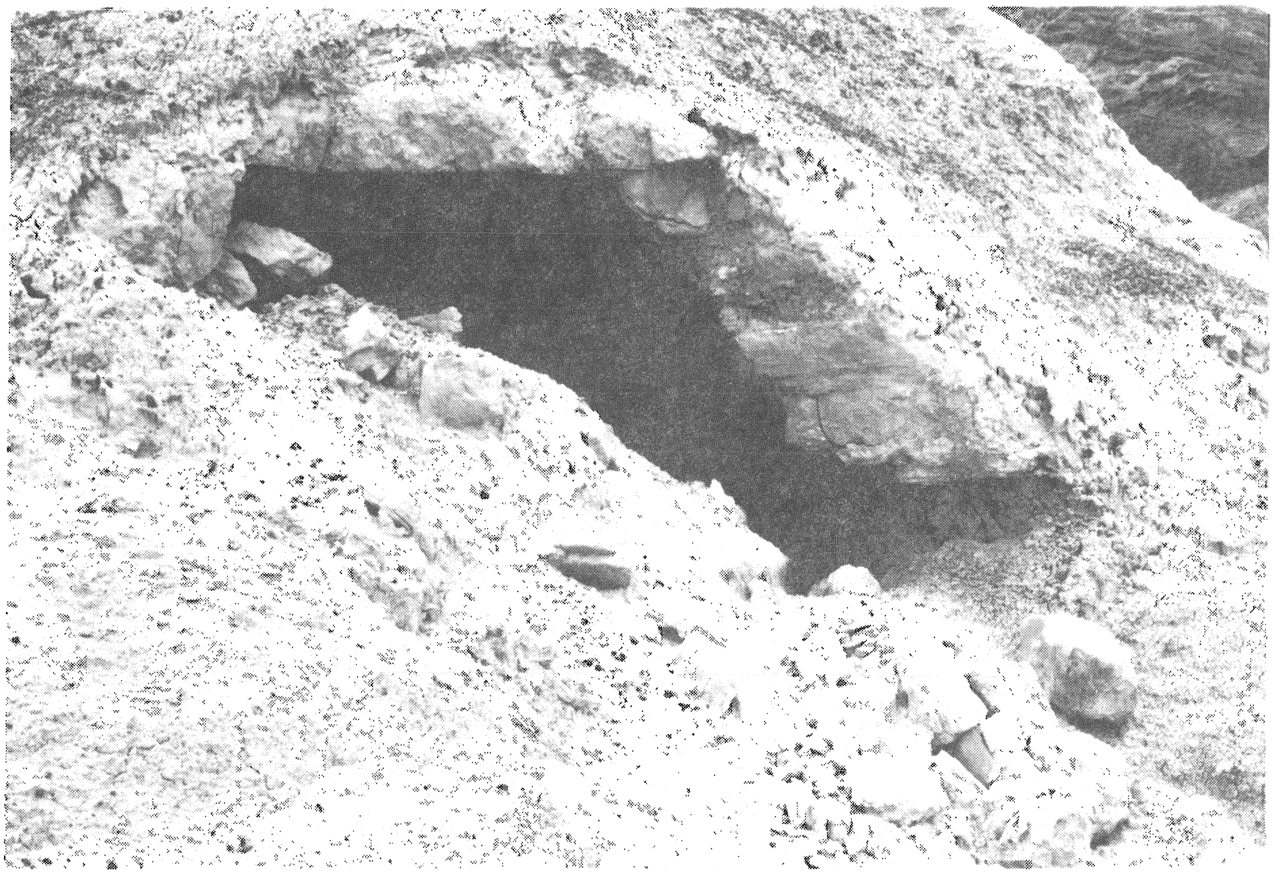

Figure 15 Outlet of a large piping tunnel in the lacustrine deposits of the Tecopa basin. This tunnel is approximately 3 m wide and 2.5 m high. Tunnels as much as 100 m long with rooms up to 5 m high and 8 m wide have been located within these lake beds (Rogers, 1981).

drainageways that incise the basin below the level of the high lake shoreline, (b) the Greenwater surface (between 0.25 and 0.15 m.y. old) which bevels the lacustrine deposits of the basin, and (c) the Sperry surface (somewhat older than 0.3 to 0.5 m.y.) which grades to the approximate level of the high shoreline but is truncated by that shoreline.

(3) Comparison of the relative positions of the original lake bed surface and the post-lake erosion surfaces (the Amargosa and Greenwater surfaces) provides a general picture of post lake dissection: (a) rapid dissection of the original lake bed surface to a baselevel of approximately 440 m (the level of the Greenwater surface); (b) formation of the Greenwater surface by complete removal (at an average rate between 8 and 25 $cm/10^3$ yr) of 20 to 30 m of lacustrine deposits from the central part of the basin; and (c) rapid dissection of the Greenwater surface to the present baselevel of 410 m by local erosion of as much as 30 m and general erosion (at an average rate between 6 and 10 $cm/10^3 yr$) of 10 to 20 m of lacustrine deposits from the central part of the basin. The initial erosional pulse and formation of the Greenwater surface probably took place between 0.5 and 0.25 m.y.B.P., and the onset of the second erosional pulse probably occurred between 0.25 and 0.15 m.y.B.P.

(4) Alluvial systems in most upper and middle piedmont areas of the basin have not been significantly affected by draining of the lake and dissection of the lacustrine deposits. Post-lake dissection has followed a general pattern of relatively deep and complete dissection in the central part of the basin accompanied by a dramatic and systematic decrease in both the depth and completeness of dissection with increasing distance from the basin axis. As much as 50 m of lacustrine deposits have been completely eroded from axial areas. However, significant dissection ends near the level of the high lake shoreline, and many alluvial systems have aggraded slightly in middle piedmont areas, probably in response to Quaternary climatic change.

(5) Post-lake dissection of the lacustrine deposits of the Tecopa basin has been strongly affected by concentrated subsurface flow processes. Indications of long-continuing concentrated subsurface flow (including sinkholes as much as 30 m wide and 10 m deep, vertical pipes 1 to 2 m in diameter and at least 15 m deep, blind valleys up to 10 m deep and several 100 m long, and tunnels as much as 100 m long) are widely and abundantly distributed across large areas of the basin. In several areas, these large piping landforms have coalesced into chaotic landscapes of irregular low ridges and hills lacking integrated surface drainage networks. Several geologic, geomorphic, and climatic factors combine within the basin to produce ideal conditions for large-scale piping: (a) sub-horizontally bedded, poorly to moderately indurated lacustrine deposits with low intrinsic permeabilities; (b) high infiltration capacities produced by extensive cracking; (c) high silt-clay contents with high percentages of expandable clays; (d) high potential evapotranspiration coupled with low mean annual rainfall and the occasional occurance of high intensity, short duration storms; (e) deep dissection providing abundant free faces and steep hydraulic gradients adjacent to these free faces; and (f) strongly indurated layers of carbonate-cemented mudstone which form and protect the roofs of the larger pipes. Where all of these factors come together, concentrated subsurface flow processes locally outstrip overland flow processes and large-scale piping landforms dominate the landscape.

TABLE 1

PRINCIPAL ASH BEDS WITHIN THE LACUSTRINE DEPOSITS OF THE TECOPA BASIN
(Data summarized from Hillhouse, in press)

ASH BED (nomenclature of Sheppard & Gude, 1968; Hillhouse, in press)	ISOTOPICALLY DATED SOURCE UNIT	THICKNESS	STRATIGRAPHIC POSITION
Tuff A	Ash of the Lava Creek Tuff (0.62 m.y.)	0.5 to 4.0 m	4 to 8 m above Tuff B; 10 to 30 m below the uppermost lake beds
Tuff B	Ash of the Bishop Tuff (0.73 m.y.)	0.05 to 3.5 m	About 18 m above the base of Tuff C; 4 to 8 m below the base of Tuff A
Tuff C	Ash of the Huckleberry Ridge Tuff (2.02 m.y.)	0.1 to 0.75 m	About 17 m above the oldest exposed lake beds

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