Quaternary Geomorphology and Geochronology of Owens Valley, California: Geological Society of America Field Trip

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INTRODUCTION

Owens Valley, in the shadow of the Sierra Nevada, has a rich and varied Quaternary geologic history of glaciation, volcanism, seismicity, and alluviation. During this field trip we will examine the results of each of these processes while reviewing the Cenozoic history of the region.

This field guide is divided into two sections. The first section includes an introduction to some of the major geomorphic processes active in Owens Valley and outlines the Quaternary history. The second section provides descriptions of each field-trip stop.

SECTION 1: GEOMORPHOLOGY OF OWENS VALLEY

GEOGRAPHY

Owens Valley, bounded on the west by the Sierra Nevada and on the east by the White and Inyo ranges, is a complex graben which defines the western boundary of the Basin and Range Province (Fig.1). The valley trends northwest-southeast, is approximately 30 km wide, and extends for 150 km from Rose Valley in the south to the Volcanic Tableland just north of Bishop, California. The valley floor rises gently from about 1050 m in elevation south of Owens Lake to 1300 m just north of Bishop. Chalfant Valley, the northern extension of Owens Valley, lies between the Volcanic Tableland and the White Mountains. The topography is dramatic, with both the White Mountains and the Sierra Nevada rising to elevations of 3500 - 4000 m. The Inyo range is slightly lower, with crest elevations of 3050 - 3350 m. Flanking both the White Mountains and the Sierra Nevada, south of Big Pine, are broad piedmonts of coalescing fans (bajadas), which rise from ~1200 to ~1800 m in elevation.

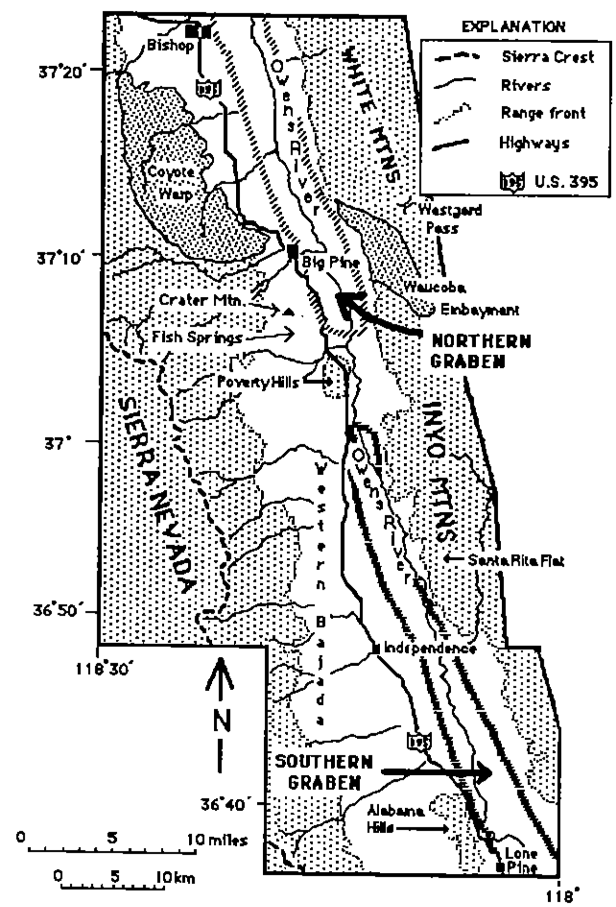


Fig. 1. Schematic map of Owens Valley showing the locations of the deep basins of the buried graben. The Owens Valley fault zone lies on the western margin of the basins.

Under the influence of the Sierra Nevada rain shadow, the local climate is semiarid with hot, dry summers and cold winters. Average annual precipitation in Bishop is 14.5 cm, with most falling between November and March. Mean annual temperature (in Bishop) is 12.8° C: 2.8° C in January and 25.0° C in July (NOAA data). Perennial creeks flow on the Sierra side, whereas the Whites and Inyos support only intermittent stream flow. The sparse rainfall supports alkali grasslands and saltbush communities on the valley floor,blackbrush, sagebrush, and a variety of grasses on bajada surfaces, and a sparse piñon - juniper woodland at higher elevations. There are ancient Bristlecone Pine forests at high elevations (3050 - 3700 m) in the White Mountains.

The southern end of Owens Valley contains Owens Lake, now a playa due to removal, in the Los Angeles aqueduct, of water from Owens River and its tributaries . During pluvial periods in the Pleistocene, Owens Lake overflowed to the south, and Sierra Nevada water eventually reached Lake Manly in Death Valley.

TECTONICS

In cross section, Owens Valley is asymmetric, consisting of a western bedrock bench shallowly buried by alluvial fans, and a narrow eastern zone containing two major buried basins (Fig. 2). The basins are filled by fluvial and lacustrine sediments, and have no surface expression. The basins are separated near the Poverty Hills, where bedrock is only ~300 m beneath the surface.

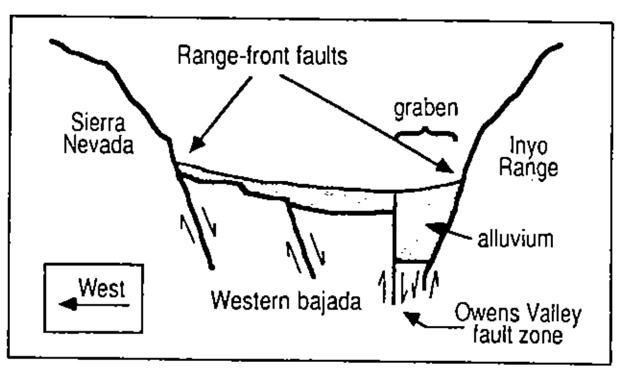


Fig. 2. Schematic east-west cross section of Owens Valley near 37° N latitude.

Geophysical studies [*Pakiser et al.*, 1964; *Hollett et al.*, 1989] have shown the northern basin to be as deep as 1.2 km (from the alluvial surface); the southern basin is as deep as 2.1 km. Its floor has subsided nearly 4 km relative to the crest of the Sierra Nevada.

From Big Pine Creek to Bishop Creek, the simple Sierra Nevada escarpment is disrupted by a high crustal block, the Coyote warp [*Bateman*, 1965]. Although truncated on the east by normal faults, the boundary between this block and the Sierra Nevada to the west is less distinct. It has tilted to the northeast. Analogous blocks occur along the eastern side of Owens Valley, including Santa Rita Flat and the Waucoba embayment.

Valley Formation

During much of the Tertiary Period, the ranges bordering present-day Owens Valley were probably united as a broad upwarp. They share common Mesozoic granitic rocks, well exposed in the more heavily eroded Sierra Nevada but still widely capped by Paleozoic marine sedimentary rocks in the White-Inyo Ranges. Subsidence of Owens Valley has occurred along normal faults that delineate the range fronts on either side.

The formation of Owens Valley must be viewed in the context of the uplift of the Sierra Nevada, and of the White - Inyo Mountains. Christensen [1966] and Crough and Thompson [1977] thought that uplift began in the southern Sierra and propagated northward. This view is consistent with the tectonic interpretation of Hay [1976], that the latest uplift began with the northward migration of the Mendocino triple junction and accelerated about 4.5 Ma, coincident with increased slip along the San Andreas Fault and the opening of the Gulf of California. The triple junction is thought to have been at the latitude of Mt. Whitney at ~10 Ma [*Atwater and Molnar*, 1973].

The upland regions of the Sierra Nevada contain benches and plateaus thought to be remnants of ancient surfaces eroded during hiatuses in uplift. In the central Sierra Nevada (Mammoth Lakes region), Huber [1981] reported evidence of ~1 km uplift in the past 3 million years, with much of it in the latter half of that period. Bachman [1978] showed that uplift of the White-Inyo Mountains, at least relative to Owens Valley, began late in the Pliocene Epoch, and that the ranges were prominent by about 2 Ma. DePolo [1989] established that in the White Mountains uplift began around 3 Ma. The pattern that emerges is one of episodic uplift on the million-year time scale or longer, with continuing tectonism in the Pleistocene.

Owens Valley appears to have begun subsiding during the Pliocene Epoch, deepening during the Pleistocene. Bateman [1965], on the Coyote warp southwest of Bishop, found ~9.6-million-year-old lavas that flowed towards present-day northern Owens Valley, suggesting that a shallow depression existed there at that time. Range-front faulting of dated basalt flows in the southern Inyo Range suggests that the major subsidence of southern Owens Valley began by 5-6 Ma [*Giovanetti*, 1979; *Bacon et al.*, 1979]. Matthes [1950] noted geomorphic evidence from the Sierra escarpment near Mt. Whitney that indicated that southern Owens Valley may have been as deep as ~1 km during the Pliocene. However, Bachman [1978] interpreted the Waucobi lake beds of central Owens Valley to indicate that by ~2 Ma there was still little relief between the valley floor and the White-Inyo Mountains. Though the above evidence is not compelling, it can be interpreted to mean that the southern Owens Valley subsided first.

Observations of the Sherwin till and Bishop Tuff permit independent estimation of the amount of subsidence of Owens Valley since ~0.8 Ma. Sherwin till deposited on the flank of the Long Valley volcano weathered subaerially for ~0.1 Ma until it was buried by the Bishop Tuff, erupted at 0.73 Ma [*Mankinen et al.*, 1986]. On the Coyote warp, Coyote Creek has eroded through the Sherwin till and the underlying bedrock to a depth of ~240 m. Provided the erosion of Coyote Creek has been due only to base level changes associated with the subsidence of Owens Valley, this estimate is a minimum for valley deepening since ~0.8 Ma.

The Bishop Tuff has been buried by as much as 210 m of alluvium south of Bishop [*Hollett et al.*, 1989], due in part to southward tilting of the tectonic block and deepening of the graben. This depth and the erosion of Coyote Creek together provide a rough estimate of the total subsidence of northern Owens Valley - ~450 m - since ~0.7 Ma.

The eastern escarpment of the Sierra Nevada preserves evidence that the subsidence of Owens Valley occurred in three main episodes. The escarpment shows two zones of faceted ridges, defining bevels of ~950 m (upper) and ~750 m (lower). Ridges from the apices of the upper zone of facets to the Sierra crest define a third plane, which is probably a heavily eroded facet, perhaps the western side of the proto-Owens Valley. It is the discreteness of the facets that suggests that subsidence was episodic. Similar facets are observed on the western escarpment of the White Mountains.

At first, Owens Valley was a shallow graben east of the present-day crest. Subsequent displacement occurred across the range-front faults in two major episodes separated by a significant hiatus, or by a change in faulting rate. If down-faulting rates were ~1mm/year, typical of rapid faulting near Long Valley today, the tectonic episodes must have been on the order of a million years in duration. The existence of two or three episodes of subsidence in the late Tertiary and Quaternary Periods is consistent with this estimate.

Neotectonic Activity

Owens Valley has been very active tectonically during the Quaternary. The three main fault systems in Owens Valley, the normal range-front faults on either side of the valley, and the right-lateral oblique-slip Owens Valley fault zone along the central axis (Fig. 2), all show evidence of slip during the Holocene. Studies of range-front faulting have been conducted by Gillespie [1982], dePolo [1989] and Bursik and Sieh [1989]. Range-front faulting in the Inyo Mountains. has apparently received little attention. The Owens Valley fault zone has most recently been studied by Lubetkin and Clark [1988], Martel et al. [1987] and Beanland and Clark [1991].

The range-front faults displace alluvial, colluvial and glacial deposits at the base of the Sierra, White and Inyo escarpments. Motion on the western range-front faults is primarily normal with displacements of as little as 6 m on late-Pleistocene ("Tahoe") moraines, although greater displacements (>25 m for Tioga) have been measured to the north, in Long Valley. Late Pleistocene vertical-displacement rates across the west-side range-front faults have generally been <0.2 mm/year. This value is less than the average rate at Coyote Creek (~0.3 mm/year across the range front), and is markedly lower than the rates of 1 -2 mm/year found north of Owens Valley, for Mono Basin and the White Mt. fault zone.

The Owens Valley fault system runs along the axis of Owens Valley, displacing alluvial and lacustrine deposits. Motion on this fault is right-lateral oblique slip. Total vertical displacement is ≤2.1 km; horizontal displacement is several km. Strain measurements show modern displacement rates of 2.2 mm/year across the range-front and Owens Valley fault zones, near the town of Lone Pine

[*Savage et al.*, 1975; *Savage and Lisowski*, 1980]; the late-Pleistocene vertical displacement rate on one strand of the Owens Valley fault zone may be as much as 0.6 mm/year. These values greatly exceed the long-term average rate of ~0.4 mm/year for the southern Owens Valley fault zone [*Martel et al.*, 1987]. Near Bishop, vertical displacement of <0.3 mm/year is estimated from the burial of the Bishop Tuff.

An earthquake in 1872 resulted in ground breakage along about 100 km of the fault. The moment magnitude of this quake was estimated at 7.5-7.7 by Beanland and Clark [1991]. The 1872 quake [*Oakeshott et al.*, 1972] was felt as far away as Oregon and Salt Lake City; the effects of the quake in Yosemite Valley, ~150 km away, were dramatically reported by John Muir. Stop 13 examines a scarp that records displacement of the 1872 quake.

Even within the present episode of valley subsidence, tectonic rates have apparently been variable. There is growing evidence that subsidence rates changed around 0.7 Ma, possibly following the eruption of the Bishop Tuff. DePolo [1989] found that the vertical displacement along the front of the White Mountains increased to its present high rate (~1 mm/year), following a period of relative quiescence. In Deep Springs Valley, east of the Waucoba embayment, most subsidence evidently postdated 0.7 Ma [*Reheis and McKee*, 1991]. In central Owens Valley, subsidence may have increased after 1.2 Ma and then slowed before ~0.5 Ma [*Gillespie*, 1991c].

Locally, range-front faulting rates appear to have have varied on a time scale of 10^4-10^5 years. Detailed studies of offset lava flows and alluvial fans at Sawmill Creek, near the town of Independence, indicate that vertical displacement rates of ~0.4 mm/year between 120 and ~65 (?) ka decreased to <0.2 mm/year thereafter. Bursik and Sieh [1989] concluded the subsidence of Mono Basin across the range-front faults was faster before 70,000 years ago, slowing since then. However, at Fish Springs, Martel et al. [1987] found no compelling evidence of late Pleistocene changes in the vertical displacement rate across a strand of the Owens Valley fault zone.

LATE QUATERNARY VOLCANIC HISTORY

The Mono Basin-Long Valley-Owens Valley region has undergone many episodes of volcanism during Plio-Pleistocene time, and continues to be one of the most active eruptive centers in North America. The late Cenozoic volcanic history of the region is detailed in Bailey et al. [1976], Bailey [1982, 1984], and in several articles in a special volume of *Journal of Geophysical Research* [v. 90, 1985, pp. 11,111-11,289].

Volcanism in the region began ~3.2 Ma, early in the history of range-front faulting along the east side of the Sierra Nevada. Early eruptions were largely basaltic, but the culminating event in the region was the catastrophic eruption at 0.73 Ma of 600 km^3 of silicic magma from beneath modern Long Valley [*Hill et al.*, 1985]. The resulting Bishop ash deposits covered 1500 km^2 to a thickness of up to 1500 m; the subsequent collapse of the roof of the magma chamber formed the Long Valley caldera. Since that event, a resurgent dome has formed near the center of the valley and smaller, less explosive eruptions of coarsely porphyritic rhyolite and more mafic magmas have periodically occurred around the periphery of the caldera.

During this later eruptive phase, a similarly bimodal system of small volcanic eruptions has developed along a series of fissures stretching 40 km northward from the western rim of the Long Valley caldera to Mono Lake. These fissure and dome eruptions, ranging in age from 300 to ~0.2 ka, include Devil's Postpile (trachybasaltic/andesitic lavas), Mammoth Mountain (quart latite/rhyodacite flows), the Inyo and Mono craters (rhyolite dome and flows), and Black Point and the islands of Mono Lake (basalt to rhyodacitic flows and domes) [*Hill et al.*, 1985; Fig 3].

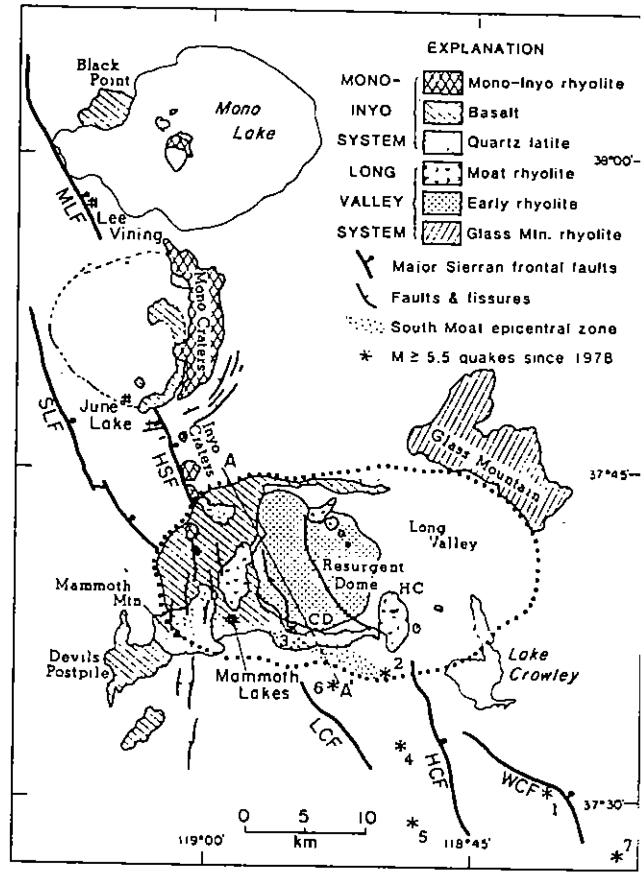


Fig. 3. Geologic map of the Long Valley/Mono Basin region showing the distribution of volcanic rocks related to the Long Valley caldera magmatic system and the younger Inyo/Mono magmatic system. HSF, Hartley Springs fault; HCF, Hilton Creek fault; SLF, Silver Lake fault; WCF, Wheeler Creek fault; CD, Casa Diablo; HC, Hot Creek (from Hill et al., 1985).

The most recent eruptions have occurred along the Inyo and Mono craters chains, and in Mono Lake. Most of the presently exposed domes and deposits here are Holocene [*Wood*, 1977; *Stine*, 1984; *Miller*, 1985], although extensive deposits of earlier eruptions underlie the surface deposits [*Lajoie and Robinson*, 1982; *Bursik and Sieh*, 1989]. Eruptions have occurred within the last 600 years in both the Inyo Craters and Mono Craters, but the youngest events in the region were eruptions of dacitic lavas and pyroclastic flows on Paoha and Negit islands in Mono Lake apparently less than 20(years ago [*Sieh and Bursik*, 1986; *Stine*, 1984]. Whereas the eruptive record suggests that much of the recent volcanic activity in the region has centered in the Mono Basin, a swarm of shallow seismic events, including several of magnitude 5 to 6, occurred along the southern margin of the Long Valley caldera between 1978 and 1984 [*Lide and Ryall*, 1985]. These events were accompanied by inflation of the resurgent dome, leading the USGS to issue a volcanic hazards warning for the Mammoth Lakes area during this time.

The Big Pine volcanic field, in central Owens Valley, is one of several late Cenozoic bimodal subalkali - alkali basalt fields in the southwestern United States. The volcanic field has been interpreted as a group of magma plumes, rising rapidly from an upper mantle source. Initial eruptions consist of this material, poor in assimilated upper crustal rocks. Subsequent flows from the same vent are increasingly affected by differentiation of the source material and by assimilation of granitic country rocks.

The field contains ≥40 vents within a ~1000-km^2 region centered south of the town of Big Pine. Roughly 120 km^2 (~2.5 km^3) of lava is now exposed, and more is buried. The greatest volume was erupted from the northern margin of the field, near the septum between the two grabens that comprise the deepest part of Owens Valley. Vents are generally localized along normal faults bounding the ranges. The largest cones contain ~0.05-0.15 km^3 of agglutinate and cinders and rise to heights of 220-260 m above their bases; the largest flows are 9 km in length and 3 km in width. The field contains a single rhyolite dome (1.0 Ma) consisting of ~0.05 km^3 of pumiceous rhyolite and obsidian.

The Big Pine field appears to have been active episodically throughout the entire Quaternary Period [*Gillespie*, 1991a; see Plate I for map]. The oldest rocks may be late Pliocene dikes; however, the oldest well dated lavas are 1.2-Ma ridge-capping flows and eroded cinder cones. Most of the exposed flows date from ~0.1-0.5 Ma. Flows of this age range are lightly mantled by silt and partly alluviated; the flanks of coeval cones are steep-sided and not gullied, and their bases are buried by alluvium. Erosion is less than for coeval cones from (for example) Cima volcanic field, ~150 km to the south, perhaps because there has been less influx of eolian silt in Owens Valley, and therefore increased percolation of rain and less runoff. The youngest flows and cones, erupted only shortly before the end of the latest Pleistocene glaciation (~10-25 ka), are virtually unweathered and uneroded.

Late Quaternary Climate History

The climate of the basins that bound the eastern side of the Sierra Nevada has undergone profound changes during the Quaternary Period. Lacustrine, floral, and glacial records indicate a complex history of fluctuations in precipitation and temperature during the most recent glacial cycles.

The net annual moisture budgets for Mono Basin and Owens Valley, part of the internal drainage of the Great Basin, were substantially increased during recent glaciations. Under current semiarid conditions, Mono Lake and Owens Lake are the only large perennial bodies of water in the region, although lake levels appear to have fluctuated throughout the Holocene. During the Late Wisconsin (Table I) and earlier Pleistocene glaciations, the effective moisture budget of the region was sufficient to increase the volume of these lakes, and integrate the Owens River system from basin to basin as far as Lake Manly (Death Valley) [e.g., *Smith and Street-Perrott*, 1983; *Smith*, 1984; Fig. 4].

Paleoecologic studies of the region, based primarily on pollen and packrat-midden data, indicate complex changes and shifts in biogeography during the late Pleistocene and Holocene Epochs. Currently, Owens Valley lies within a vegetation zone that is transitional between the cool deserts of the Great Basin to the north and the warm Mojave and Sonora deserts to the south. This boundary corresponds roughly to the present northern limit of summer incursions of monsoonal moisture. Increased winter moisture, weakened monsoons, and regionally cooler temperatures during the Late Wisconsin glaciation pushed the boundary southward into the Sonoran desert. These climatic shifts also caused elevational lowering of floral communities by as much as 1000 m, such that much of the scrub-covered lower flanks of Owens Valley was dominated by open-canopy juniper and pine woodland [*Spaulding et al.*, 1983; *Spaulding and Graumlich*, 1986]. At the glacial/interglacial transition, between about 12 and 8 ka, the cooler, wetter glacial conditions were replaced by increased temperatures and summer precipitation, causing a northward shift in the warm/cool desert boundary. After 8 ka, monsoonal incursions weakened and the boundary shifted to roughly its present position in the southern Owens Valley.

Deposits of former glaciers in the mountains next to the Owens and Mono basins provide a proxy record of climate. The mass balances of glaciers are highly sensitive to changes in climate. By reconstructing equilibrium-line altitudes (ELAs) of former glaciers

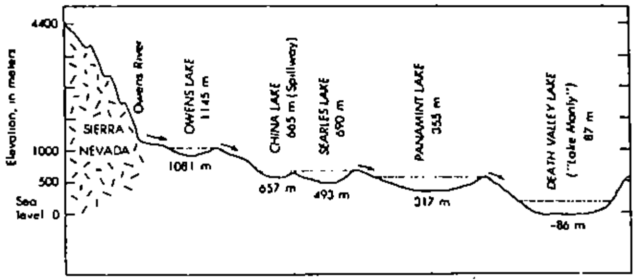


Fig. 4. Schematic cross section of pluvial lakes downstream from the Owens River showing elevations of floors of basin and lake surfaces during high stands (from Smith and Street-Perrot, 1983).

Table 1: Summary Chronology of Glacial Deposits in the South-central Sierra Nevada.

AGE	TIME (ka)	GLACIAL ADVANCES *w/ Numerical Age Constraints*
HOLOCENE	0-10	NEOGLACIATIONS (Matthes, Recess Peak)
		• *9.99 ka (^{14}C, Adam, 1967)* • *10.6 ka (^{14}C, Metzger, 1986)* • *9.5, 11 ka (^{14}C, Batchelder, 1980; Price, 1982)*
LATE WISCONSIN	10-35	HILGARD (?)
		• *15.6 ka (^{14}C, Batchelder, 1980; Price, 1982)*
		TIOGA • *14.1 ka (^{14}C, Marchand and Allwardt, 1981)*
		• *21 ka (^{14}C, Lubetkin, 1980)* • *23, 27 ka (^{14}C, Bursik and Gillespie, in prep.)*
		TENAYA
MIDDLE WISCONSIN	35-65	
EARLY WISCONSIN	65-79	"LATE" TAHOE (stage 4 or 5 (?))
"EOWISCONSIN"	79-122	
SANGAMON	122-132	• *118, 119, 131 ka (^{40}Ar-^{39}Ar, Gillespie 1982)*
ILLINOIAN	132-302	"EARLY" TAHOE (stage 6)
		MONO BASIN
PRE-ILLINOIAN	302-	• *463 ka (^{40}Ar-^{39}Ar, Gillespie, 1982)* • *730 ka (K-Ar, Bishop Tuff; Mankinen et al., 1986)*
		SHERWIN
		MCGEE
		• *2.71 Ma (K-Ar, Mankinen and Dalrymple, 1979)*

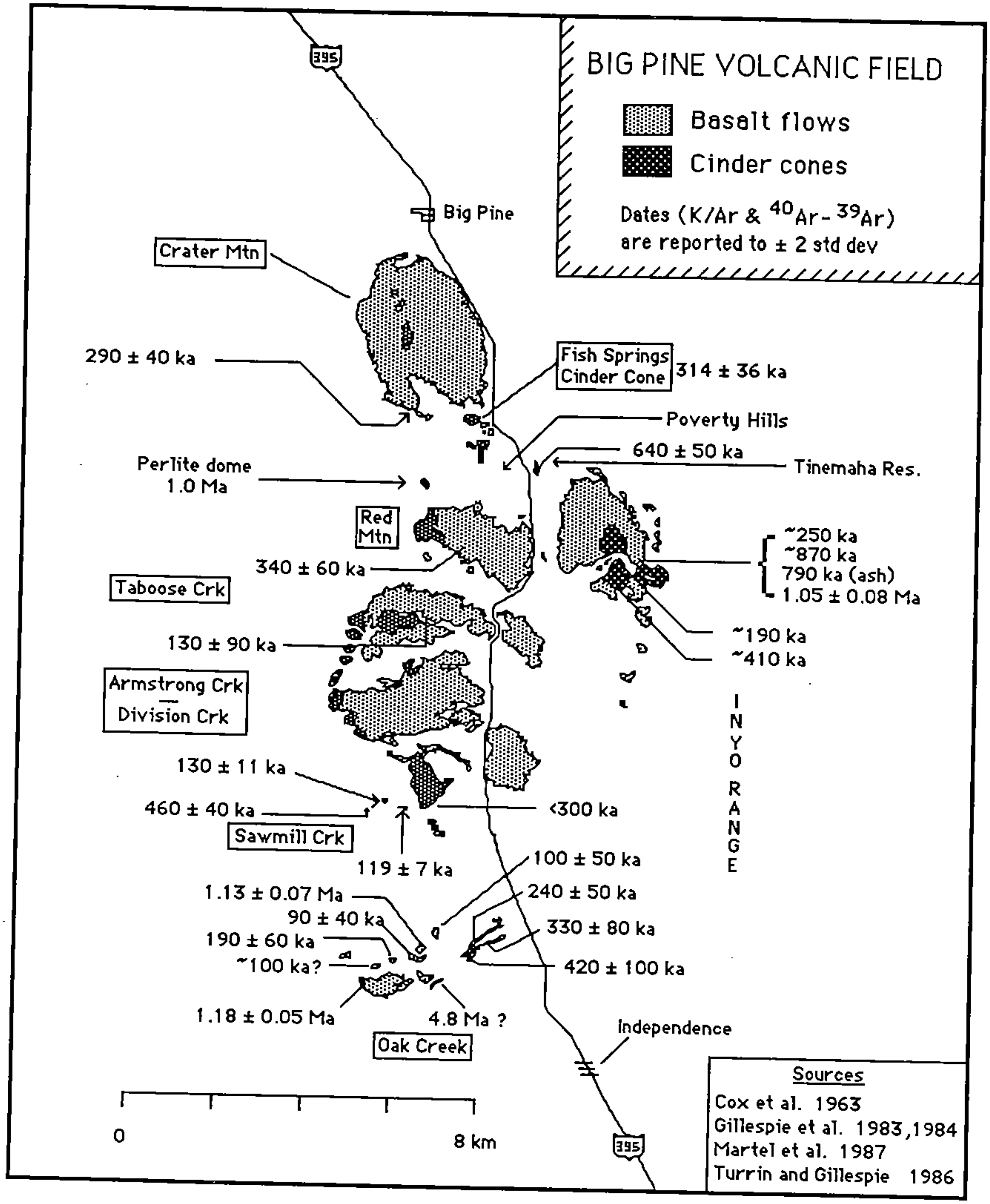


PLATE I. Map showing radiometric ages and location of samples for the Big Pine volcanic field for different glaciations along a regional transect, it is possible to detect changes in climate. Applying this technique in the southern Sierra, Gillespie [1991b] found that the N-S ELA gradient of the Tahoe glaciation was significantly discordant from that of other late Pleistocene glaciations. This discordance probably reflects an inadvertent "lumping" of two distinct glaciations (probably ^{18}O stages 4 and 6) rather than a substantially different climate for a single glaciation. This conclusion is important because it indicates that Early Wisconsin glaciers (stage 4) in the region were larger than Late Wisconsin (stage 2) glaciers, a relationship that contrasts with the marine global ice-volume record. Gillespie and Clark are currently expanding this research into the northern Sierra Nevada and adjacent ranges to investigate further these initial findings.

LATE QUATERNARY GLACIAL HISTORY

Glacial-geologic mapping in the Sierra Nevada indicates that at least seven Pleistocene glaciations and three Holocene advances have occurred in the range (see Wahrhaftig and Birman [1965] and Fullerton [1986] for detailed summaries). Blackwelder's [1931] original fourfold sequence of Pleistocene glaciation for the eastern Sierra Nevada has been refined and expanded by subsequent work (Table I). The McGee glaciation is evidenced only by scattered erratics and weathered till remnants preserved high above later deposits on isolated mountain tops and ridges. Sherwin deposits are much more extensive than either earlier or later glacial deposits, and it is generally thought that this was the largest Pleistocene glaciation in the Sierra Nevada. Only moraines of the four youngest Pleistocene glaciations, Mono Basin, Tahoe, Tenaya, and Tioga (of which we will see examples), are morphologically preserved, and apparently represent late-middle Pleistocene through late Pleistocene glacial periods (stages 6 through 2). In addition to the Pleistocene glaciations, at least three smaller Holocene (and/or latest Pleistocene) glacial advances have been proposed (Table I).

Uncertainties about numerical ages have led to persisting controversies over the timing and correlation of these glaciations at both range-wide and drainage-to-drainage scales. McGee till overlies and is probably substantially younger than a basalt flow dated at 2.71 Ma [*Dalrymple*, 1964]. A younger limiting age for the glaciation is possibly given by a deeply weathered till, tentatively designated as McGee till, that is separated from overlying Sherwin till by a well developed paleosol. This relationship suggests that the McGee (?) till is at least 100 thousand years older than the Sherwin till [*Sharp*, 1968; *Birkeland et al.*, 1980]. These same workers argued that the Sherwin glaciation predated deposition of the Bishop Tuff (0.73 Ma) by at least 50 thousand years on the basis of an exposure near Long Valley in which a thick paleosol that is developed in Sherwin till is overlain by the tuff. The Casa Diablo diamicton may represent a glaciation intermediate in age between the Sherwin and Mono Basin/Tahoe glaciations, but its timing and genesis are questionable.

The timing of the four youngest Pleistocene glaciations is more controversial. In general, the Mono Basin has been designated Illinoian (stage 6), the Tahoe has been alternatively assigned Illinoian or early Wisconsin (stage 4), and the Tenaya and Tioga are generally felt to be Late Wisconsin (stage 2). However, only broad limiting numerical dates exist for these deposits (Table I). The best constraints on the age of Tahoe glaciation come from Sawmill Creek near Independence, where a double moraine complex is separated from an underlying till by a basalt flow that dates to the last interglaciation [*Gillespie*, 1982; *Gillespie et al.*, 1984; Table I]. Earlier workers argued over whether the overlying deposits represented one (Tioga) or two (Tioga and Tahoe) glaciations. However, by quantitative analyses of boulder weathering, Gillespie [1982] established that the two overlying moraines were from distinct glaciations, and thus that the Tahoe glaciation in this drainage was Early Wisconsin in age. Correlating to other drainages in the area, Gillespie [1982] felt that moraines designated as 'Tahoe' commonly included deposits of two distinct and separable glaciations, one Illinoian ("early" Tahoe) and one Early Wisconsin ("late" Tahoe) in age.

FAN SEDIMENTATION PROCESSES

Broad piedmonts of coalescing fans (bajadas) are characteristic landforms in the Basin and Range. Owens Valley fans, unlike those in Death Valley [*Denny*, 1965], are debris-flow fans and not, strictly speaking, alluvial fans [*Beaty*, 1963; 1970; *Hubert and Filipov*, 1989; *Whipple and Dunne*, 1991]. The distinction is non-trivial.

Because the processes of sediment delivery and deposition on debris-flow fans are distinct from those on alluvial fans, basins with different sediment delivery mechanism(s) may respond differently to episodes of climatic change. For example, the debris-flow fans on either side of Owens Valley have different characteristic surface morphologies and have experienced different "alluviation" histories. Fans flanking the Sierra Nevada most recently aggraded in the late Pleistocene whereas those below the White Mountains have continued to aggrade throughout the Holocene

Climate Change and Fan Aggradation

We propose that aggradation along the Sierra Nevada has been episodic and that phases of fan building are correlated with the late Pleistocene glacial advances [*Gillespie*, 1982]. During glaciations, supra-glacial and morainal debris provide an abundant source of poorly sorted, unstable debris which would be particularly susceptible to mobilization as debris flows. During interglaciations, debris-flow generation is greatly retarded because: 1) After deglaciation the source of morainal debris is reduced as moraines stabilize; 2) basin side slopes have been stripped of mobilizable colluvium during the preceding glaciation; and 3) the development of a stepped longitudinal valley floor profile effectively eliminates the upper two thirds of some basins as a potential sediment source.

Conditions are quite different in the White Mountains. Steep tributary gullies (little more than bedrock chutes) are choked with an abundant supply of loose, poorly sorted colluvium [*Beaty*, 1990], which contains abundant fines and readily mobilizes into debris flows if saturated during intense thunderstorms or rapid snowmelt. Debris-flows are a relatively common occurrence in today's semi-arid climate [*Beaty*, 1963; 1970]. Although the response of the White Mountain fan system to climatic changes is not easily predicted, the primary controls on debris flow activity are the frequency of precipitation events capable of mobilizing debris flows, the effectiveness of vegetation at stabilizing talus slopes, and the long-term availability of poorly sorted source debris.

Morphology of the Owens Valley Fans

The fans flanking the White Mountains and those at the foot of the Sierra Nevada exhibit important differences in both gross morphology (size and slope) and surface morphology (surface texture and channel network patterns). The White Mountain fans are generally smaller and steeper than the fans along the Sierra Nevada - a difference which probably reflects differences in the balance between uplift and sedimentation rates. On a finer scale, the differences in fan surface morphology are probably related to differences in lithologic and hydrologic conditions, as discussed below.

The surfaces of the Sierran fans are characterized by many well-preserved abandoned, boulder-lined channels with little relief on levees, narrow bouldery debris-flow snouts, and generally smooth low-relief interfluves (Figs. 5 and 6). The surfaces of the White Mountain fans are dominated by on-lapping wide terraces defined by broad overbank debris flow lobes that add relief to interfluves, broad channel-margin levees built up to levels well above general elevation of the surrounding fan surface, fewer abandoned channels, and fewer large boulders (Fig. 7).

Sedimentation Processes on Debris-Flow Fans

Debris-flow and fluvial processes play distinct roles in fan construction. The nature of Owens Valley fan sediments suggests that fan surface aggradation is accomplished by debris-flow deposition [*Beaty*, 1963; *Hubert and Filipov*, 1989]. However, fluvial processes are generally responsible for the channelization of the fan surface. The fluvial channels serve as an important guide for debris-flows traversing the fan surface, and thus influence the pattern of debris-flow deposition. Debris flows follow existing stream courses, depositing material on various parts of the fan depending on (1) the nature of the channel system, and (2) the hydrographs and rheology of the debris flows [*Whipple and Dunne*, 1991]. Mobile, fluid debris flows traverse the length of the fan, leaving thin overbank deposits wherever peak discharge exceeds channel capacity. Generally, these flows aggrade levees, smooth the fan surface, and aggrade the lower fan. Conversely, dry, relatively immobile debris flows, come to rest on the upper parts of the fan, block channels, add to surface irregularity and relief, and often cause channel avulsions by diverting floodwaters around plugged channel reaches.

Over the long-term, the frequency distribution of debris-flow rheologies delivered to the fan probably exerts the most important influence on fan surface morphology, because: 1) the spatial pattern and temporal frequency of channel-blockage events directly determine the structure of the network of abandoned channels; and 2) debris-flow driven channel avulsions are the mechanism by which long-term shifting of depositional loci is accomplished. Thus, the linkages between source area lithology, climate, hydrology, and debris-flow rheology are crucial to understanding the morphological response of debris-flow fan surfaces to climatic change.

Fig. 5. Channel on Sierran bajada, with debris-flow boulder levees. Channel is approximately 22 meters wide. View is down-fan.

Fig. 6. Oblique aerial view of the Sierran bajada, 15 km south of Independence, CA. Note characteristic pattern of discontinuous abandoned channels (20-30 m wide) with narrow boulder levees.

Lithologic Control on Fan Morphology

Source area lithology plays an indirect, but important role on fan morphology by controlling debris-flow granulometry and, hence,

Fig. 7. Oblique aerial view of Jeffery Mine Canyon fan (Stop 9), below White Mountain Peak. Incised channel is 20-25 m wide. Note pattern of broad levees and wide, on-lapping overbank debris-flow lobes.

rheology. Although the process is not well understood [see *Pierson,* 1986; *Iverson and Denlinger,* 1987], boulder-rich, sandy debris flows derived from granitic source terrains have a greater tendency to segregate into a relatively immobile bouldery snout and a wetter, more mobile tail than more clay-rich debris flows characteristic of volcanic and metasedimentary sources. This segregation process can significantly alter the frequency distribution of debris-flow rheologies delivered to the fan surface. In addition, the relatively immobile bouldery fronts may be critical in establishing channel plugs which are not easily washed away by subsequent floodwaters.

ALLUVIATION HISTORY

Alluviation in Owens Valley has been episodic, with periods of aggradation related to climatic fluctuations and glacial advances and retreats. Interestingly, aggradational pulses on the two sides of the valley do not appear to have been synchronous, and different mechanisms of control must have been at work. The most obvious difference is the glaciation of most of the Sierran canyons, and the lack of glaciation in the eastern ranges. Even between glaciations, the eastern ranges are drier, because of their position in the Sierra rainshadow.

Deposition down gradient from glaciated valleys appears to have been synchronized to glaciations. At Pine Creek (near Bishop, CA), small debris-flow fans, deposited below lateral moraines extending onto the bajada, grade to the top of now-vanished glaciers. Weathering and soil development on these fans is comparable to the development on fans of the trunk stream, which can be traced to terraces between the breached moraines. Evidently, aggradation of the fans occurred during glaciations, terminating after retreat of the glaciers.

On the Sierra bajada, late Pleistocene fans are widespread. They are well exposed, in part because Holocene deposits are notably sparse. In contrast, Holocene aggradation has been extensive below the Inyo and White Mountains., and late Pleistocene fans there are not commonly exposed. In much of the western Great Basin, they appear to be missing from the record entirely [e.g., *Sawyer,* 1991; *Slate,* 1991], and the possibility exists that aggradation during glacial periods was minimal. The question of the "missing" fans remains a perplexing topic of ongoing study.

In the long run, the filling of Owens Valley has been outpaced by tectonic deepening of the grabens. However, for the past ~0.5 Ma, as determined from dated lavas, soil profiles [*Burke et al.,* 1986] and geologic evidence [*Gillespie,* 1982, 1991c], base level for the streams of the eastern Sierra appears to have been stable, leading to a remarkable western bajada on which fans of markedly different ages are all found within a few meters' elevation of each other. The typical locus of deposition (intersection point) has neither shifted towards the valley floor nor towards the mountains during this interval.

From the stability of the base level we may infer that there was quasi-equilibrium between deposition and valley subsidence since ~0.5 Ma. However, the distal ends of the fans are all at about the same elevation, from Lone Pine to the Poverty Hills, even though the floor of the valley rises to the north. It may be that the present day Owens River is not the base to which the Sierran streams grade; rather it may have been pluvial Lake Owens, whose maximum late-Pleistocene levels were stabilized near 1140 m elevation by alluvial fans south of Owens Valley. During interglacial periods, when the lake level dropped, base levels would also be lower, but during these times fan building beneath the Sierra Nevada appears to have been minimal.

RANGE FIRES

Range fire is an important geomorphic agent in the semiarid setting of Owens Valley, accelerating the physical weathering of granitic boulders, producing grus, and allowing increased eolian sediment transport by removing vegetation [*Bierman and Gillespie,* 1991a]. Our observations in addition to those of Blackwelder [1927], Birkeland [1984], Burke and Birkeland [1979], Gillespie [1987], and Evenson et al. [1990] suggest that fire is an important agent of rock weathering throughout the Great Basin.

Fire affects rock and geomorphic surfaces in several ways, some of which will be observed at stop 11. (1) Uneven heating and expansion and perhaps vaporization of endolithic moisture induce spalling, the loosening and removal of large pieces of the rock surface. (2) Removal of vegetation causes increased eolian transport of grus and may contribute to the smoothing of geomorphic surfaces through time. (3) Intense heating may interfere with the application of exposure age methodologies by removing rock varnish and accelerating the loss from rock of cosmogenic and radiogenic gases such as Ar, He, and Ne.

Fires that we and others have observed in Owens Valley, began at a point and spread as an irregular flame front. We estimate that the dwell time for rangefires in any one area is between 2 and 6 min. Areas of unburned vegetation remain within rangefire burns, implying that heating in these fires is spatially heterogeneous.

The temperature of boulders during range fires is not well constrained. Gillespie et al. [1989] provide an estimate of about 700 °C for the surface of a boulder exposed in a campfire. Zschaechner [1985] reported that maximum temperatures recorded by temperature-sensitive paint placed in sagebrush-fueled fires ranged from 540 to 980 °C. Observations by firefighters suggest that the most extreme temperatures occur in range fires during brief, intense fire storms, characterized by tornado-like winds vortexes. After passage of such fire-storms at night, exposed rocks are incandescent (T. Willhoite, 1985, personal commun.), implying temperatures in excess of 650 °C. Because heating duration is short only the surface of rocks significantly heated (Fig. 8).

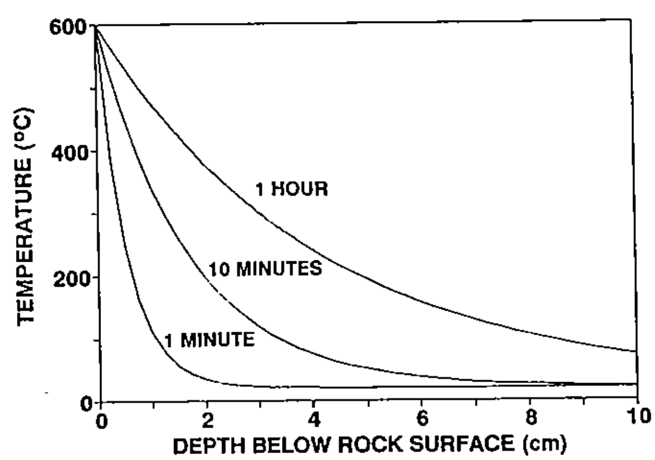


Fig. 8. Estimated depth-temperature profile of a rock heated by a 600° C range fire. Curves show temperature of rock at different depths below the surface after exposure to fire for 1 minute, 10 minutes, and 1 hour.

Fig. 9. Map showing location of field trip stops.

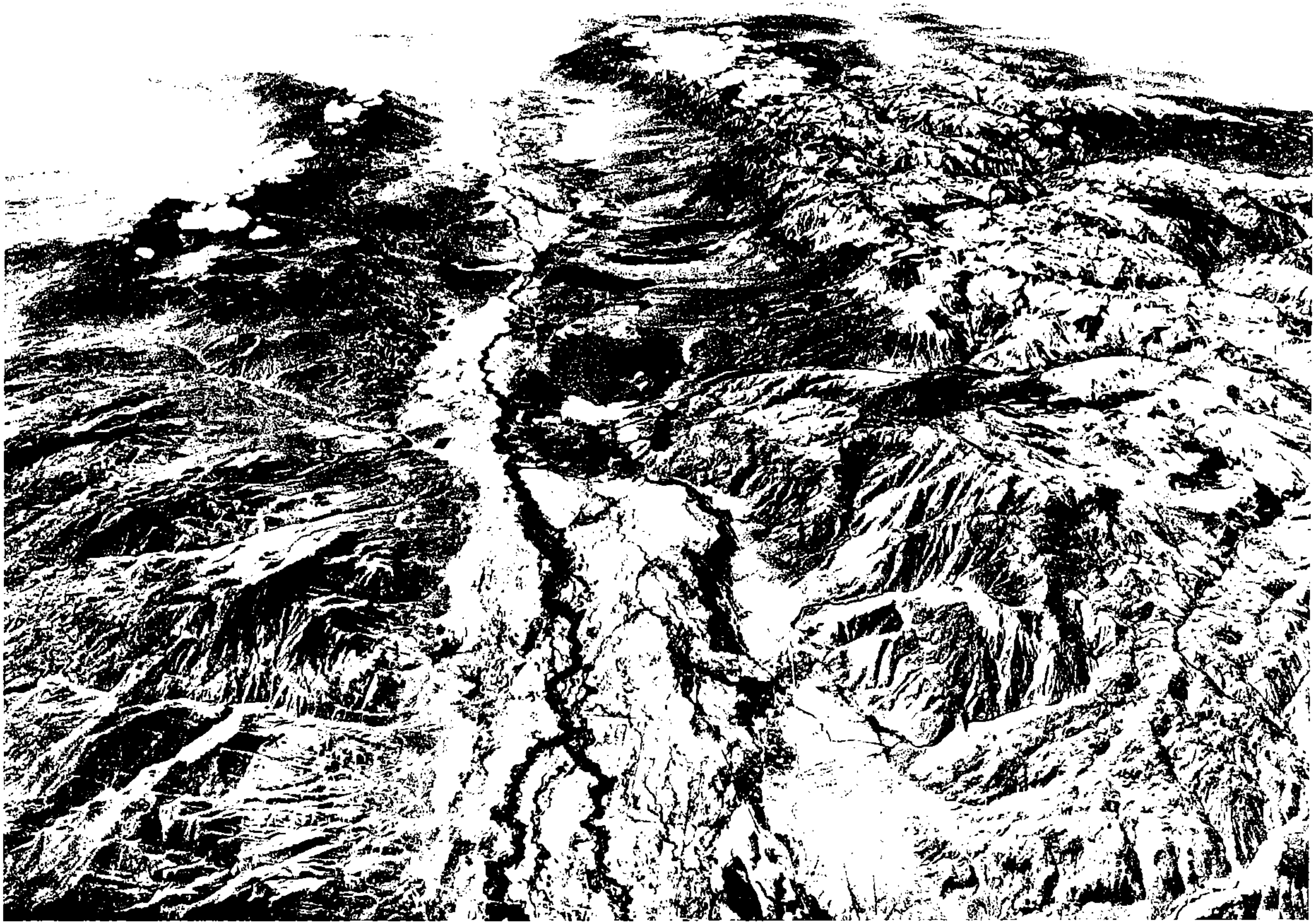


Fig. 10. High-altitude oblique aerial photograph looking south down Owens Valley from near Bishop. Sierra Nevada are to the right; the upland surface of the Coyote warp is on the lower right-hand side of the photograph. The Inyo Mountains are to the left. The Big Pine volcanic field is in the middle distance; Owens River and the nearby scarps of the Owens Valley fault zone in the middle of the valley lead south towards Owens Lake, near the top of the photograph. The gap in the Inyo Mountains east of the volcanic field is the Waucoba embayment.
(U. S. Air Force photo taken for U. S. Geological Survey, 018L-057).

SECTION 2: FIELD TRIP STOPS

STOP 1 - BOOTLEG DEBRIS FLOW ON HWY 395

Location. Stop at the pullout 0.5 mi south of the Bootleg campground, about 86 miles south of Reno on US 395.

Purpose. The small fan at this stop was built in a manner analogous to the debris-flow fans flanking Owens Valley. The immediate proximity of the source area and the fan at this stop affords an opportunity to view the hillslope-fan geomorphic system in its entirety. We will discuss debris-flow initiation, the mechanics of debris-flow motion, and the role of debris-flows in fan sedimentation.

Description. Several debris-flow tracks run down the steep, vegetated talus cone and hillsides which form the source area for this fan. The most recent debris-flow track swings across to the right behind the trees in the middle ground (Fig. 1-1). The road cut 100 meters ahead confirms that the small fan at this site was built by repeated debris-flow deposition. Here, as in the rugged drainages of the western slope of the White Mountains and in the *unglaciated* basins of the southern Sierra Nevada, debris-flows are usually initiated as failures of talus cones during intense summer thunderstorms. Hydrologic conditions at the time of failure and the granulometry of the source material determine the mechanical properties of the debris flow (see below). Debris-flow volumes often exceed the size of the initial failure because, on the steep slopes of talus cones, debris flows can be erosive, scouring material from the gully bed enroute to the fan.

A view up the small fan reveals that the debris flow followed a pre-existing channel across the fan, depositing material along channel margins and contributing to the accumulation of well-defined levees before spreading out at the distal end of the channel (Fig. 1-1). Note the matrix-supported structure of the debris-flow deposit and the abundance of large boulders and logs at the flow margin. In some instances, boulder accumulations at the flow front can significantly influence debris-flow behavior [*Pierson*, 1986]. We will see evidence of this effect in Owens Valley. At the deposit margin, over-ridden grasses appear to have been gently pushed aside, testifying to the non-erosive behavior of the debris flow on the moderate (~12 %) slope of the fan surface. Observations such as this are common on debris-flow fans and it is generally held that debris flows do not contribute to channel deepening on fans [*Pierson*, 1980; *Suwa and Okuda*, 1983].

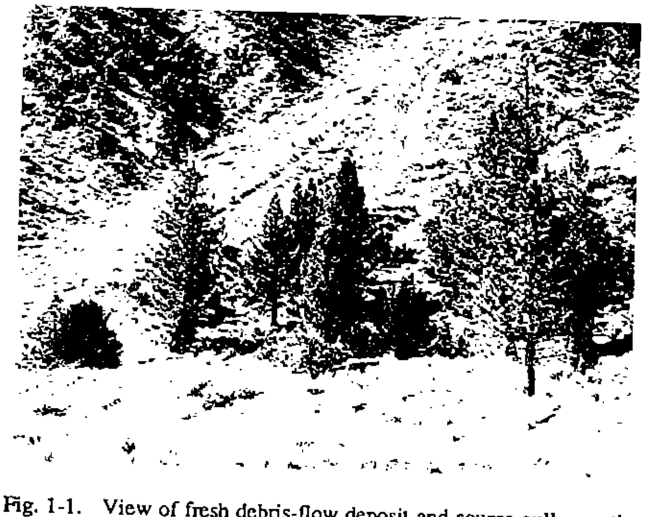


Fig. 1-1. View of fresh debris-flow deposit and source gully, south of Bootleg campground on US 395.

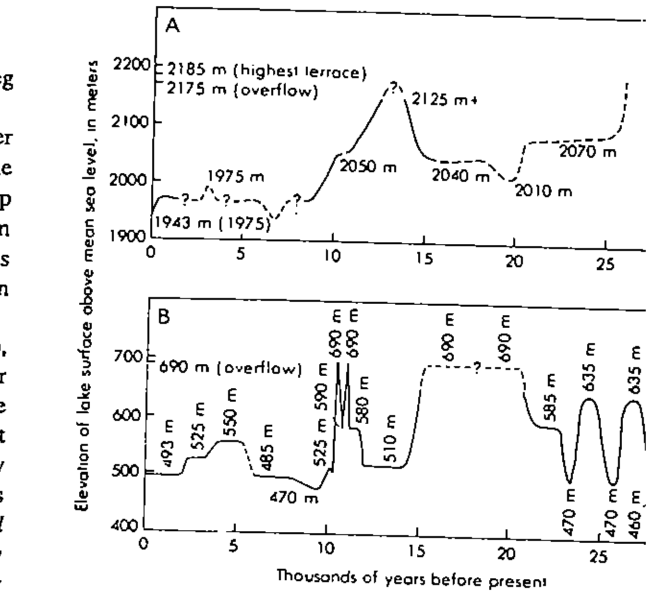


Fig. 2-1. Lake surface fluctuations of (A) Lake Russell in Mono (from Lajoie in Smith and Street-Perrott, 1983), and (B) Searles (from Smith and Street-Perrott, 1983) during the past 30 ka.

STOP 2 -- CONWAY SUMMIT OVERLOOK

Location. Pull out on the Mono Basin overlook, on US 395 south of Conway Summit, between the towns of Bridgeport Lee Vining.

Purpose. Overview of Mono Lake, Mono Basin, and glaciated Sierra Nevada.

Description. We have just climbed south out of Bridge Basin. The jagged Sawtooth range, and the Matterhorn, regarded as the northern limit of the "High Sierra." The view s from Conway Summit includes much of the central Sierra, from Tioga Pass region south to Mt. Morrison. The Sierra crest in area is typically 3600 - 3900 m elevation, although somewhat lo near Mammoth Mountain. The culmination of the Sierra is far south, between the latitudes of Big Pine and Lone Pine, wh many peaks exceed 4000 m elevation.

Below the lookout point lies Mono Basin, partly filled by M Lake. Mono Lake is one of the oldest continuously existing lake North America. Drill core data indicate that it has been extant si the Bishop Tuff was erupted (~0.73 Ma) and probably lon [*Scholl et al.*, 1967]. Throughout its history, the lake level fluctuated widely in response to climatic changes (Fig. 2-1). example, during the latter stages of the Tioga glaciation, the l may have reached a depth of ~283 m (compared to its modern p diversion maximum depth of ~45 m), and may have actually h overflowed its drainage divide for a short while [*Russell*, 18 *Lajoie*, 1968], flowing into Adobe Valley to the southeast a eventually reaching Owens Valley. On the northern shore of lake is Black Point, a large sublacustrine volcano that last erupted ~13 ka, when the lake level was just below its top.

Since diversion of the major rivers feeding the lake by the L Angeles Department of Water and Power in the 1940s, the lake le has lowered ~12 m. However, the lake level has fluctuat naturally throughout the Holocene, and on occasion it has been ev lower than today [*Smith and Street-Perrott*, 1983].

Glaciers from the Sierra Nevada descended to the edge of Mo Lake. The town of Lee Vining is built on the delta of Lee Vini

Fig. 2-2. Oblique aerial view looking south toward Mono Craters in the foreground with Long Valley and the Sierran crest in the background. Photo by Roland von Huene [in *Rinehart and Smith*, 1982]..

canyon. To the south are Bloody Canyon and Grant Canyon. K. Lajoie showed that Lake Russell entered between the Grant Lake moraines following the first stages of glacial withdrawal. It thus appears that pluvial and glacial maxima were not exactly coincident.

The islands in the lake are Paoha and Negit islands, made famous by Mark Twain during his sojourn at Lee Vining. Both islands apparently have been active during the 19th century [*Stine*, 1984, 1987]. South of Mono Lake are the Mono Craters - rhyolite domes erupted since ~36 ka (Fig. 2-2). The youngest, Panum Crater, is the closest to the south shore of the lake. It erupted only ~600 years ago. It will be our lunch stop.

It is worth a moment to inspect the road cut on the north and east sides of US 395, just behind the overlook. This exposure is thought to be diamicton - till or "outwash" - of Sherwin age. The Sherwin glaciation, which we shall discuss again, preceded the eruption of the Bishop Tuff by as much as 0.1 million years. Sherwin till is widely preserved in Bridgeport Basin, but it is scarce in Owens Valley. The presence of Sherwin till at Conway Summit shows that subsidence of Mono Basin largely postdated ~0.8 Ma.

STOP 3 -- PANUM CRATER

Location. From Lee Vining, travel south on US 395 roughly 5.5 miles and turn left (east) on Highway 120. Continue on this road for about 3 miles and turn left at marked dirt road leading to Panum Crater. Stop at the parking lot at the foot of the crater (about 1 mile).

Purpose. At this stop, we will break for lunch at the base of the southwestern rim of Panum Crater. This stop affords excellent views of many of the results of the geologic processes that have characterized the late Quaternary development of Mono Basin.

Description. To the south and southeast the skyline is dominated by the northern end of the Mono Craters, of which Panum Crater is the northernmost and youngest. To the southwest and west lies the Sierra Nevada range front, here defined by the Silver Lake fault, from which several important late Pleistocene morainal sequences emerge, including the Bloody Canyon/Sawmill Canyon sequence that we will visit next. Farther north, the range front steps to the east at Lee Vining Canyon before continuing northwestward along the Mono Lake fault. The base of this scarp, along which you have just driven, forms the western margin of Mono Lake before rising northward into the ridge west of Conway Summit.

Panum Crater defines the northern end of the north-trending arc of the Mono Craters volcanic complex, a string of primarily Holocene siliceous plug intrusions (Fig.2-2 and Fig. 3). Panum Crater was formed during the most recent eruption at the Mono Craters volcanic complex, between A.D. 1325 and 1365 [*Sieh and Bursik*, 1986]. The eruption closely preceded a similar eruption to the south in the Inyo Craters, which lie between the Mono Craters and the northwestern rim of the Long Valley caldera, suggesting a genetic relationship between these recent events [*Sieh and Bursik*, 1986]. Almost all of the other exposed domes in the Mono Craters chain are Tioga in age or younger [*Wood*, 1977; *Sieh and Bursik*, 1986]. However, lacustrine sediments show that rhyolitic volcanism chemically similar to the present domes has occurred in the basin since about 36 ka [*Lajoie and Robinson*, 1982]. These observations indicate that extensive volumes of early Mono Craters deposits have been buried by subsequent eruptions [*Sieh and Bursik*, 1986].

STOP 4 -- BLOODY CANYON MORAINES

Location. From Panum Crater, travel west to US 395 and turn north. After 0.4 miles, turn left (southwest) onto the Grant Lake - June Lake road. Sawmill and Bloody Canyons are visible to the west. Access is by dirt road that branches off southwest 1.3 miles from US 395, just where the paved road turns south. After 0.4 miles, turn right (northwest) onto a second dirt road and follow it about 0.9 miles to the breached moraines of Sawmill Canyon. Turn left and drive up Sawmill Canyon, between the moraines. The stop is about 1.2 miles up the road, near 7600 ft elevation.

Purpose. Bloody Canyon is the type area of the Mono Basin till, probably dating from oxygen-isotope stage 6. The complex of moraines is one of the most complete and best studied in the Sierra Nevada. Recently, Phillips et al. [1990] reported dates for several of these moraines based on analyses of ^{36}Cl created by cosmic radiation (Fig. 4-1). These dates contradict the stratigraphy deduced by field inspection, relative-weathering, and soil-development studies, and compel us to reevaluate the field setting and these techniques.

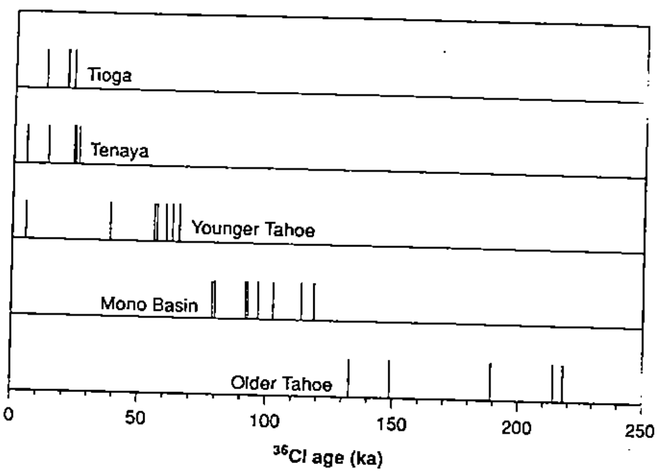


Fig. 4-1. Cosmogenic ^{36}Cl exposure ages for moraines of Bloody Canyon. Figure from Phillips et al. [1990].

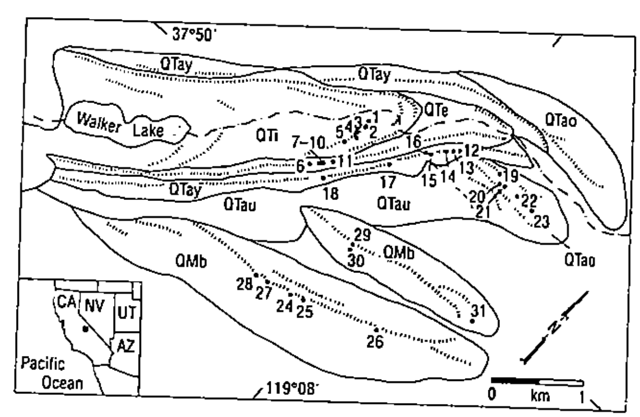


Fig. 4-2. Map of glacial deposits at Bloody Canyon, from Phillips et al. [1990], showing their ^{36}Cl sample locations. QTi, Tioga deposits; QTe, Tenaya deposits; QTay, younger Tahoe deposits; QTao, older Tahoe deposits; QTau, undifferentiated Tahoe deposits; and QMb, Mono Basin deposits. Hatchures indicate moraine crests.

Description. The moraine complex at Bloody/Sawmill Canyons is especially well preserved because of a geological "accident": the course of the glaciers shifted to the north more than 100,000 years ago (Fig. 4-2). Because of this, smaller moraines - and moraines that were faulted down east of the range-front faults [*Clark*, 1979] - were protected from burial by larger moraines deposited during subsequent glaciations. Although the extensive Sherwin moraines have evidently been faulted down sufficiently to have been buried, post-Sherwin and pre-Tahoe moraines of three different ages (rc' weathering data) have been preserved along Sawmill Canyon, the old, southern course of Walker Creek. The crests of the older moraines are shown in Fig. 4-2, but Phillips did not differentiate them from the largest moraine, the type locality for Mono Basin till.

After the course of the glaciers shifted northward, parallel to modern Walker Creek, at least four advances of glaciers left moraines. The largest of these was been assigned to the "Tahoe" glaciation by Sharp and Birman [1963], but Gillespie [1982] thought that the moraine was more complex, consisting of a young (stage 4?) moraine overtopping and older (stage 6?) one, which protruded from the outside slope of the right lateral moraine, above the crest of the Mono Basin moraine. What appears to be the same moraine, based on assessment of crest elevations and slopes and soil development, also protruded beyond the overlying till. The 36Cl data seem to indicate that the protruding snout is considerably older than the Mono Basin moraine (185 vs. 105 ka), and it is this implication that is controversial.

Birkeland and Burke [1988] speculated that ~10 m of till had been eroded from the crest of the steep-sided Mono Basin moraine. Because the cosmogenic isotopes are produced only within a meter or so of the exposed surface, rapid erosion could lead to erroneous apparent ages. On the other hand, the relative dating and soil studies that were the basis for earlier analyses are somewhat subjective, and don't give numerical ages. It is also possible that the stratigraphy was incorrectly deduced.

STOP 5 -- LONG VALLEY CALDERA

Location. Continue south on US 395. Drive about 11 miles south of Route 203 and turn north toward Crowley Lake and South Landing. Pull out several hundred meters north of US 395.

Purpose. This brief stop offers an overview of geologic features that have shaped the Quaternary history of Long Valley.

Description. Lake Crowley is in the foreground, while to the west the faulted end moraines of several large late Pleistocene glaciers emerge from the Wheeler crest of the Sierra Nevada, which forms the southwestern flank of the valley. To the north and west, Mammoth Mountain, a large late Pleistocene volcanic dome complex mantles the rim of the valley. In the middle distance, just to the east of US 395, is the large resurgent dome that formed after the collapse of the Long Valley caldera. Directly to the north, across the lake, Glass Mountain Ridge forms the rim of the caldera. Lake Crowley was impounded in 1941 by the Los Angeles Department of Water and Power (LADWP) as part of the Owens Valley/Mono Basin aqueduct system. There was a long, bitter fight between LADWP and the local ranchers during the early part of the century over water rights, summarized by Lipshie [1976] and Nadeau [1974].

During much of the middle and late Pleistocene, after the eruption of the Bishop Tuff, the Long Valley caldera was filled by a lake. Old shoreline terrace deposits occur at elevations up to 7600 ft (2320 m) in the central part of the caldera [*Lipshie*, 1976]. Later, lake levels steadily dropped and the lake finally disappeared about 0.1 Ma when the Owens River cut down to the level of the caldera floor. A delta, that was built into the lake by the Tahoe glacier at Convict Creek, has been offset by the Hilton Creek fault where US 395 takes a bend and drops to the south as it approaches Lake Crowley. This scarp exposes the terminal moraine and deltaic deposits.

STOP 6 -- SHERWIN TILL AND BISHOP TUFF

Location. Stop at the road cut on US 395 at the top of the Sherwin grade. Park in the turnout (to the right when southbound) immediately south of the intersection of old US 395 (lower Rock Creek Road) with the modern highway, just before starting down the long grade to the town of Bishop and Owens Valley. The road cut is across the freeway, which can be busy at times.

Purpose. We will inspect the Sherwin till and the overlying Bishop ash and Tuff.

Description. Before crossing the freeway, take a moment to glance south, down Owens Valley. The Sherwin grade is the flank of the old volcano that erupted the Bishop Tuff, which mantles this slope. The tuff is buried as much as 210 m below the alluvium south of Bishop. The High Sierra rise steeply from Owens Valley, on the west. The upland surface in the middle distance is the Coyote warp. The large basalt cinder cone in the valley itself is Crater Mountain, astride the Fish Springs horst. The White Mountains. rise to the left and, beyond Crater Mountain, the Inyo Mountains are their southward continuation.

Before the eruption of the Bishop Tuff, Sherwin till from the Sierra was deposited over the flank of the volcano. Reworked Sherwin till has been found east of here, beneath the tuff in the gorge of Owens River. At the road cut in front of you, the eroded surface of the till, draped by 3-5 m of airfall pumice, has been well exposed. Granitic boulders within the till are grusy but not oxidized; some can be disaggregated by hand. R. P. Sharp felt that this level of weathering, which took place prior to burial, indicated an age equivalent to the Tahoe till today - that is, the Sherwin till predates 0.73 Ma by ~50 to 150 ka.

The white pumice is bedded concordant with the eroded surface upon which it fell. Above is the pinkish tuff, horizontally bedded upon deposition (and discordant with the pumice). The maximum thickness of the tuff is ~240m. It was erupted in at least two episodes, indicated by two distinct cooling units [*Sheridan*, 1970, 1971]. These events appear to have been close in time, because no eolian or fluvial deposits and no soil have been found between the two units.

Protruding from the weathering road cut are thin vertical clastic dikes, filled with rounded granitic pebbles. Originally vertical, these dikes now dip ~80° to the north, the result of post-eruption tectonic tilting. Wahrhaftig [1965] suggested that the gravel filling the dikes was derived from Tahoe or older outwash deposited over the tuff.

STOP 7 -- FISH SPRINGS CINDER CONE

Location. Five miles south of Big Pine, turn south from modern US 395 onto the old highway. After 1.2 miles, a dirt road leads west to the prominent red cinder cone. Park at the eastern base of the cinder cone, in an old quarry.

Purpose. We will climb the cinder cone, discuss offset on the Owens Valley fault zone, and inspect faulted alluvial fans.

Fig. 7-1. Oblique aerial view west toward the Fish Spring strand of the Owens Valley fault zone. The scarp displaces alluvial fans of a range of mid to late Pleistocene ages and a ~0.3-Ma basaltic cinder cone. The base of the cinder cone is buried by the fan deposits. This relationship is exposed in the scarp.

Description. To casual observers, the Fish Springs and other cinder cones of the Big Pine volcanic field appear to be quite youthful - no more than ~50 ka old. Yet the Fish Springs cinder cone is much older. It was dated at ~0.31 Ma by ^{39}Ar-^{40}Ar analysis of partially degassed granitic xenoliths [*Martel et al.*, 1987].

The cinder cone was erupted along a strand of the Owens Valley Fault Zone. Over most of its length, the right-lateral oblique-slip fault zone trends NW-SE, but near Fish Springs it turns north. The cinder cone is displaced vertically 78 m by the fault, from which an average displacement rate of ~0.25 mm/year may be deduced (Fig. 7-1 and 7-2). This value is similar to the estimated long-term average of ~0.3 mm/year for the Owens valley fault zone near Big Pine [*Martel et al.*, 1987]. By careful geometric reconstruction of the faulted cinder cone, Martel [1984] found horizontal displacement here to be negligible.

Before much offset had taken place, the cinder cone was buried deeply by debris flows from Birch Creek. After abandonment, the alluvial surface was faulted 31 m. If the average displacement rate is correct, an age of ~124 ka is indicated. If aggradation really is synchronous with glaciation, this fan must date from late stage 6. The very existence of a faulted, unincised ancient surface is evidence of rapid aggradation, followed by a hiatus.

The surface of the stage-6 fan west of the fault scarp has been deeply weathered. Granitic boulders are grusy or deeply pitted, and eroded flush to the surface instead of standing out prominently, as on younger fans. A distinct soil profile has developed and is clearly visible in road cuts. Careful observation of the degree of boulder weathering on the fan surface reveals several distinct units (the exact number of which appears to depend on the enthusiasm and experience of the observer!). The older of these units are offset equally by the scarp. However, the youngest units are offset differentially. Some of these fans may be seen just north of the cinder cone. The oldest unit was draped over the scarp, building up a subsidiary fan below it that has been offset only ~7 m at its head. This fan may correspond to the Tahoe glaciation, or to a later stade. Inset within the fan, on its south side (closest to the cinder cone), is a younger terrace of bouldery levees - offset only 3.3 m and probably of Tioga age. Finally, the channel of the intermittent stream that built these fans has been offset ~1.3 m, evidently during the 1872 earthquake. The knick point has migrated only a few meters west of the scarp.

Ages for these fans can be estimated from the average slip rate of the fault. Alternatively, if they could be dated directly, it would be possible to estimate variability in the slip rate for the late Pleistocene. If the latest offset represents a "characteristic event," there have been about 60 earthquakes since construction of the cinder cone, with an average recurrence interval of ~5 ka. This value is in reasonable agreement with the with the estimate of Lubetkin and Clark [1988] near Lone Pine (Stop 13).

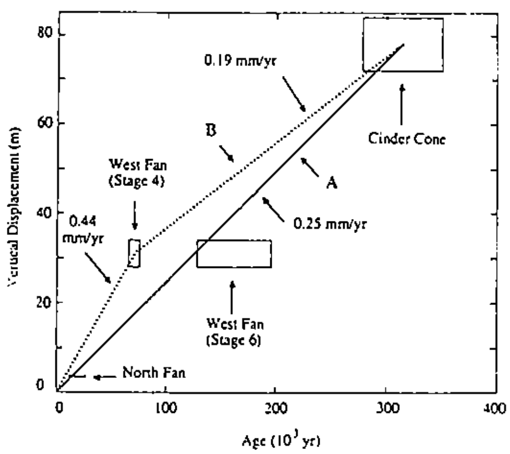


Fig. 7-2. Two possible displacement histories for the Fish Springs fault, based on different assumed ages for the "west" fan. The "west" fan is the older faulted fan west of the cinder cone, in which the vertical displacement of the scarp is highest (Fig. 7-1). The "north fan" is draped over the scarp just north of the cinder cone. The age of the undated west fan was assumed to be either oxygen stage 4 or 6. Subsequent studies favor the older age. From Martel et al. [1987].

STOP 8 -- WAUCOBI BEDS OVERLOOK

Location. Wildlife overview in the Poverty Hills, just north of Tinnemaha Reservoir off of US 395. The dirt road to the lookout is no longer signed (May, 1991). It leads to the left (east) up the low hill between the freeway and the reservoir.

Purpose. Overview of Plio-Pleistocene lacustrine sediments predating much of the subsidence of Owens valley.

Description. Late Pliocene Owens Valley was occupied in part by ancient Lake Waucobi [*Bachman*, 1978]. The lake sediments are visible to the east as whitish strata underlying darker fan deposits, in the sloping reentrant at the northern end of the Inyo Mountains. The lacustrine sediments are preserved in the Waucoba embayment, east of the town of Big Pine (Fig. 1). The upper Waucobi beds date to ~2 million years ago [*Sarna-Wojcicki et al.*, 1984]. They dip gently to the west, from ~2° for the upper strata to ~6° or more for the lower, and record the progressive westward tilting of the Inyo crustal block during the late Pliocene Epoch. The beds are in sharp contact with overlying alluvial deposits. Both are extensively faulted, and are truncated by the range-front faults [*dePolo*, 1989]. West of the range front, the lake beds are probably buried by late Pleistocene fans and valley fill.

Granitic sands and pebbles within the lake sediments suggest a western source in the granitic rocks of the Sierra Nevada. Perhaps the escarpment of the Sierra Nevada was farther east during the Pliocene than today. Other explanations, including ice rafting, have also been suggested.

From the tilting of the Waucobi strata it appears that Owens Valley deepened >600 m while the lake was extant, and >300 m subsequently. These numbers are approximate minima because the hinge points of the tilting block are uncertain, as is the western limit of the original lake beds. Exposures of the Waucobi beds are truncated along the range-front faults, and this zone may delimit the western side of the tilted block. The Waucobi beds do not provide an estimate for the deepening of the graben west of the tilted block.

From the wildlife overlook are visible three basaltic cinder cones and related flows on the flank of the Inyo Mountains, east of the reservoir. This eruptive center, part of the Big Pine volcanic field, is located at the southern end of the White Mountain fault zone [*dePolo*, 1989]. The oldest (and highest) cone dates from ~ 1 Ma; the youngest (northern) dates from ~0.2 Ma. Between the Poverty Hills and these cinder cones is the shallow bedrock sill separating the deep northern and southern grabens of Owens Valley, bordered on the west by the Owens Valley fault zone. The fault zone splays through the Poverty Hills, but essentially passes underneath the freeway to our west. Water wells drilled above the septum between the buried basins pass through multiple buried basalt flows from the Inyo Mountains. [*Hollett et al.*, 1989].

STOP 9 -- JEFFERY MINE CANYON FAN

Location. Proceed north from Bishop on Hwy 6 for 20 miles. Turn east on the White Mountain Ranch Road. Past the ranch, 0.6 mi ahead, turn left on the dirt road at the powerhouse. Follow this track for 0.2 mi and take the second right on a dirt road up the center of the Jeffery Mine Canyon (JMC) fan. Stop 0.8 mi up the fan, where the dirt road crosses the unvegetated deposit of the 1958 debris flow [*Hubert and Filipov*, 1989].

Purpose. At this stop we look at the surface morphology and stratigraphy of a fan typical of those derived from the metasedimentary and metavolcanic rocks exposed in this part of the White Mountains [*Crowder and Sheridan*, 1972]. We will discuss channel formation, debris-flow mechanics, and the evolution of fan surface morphology.

Fig. 9-1. Aerial photograph of Jeffery Mine Canyon fan (BLM CA01-77, #1-32-27). Light colored deposit is the 1958 debris-flow. Cross-section profile in Fig. 9-2 is indicated (section line) as is the location of the exposure pictured in Figure 9-3 (x).

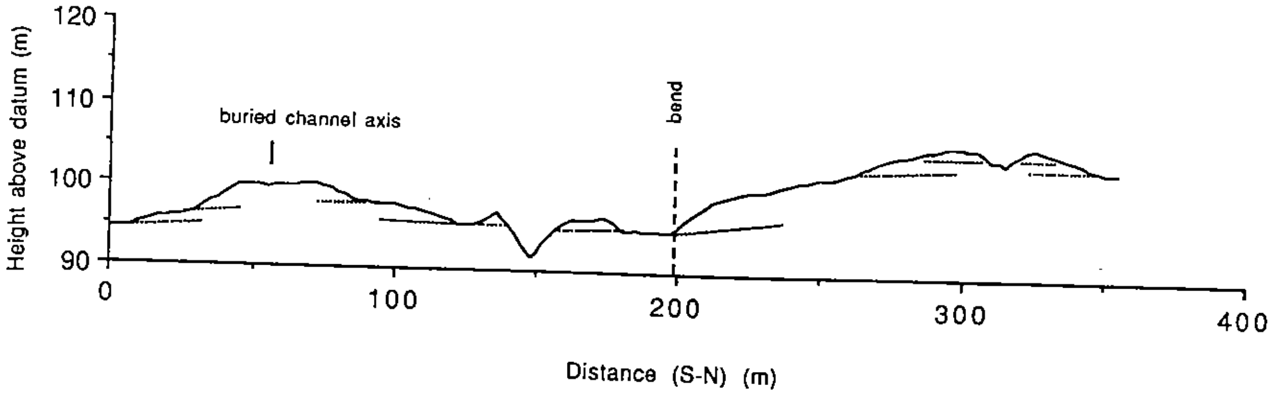


Fig. 9-2. Cross-fan topographic profile Dotted lines indicate the subsurface projection of on-lapping debris-flow contacts.

Description. The lower fan is characterized by a scatter of large boulders and the subtle lobes characteristic of deposition by highly mobile debris flows. Surface relief becomes more pronounced as we move up-fan where less mobile debris flows come to rest. At about 0.5 mi up the trail, a gully on the right-hand side has exposed a section of three thin, superposed debris-flow deposits. Note the sharp, non-erosional contacts between the flow units. Stop on the surface of the 1958 debris-flow deposit.

Surging behavior is commonly observed in flowing debris and 2 surges are readily distinguished on the surface of 1958 deposit and in channel wall exposures 100-200 m up fan. On-lapping snouts of the 2 surges allow estimation of deposit thicknesses and estimation of flow yield strengths (940-1200 Pa and 1100-1300 Pa, respectively -- a relatively mobile debris flow).

As we walk up the 1958 deposit towards the incised reach of the creek, observe the levees built by the 1958 flow. Over time, repeated in-channel deposition, overbank flow, and consequent levee aggradation will elevate the channel floor above the general level of the surrounding fan surface. On these fans, channels are usually re-excavated immediately by floodwaters following debris flows [*Beaty*, 1963; *Hooke*, 1987] and individual channel courses are relatively long-lived. Eventually, channels are plugged by massive debris flows and new channels are formed by diverted floodwaters. Linear topographic highs along paleo-channel traces dominate the surface morphology of these fans (Fig. 9-1, 9-2).

Stop at the first good exposure on the north wall of the channel. The fanglomerate is dominated by tabular deposits of matrix-rich debris flows. These beds are very poorly sorted and ungraded, with boulders and cobbles floating in a silt-and-clay rich (20-30 % by weight) matrix. The debris-flow beds range from 0.2 to 1.5 meters in thickness. Channel-filling debris flows are not always scoured away by subsequent floodwaters and buried channel gravels are occasionally preserved. The dark grey stratum, half way up the exposure, is a fluvial channel gravel bed preserved in this manner.

Dated detrital charcoal layers can provide an estimate of debris-flow recurrence intervals [*Hubert and Filipov*, 1989]. Proceed 100 m farther up fan to the 5-m-high exposure pictured in Figure 9-3 (again on the north wall). Charcoal (as yet undated) has been found at the base of the debris-flow bed immediately above the coarse, poorly sorted fluvial unit near the base of the section (Fig. 9-3). Seven debris-flow beds overlie the charcoal. On the basis of similar observations, Hubert and Filipov [1989] estimated a debris-flow recurrence interval of ~300 years (averaged from three ^{14}C dates).

Fig. 9-3. Stratigraphy of channel wall: incised portion of Jeffery Mine Canyon Creek. The scarp is 4.5 m high. Six debris-flow beds overlie the coarse channel gravel near the base. Detrital charcoal has been found in the debris-flow bed immediately above the channel gravel.

The surface of the JMC fan preserves a number of well-defined debris-flow lobes, many of which surround abandoned channels in the form of broad levees (Figs. 9-1 and Fig. 7). We will spend some time walking the surveyed transects (Fig. 9-2) and looking at abandoned channels, channel-plugging debris-flow lobes, and overbank debris-flow lobes.

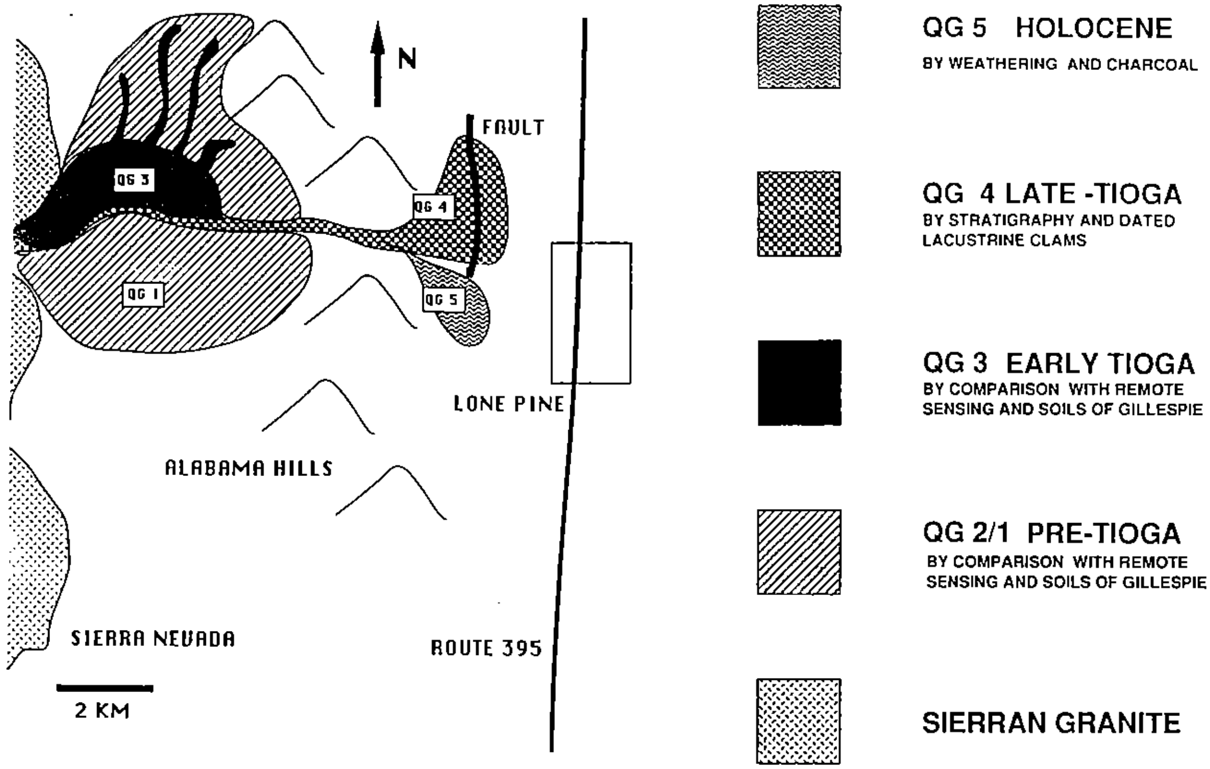


Fig. 10-1. Schematic map of age relationships of Lone Pine Creek alluvial fans.

STOP 10 -- LONE PINE CREEK ALLUVIAL FANS

Location. Drive 6 miles west from Lone Pine on the Whitney Portal Road. Ascend the Sierra front in a large switchback and park at a turn out where the road turns west into the canyon of Lone Pine Creek.

Purpose. This stop provides an overview of Owens Valley, Owens dry lake, the Alabama Hills and the Tuttle Creek and Lone Pine Creek fans.

Description.. West of the town of Lone Pine, along and to the north of the present course of Lone Pine Creek, is a complex of fan surfaces extending almost 10 km from the Sierra Nevada range front. Reconnaissance mapping indicates that these fan surfaces are of several different ages (Fig. 10-1). Older geomorphic surfaces are differentiated from younger surfaces by superposition, boulder weathering, and boulder frequency.

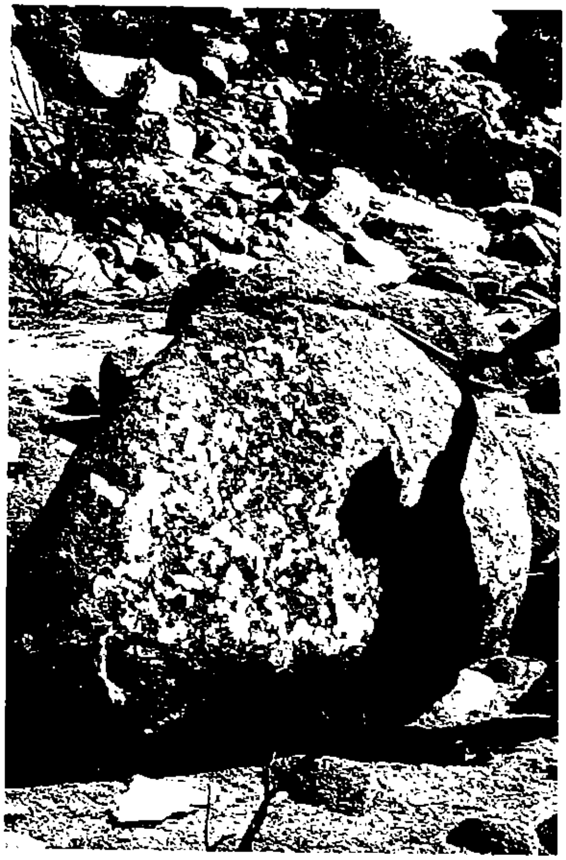
The oldest fan unit, Qg1, crops out near the Sierran range front and is well exposed in a road cut leading to the Forest Service campground and in the incised channel of Lone Pine Creek. Granodiorite boulders in these outcrops are severely weathered and most disintegrate with hammer blows. The surface of Qg1 is smooth. No channel morphology is preserved and the few boulders that crop out are extremely weathered. The age of this fan is unknown.

Fan unit Qg3 is the most widespread geomorphic surface near Lone Pine Creek. It is exposed at the top of the section along the incised channel of Lone Pine Creek and in road cuts along the Whitney Portal Road. This unit is characterized by well preserved and extensive channel morphology. Numerous but weathered granodiorite boulders crop out on the fan surface. Comparison of this fan with remote sensing data suggests that this surface may have been deposited during marine oxygen isotope stage 6.

Just west of Lone Pine and north of the Whitney Portal Road, fan surface Qg4 is exposed. The fan surface is characterized by well preserved channel morphology and relatively fresh granodiorite boulders, a few of which appear to preserve striations. The surface of this fan is offset by the Lone Pine fault, a spur of the Owens Valley fault system [*Lubetkin and Clark,* 1988]. No shorelines are present on this fan, suggesting that deposition on the fan continued after the level of Pleistocene Lake Owens dropped about 13 ka. Deposition on fan Qg 4 ceased abruptly after stream capture diverted Lone Pine Creek to the south (Fig. 13-1).

Fan Qg5 is exposed along the Whitney Portal Road, east of the Alabama Hills. The fan surface is characterized by numerous unweathered granodiorite boulders near the fan head and finer grain sediments near the fan toe. This fan postdates Qg4. Radiocarbon ages on charcoal from fluvial and debris flow strata in this fan indicate that deposition began at least 4 ka and continued through historic time.

Looking east you will see the Alabama Hills (Stop 12). This low range continues south along now-dry Owens Lake. Earlier in this century, a ferry plied the lake carrying silver ore from Cerro Gordo to the railroad at Cartago. Looking south along the range front you will see the scar of the 1989 Tuttle Creek fire (Stop 11).



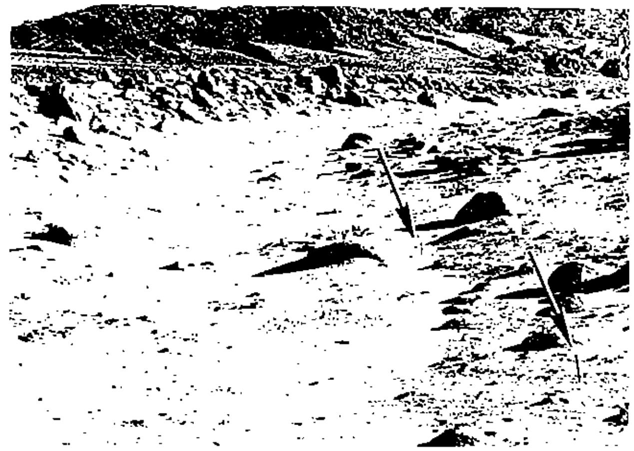


Fig. 11-2 (ABOVE). Wind ripples (1-2 cm high) formed in grus and charcoal (foreground) provide evidence of eolian transport in an abandoned debris-flow channel, Tuttle Creek fan. Arrows point to erosion pins. Photograph taken within 36 hours of the fire.

Fig. 11-1 (LEFT). Extensively spalled granodiorite boulder 36 hours after the rangefire at Tuttle Creek in 1989. Black areas are covered by soot from adjacent vegetation; light areas are spalled. Some spalls are lightly covered by soot, indicating that spalling occurred during the fire. Note large spall on ground below boulder.

Fig. 11-3. Oblique aerial photograph looking east down the Tuttle Creek fan. The scar from the 1989 rangefire is the light area in the foreground. Abondoned debris-flow channels and boulder levees are evident. The Alabama Hills are in the middle distance.

Fig. 11-4. Bouldery channel-margin levees and a bouldery debris-flow snout plugging an abandoned channel on the Tuttle Creek fan shown in Fig. 11-3. Relief on snout is >2 meters. View is up-fan.

STOP 11 -- SURFACE MORPHOLOGY AND EFFECT OF RANGE FIRE ON TUTTLE CREEK FAN

Location. Drive about 4 miles west of Lone Pine on the Whitney Portal Road. Turn south on Olivias Ranch Road and drive about 1.5 miles until you reach the Tuttle Creek burn. The burn and the ranch road extend to the fan head at which point the road becomes impassable.

Purpose. We will do two things at this stop: 1. Investigate the effect of range fire on fan surfaces and granitic boulders and 2. Contrast depositional styles on, and the surface morphology of, the fans along the Sierran front (glaciated, granitic source rocks, inactive at present) to those we looked at yesterday along the White Mountains (unglaciated, metasedimentary and metavolcanic source rocks, presently active).

Description . We observed the Tuttle Creek fire (3.2 km^2, 30 March 1989) as it burned over bouldery, scrub-covered alluvial fan surfaces and range-front talus cones. Winds during the Tuttle Creek burn ranged from 5 to 10 m/s (12-25 mph) with higher gusts. The flame front was narrow and varied in height as wind speed and direction changed. On the basis of the size and duration of the Tuttle Creek fire, dwell time for the fire at one area was 2 to 4 min, although the presence of charred piñon stumps indicates that heating was more intense and of longer duration in the vicinity of individual trees.

We examined granitic boulders affected by the Tuttle Creek range fire and observed that, on some of the boulders, nearly half of the [illegible] surface had spalled (Fig. 11-1). At Tuttle Creek, less than 5% of the boulders in the burn area spalled; however, boulders adjacent to the charred remains of trees and large bushes were heavily darkened by soot and more commonly damaged by spalling than boulders isolated from vegetation. This observation indicates that the extended and perhaps more intense heating caused by the presence of burning trees and sagebrush bushes increased the probability that a particular boulder would spall.

Our observations suggest that most granodiorite spalls are between 0.5 and 3 cm thick, cover an area between 50 and 200 cm^2, and shatter into grus or smaller pieces of rock. Photographs, taken during the two years since the burn occurred, suggest that most spalling occurs during or immediately after the fire.

Within 24 hr after the Tuttle Creek fire had been extinguished, grus and charcoal had been moved by the wind and shaped into bedforms (Fig. 11-2). Similar effects were noted in 1988 by K. Whipple at the 1985 Symmes Creek burn. Although the surface at Tuttle Creek was devegetated, erosion pins indicated little net movement of sediment during the six months after the Tuttle Creek fire, despite unusual heavy thunderstorms during August 1989.

There are a variety of landforms on this fan surface: abandoned channels, bouldery channel-margin levees, and bouldery debris-flow snouts. Compared to the White Mountain fans, the fans below the glaciated Sierra Nevada have abundant large boulders, more closely-spaced abandoned channels, narrow boulder berms instead of broad channel-margin levees, channels inset into the fan surface rather than perched on topographic highs, and narrow, elongate overbank lobes (appear as boulder-laced snouts) rather than wide lobes forming broad terraces (Figs. 11-3 and 11-4).

STOP 12 -- ALABAMA HILLS PEDIMENTS

Location. Drive east toward Lone Pine on the Whitney Portal Road. Turn north on Movie Flats road which wanders its way north through the Hills finally rejoining US 395 several miles north of Lone Pine. Today we will stop for lunch at the base of a large inselberg about 2 miles north of the Whitney Portal Road.

Purpose. This stop will provide sustenance and allow closer examination of the landforms and weathering environment of the Alabama Hills.

Description. The Alabama Hills are a low range rising several hundred meters between the Owens River and the Sierra Nevada. Within the Alabama Hills, both fine-grain metavolcanic rocks and coarse-grain granodiorite crop out. The granodiorite has been dated at 85 Ma (U/Pb by *Chen and Tilton*, 1991) and was therefore emplaced contemporaneously with the granites of the Sierra Nevada. The Alabama Hills have presumably been faulted down from the adjacent range crest. Granodiorite of the Alabama Hills weathers to a rusty yellow/orange color and so can be readily distinguished from the gray granite of the Sierra Nevada.

On the western flank of the Alabama Hills, in areas where the granodiorite crops out, are well developed inselbergs, tors, and pediments. These features are, in places, well coated with rock varnish, and their morphology reflects the influence of pervasive and consistent jointing of the granodiorite. The large-scale pattern of light and dark exposures you see reflects primarily the amount of rock varnish on a particular rock surface. In contrast to varnished surfaces which are frequently case hardened to a depth of several cm, unvarnished rock surfaces are frequently grusy and crumble under a hammer blow. In many areas, it appears that these deeply weathered and now unvarnished rock surfaces were previously covered by sediment which has now been removed. Initial mapping suggests that stripping of these surfaces may have been in response to uplift along the easternmost margin of the Alabama Hills. This uplift lowered the effective base level of streams draining the Alabama Hills, initiated incision, and resulted in the partially denuded landscape before you.

The Alabama Hills are surrounded, and in places, overrun by granitic alluvium shed from the Sierra Nevada. On the basis of relative weathering criteria, we have mapped Sierra Nevada alluvium of at least three different ages.within the Alabama Hills. In several places, boulders of Sierran provenance lie as isolated erratics on Alabama Hills pediments. This implies that pediment surfaces

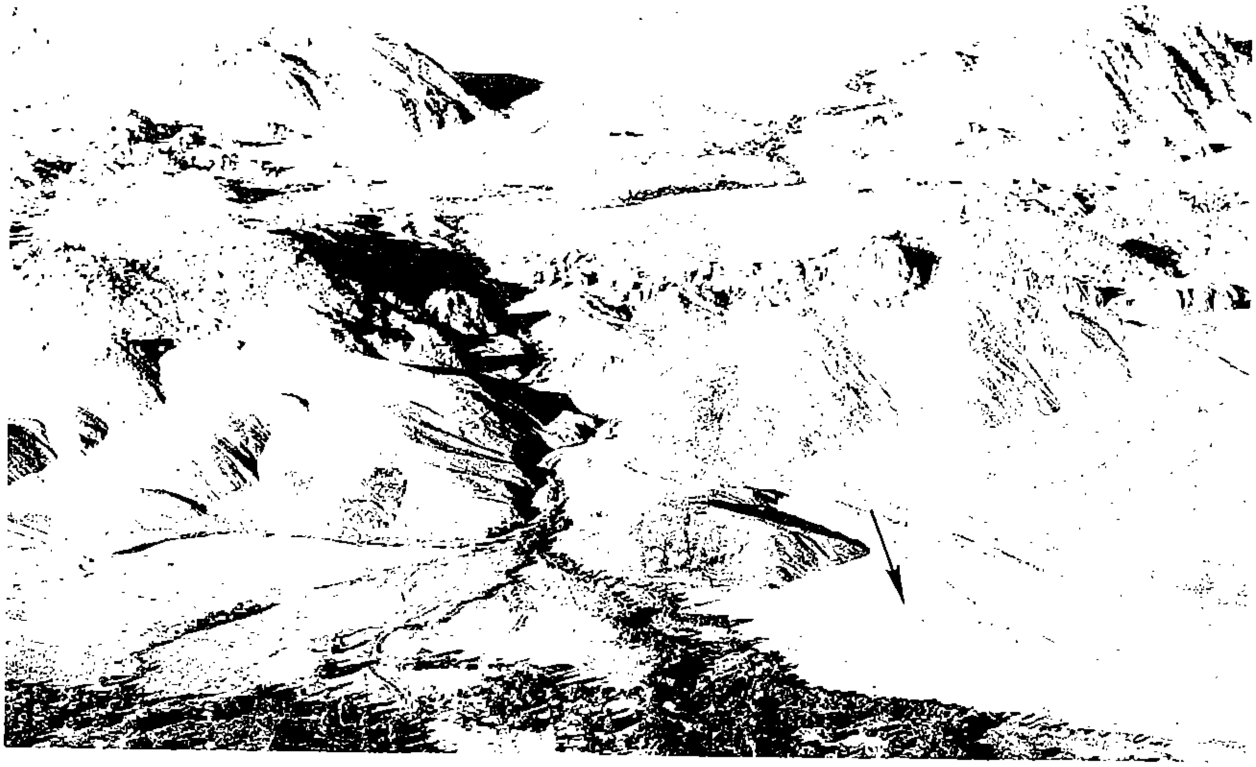


Fig. 13-1. Oblique aerial photograph looking west over Lone Pine Creek and the Alabama Hills. The scarp of the Lone Pine fault cuts fan surface Qg4 in the lower right-hand corner of the photograph. The large black arrow points to the scarp and the boulder shown in Fig. 13-2. The Qg5 fan south of the modern course of Lone Pine Creek is Holocene in age and was deposited after stream capture and incision during the Tioga glaciation. The Mount Whitney Portal road runs through the Alabama Hills just to the left (south) of the beheaded Qg4 fan.

were extant prior to alluviation and that these surfaces have subsequently been stripped of their alluvial cover.

In addition to field mapping, we have been studying the geomorphic evolution of the Alabama Hills by measuring the concentration of cosmogenic isotopes produced *in situ*. These isotope measurements will allow us to estimate the rate at which specific granitic surfaces with the Alabama Hills have eroded and may provide crude limiting ages for the alluvial boulders overlying the pediment surfaces.

STOP 13 -- LONE PINE FAULT SCARP

Location. Drive 1 mile west of Lone Pine on the Whitney Portal Road. Park on the north side of the road at an abandoned turnout with a sign for the Alabama Hills Recreation Area. Walk about 0.5 mile north from the Holocene fan, on which you are parked, to the Pleistocene fan which has been displaced by the fault.

Purpose. At this stop we will examine the Lone Pine fault scarp, observe the condition of granitic boulders on a fan of latest Pleistocene age, and view the scarp boulder on which we have begun to test the rock varnish cation ration dating method of Dorn, [1983].

Description. About a kilometer west of Lone Pine, the Qg 4 alluvial fan surface is cut by a fault scarp with up to 6 m of vertical displacement (Fig. 10-1 and 13-1). The scarp is well preserved in bouldery diamicton and has been studied in detail by Lubetkin and Clark [1988]. On the basis of scarp morphology, rock varnish on scarp boulders, and trenching, they suggest that the scarp represents three events including the 1872 quake. Lubetkin and Clark [1987] provide a detailed field description of the fault.

During the spring of 1989, we opened two additional trenches in the graben east of the fault. On the basis of stratigraphy in these trenches, primarily buried vesicular A horizons, we believe that either three or four events are preserved in a graben adjacent to the fault. We are measuring the thermoluminescence of 12 samples collected from these trenches in an attempt to constrain better the age of faulting events.

This fault scarp, because it exposes large boulders and because it records several episodes of movement, provides an excellent opportunity to determine the chemical variability of rock varnish and to test the assertion that rock varnish chemistry changes progressively with time. It was first suggested by Lubetkin and Clark [1988] that one large scarp boulder preserved varnish of two different and distinct ages (Fig. 13-2). They hypothesized that the younger (lower) varnish began to form after a faulting event exposed this portion of the rock. We agree with Lubetkin and Clark's [1988] interpretation and so collected 44 samples of varnish from this boulder (Fig. 13-2). We have chosen to analyze samples from this boulder because it provides a particularly well controlled experiment examining the effect of exposure time on varnish chemistry as substrate and exposure geometry are held constant.

Analysis of these 44 rock varnish samples using SEM/EDS, according to the protocol presented in Bierman and Gillespie [1991b] and Bierman and Kuehner [1991], suggests that varnish chemistry differs between the older and younger varnish on the fault scarp boulder. If chemical data are analyzed using confidence intervals [see *Bierman et al.*, 1991] the older varnish has higher Fe, lower Ca, and lower K than the younger varnish; the differences are significant at greater than 95% probability (Fig. 13-3).

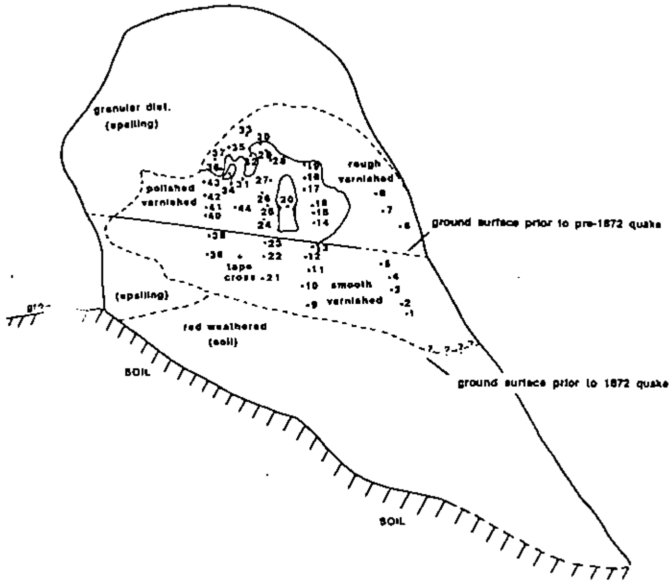


Fig. 13-2. South face of boulder on scarp of Lone Pine fault, described by Lubetkin and Clark [1987, 1988]. Polished and rough varnished surfaces are of similar age. Smoothed varnished surface was exposed after a pre-1872 earthquake and must be younger than both the rough and polished surfaces. Numbers refer to varnish sample sites.

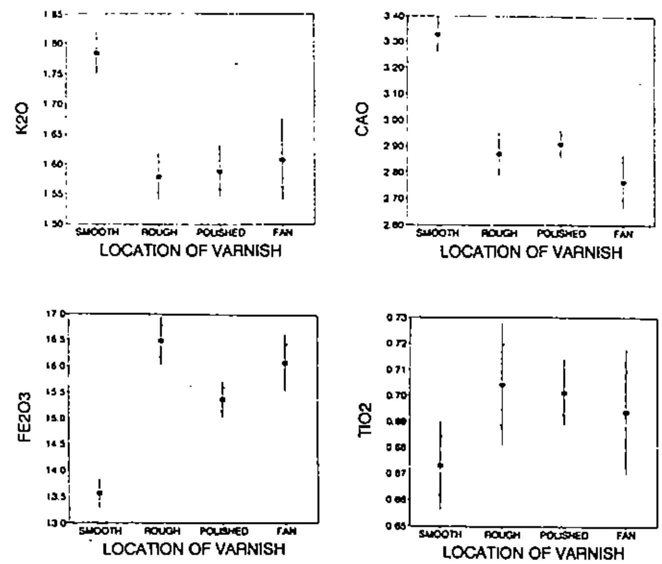


Fig. 13-3. Elemental abundances in rock varnish from boulder on Lone Pine Scarp (Fig. 13-2). "Fan" refers to varnish on boulders from fan surface Qg4, broken by the fault. Varnish from "fan," "rough" and "polished"surfaces must be older than varnish from the "smooth" surface. Bars are ±95% confidence limits.

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