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Proceedings of the Workshop "Ongoing Paleoclimatic Studies in the Northern Great Basin," Reno, Nevada, May 1993

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Edited by Larry V. Benson

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PREFACE

The workshop "Ongoing Paleoclimatic Studies in the Northern Great Basin" was jointly sponsored by the U.S. Geological Survey (USGS), the University of Nevada-Reno (UNR), and the Quaternary Sciences Center (QSC) of the Desert Research Institute, Reno, Nevada. W. Berry Lyons, chairman of the Hydrology Program (UNR), and Dale Ritter, head of the QSC, chaired the presentations. On May 16, Joseph Smoot, of the USGS, led a field trip along the Truckee River Canyon downstream from Wadsworth, Nevada. The subject of the field trip was the Quaternary-Holocene sedimentary record exposed in the Truckee River Canyon. On May 19, Larry Benson of the USGS led a field trip on the origin, order of formation, and age of carbonates deposited in the Pyramid Lake area. Joseph Liddicoat of Columbia University, Andrei Sarna-Wojcicki of the USGS, Peter Wigand of the QSC, and James Yount of the USGS led discussions in the field. W. Berry Lyons arranged for lunches, field-trip transportation, and conference facilities.

Larry V. Benson, Editor

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CONVERSION FACTORS

Multiply	*By*	*To obtain*
centimeter (cm)	0.394	inch
cubic centimeter (cm^3)	0.0610	cubic inches
gram (g)	0.00220	pounds
kilometer (km)	0.621	miles
liter (L)	1.06	liquid quart
meter (m)	3.28	feet
microgram (µg)	2.20×10^{-9}	pounds
micrometer (µm)	3.94×10^{-5}	inch
millimeter (mm)	0.0394	inch
square kilometer (km^2)	0.386	square mile

Temperature in degrees Celsius (°C) may be converted to degrees Fahrenheit (°F) as follows:

$$°F=1.8\ °C+32$$

Temperature in degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) as follows:

$$°C=(°F-32)/1.8$$

Meeting Highlights

Larry V. Benson, U.S. Geological Survey

Introduction

One of the major purposes of paleoclimatic and paleohydrologic studies is the determination of the degree of synchroneity of climate change recorded in marine and terrestrial sediments. A meeting was held in Reno, Nevada, on May 16–19, 1993, to exchange information and discuss the progress of ongoing paleoclimatic studies of one terrestrial area, the northern Great Basin of the Western United States. Several research topics were discussed including: methods of age estimation, glacial-lacustrine associations, high-frequency variations in lake level, and the frequency and duration of drought in the Holocene period.

Primary Methods of Age Estimation

Primary methods of age estimation and their application to the sediments (chemical and detrital) were discussed by G.W. Berger, C.A. Bush, and B.J. Szabo. G.W. Berger reported that fine-silt-size rhyodacitic glass shards in ash beds are suitable for thermoluminescence (TL) dating, using the additive-dose procedure originally developed for fired pottery. He also reported excellent results for a suite of independently dated tephra using newly developed sample-purification procedures. Rhyodacite glasses can be accurately dated by TL to 450,000 years before present (yr B.P.). Present in this context refers to 1950.

In a pilot study of five sections of sediment core from Walker and Pyramid Lakes, Nevada, C.W. Bush demonstrated that closed systems (with respect to uranium and its daughters) are sometimes maintained within some sections of sediment. Uranium-series methods, therefore, can sometimes be used to estimate the age of lake sediments deposited outside the age range of the radiocarbon (^{14}C) method.

B.J. Szabo presented results from a study of the tufa mounds that border Pyramid Lake indicating that uranium-series methods can be used to approximate the ages of such tufa deposits. In the Pyramid Lake Basin, tufas less than 50,000 years old contain large quantities of excess thorium, and the error in age estimates made using uranium-series methods is not small enough to confirm results from ^{14}C determinations. Carbonate cements formed between about 400,000 and about 60,000 yr B.P. at the Marble Bluff locality contain small amounts of excess thorium and seem to yield reliable uranium-series ages. The source of the excess thorium present in the Pyramid Lake Basin between about 60,000 and about 50,000 yr B.P. has not been determined. The presence of large quantities of uranium and thorium in tubular structures present at the base of tufa mounds indicates that thorium may have been transported in ground water to the lake as a colloidal suspension.

Secondary Methods of Age Estimation

Correlation within and between lake basins is highly desirable. Tephrochronologic and paleomagnetic methods show great promise in providing a means of achieving this goal. A.M. Sarna-Wojcicki provided a comprehensive overview of the field of tephrochronology. He reported that much of the Great Basin is downwind of two major volcanic areas that have been sources of sporadic, but prolific, explosive volcanic activity throughout middle and late Quaternary time. These volcanic source areas are the Cascade Range of Washington, Oregon, and northern California and the Long Valley-Mono Craters area of east-central California. Isotopic and other numerical ages determined on widespread ash layers from these source areas provide local and regional age control, whereas chemical and petrographic characterization of the layers allows unambiguous temporal correlation of associated sediments. Tephrochronology is currently (1993) being applied to sediments of Fish Lake Valley, Fort Rock Basin, Lake Bonneville, Lake Chewaucan, Owens Lake, Pyramid Lake, Walker Lake, and Washoe Lake.

Several researchers (D.B. Erbes, J.M. Glen, J.C. Liddicoat, S.P. Lund, R.M. Negrini, and J.G. Rosenbaum) gave presentations on the applicability of paleomagnetic methods to dating and correlation of stratigraphic sequences in lake basins. As discussed by R.M. Negrini, paleomagnetic secular variation (the low-amplitude variation in the intensity and direction of the Earth's magnetic field) offers a high-resolution tool that can be used for intrabasinal and interbasinal correlation. This method could be used to correlate nonmarine stratigraphies with the marine chronology.

J.C. Liddicoat discussed polarity reversals and other unusual field behaviors that serve as stratigraphic markers. Two such markers are a pole path that loops through the South Pacific and is recorded in the sediment on Paoha Island in the Mono Lake Basin and in Long Valley, California. This marker has been assigned an age of about 150,000 yr B.P. Another marker is the Mono Lake excursion present in sediments from the Lahontan and Mono Lake Basins, which has been assigned an age of about 27,000 ± 1,000 yr B.P.

J.G. Rosenbaum discussed the results of a magnetic study of sediments from Buck Lake, Oregon. One of the results of this study was the demonstration that magnetic susceptibility and pollen zonations closely corresponded with relatively warm climates corresponding to low-susceptibility intervals. In contrast, high-magnetic susceptibility characterized times when sagebrush steppe dominated the landscape.

Glacial and Lacustrine Records

In his discussion of marine-terrestrial correlation, M.J. Pavich indicated that climate-forcing functions of regular periodicity did not account for aperiodic terrestrial records that are climatically related. From a review of the literature, Pavich reported that there are hints of similar frequencies of change in some marine and terrestrial records, but the coherency of change has yet to be determined.

A. Bach, in his discussion of the glacial history of the eastern Sierra Nevada, concluded that at least seven major episodes of glaciation occurred at Bishop Creek during the past 200,000 years. Surface-exposure dating of these deposits indicates that the timing of the advances may be correlative with intervals of maximum ice sheet volume inferred from marine oxygen-18 ($\delta^{18}O$) records. Bach suggested that the relatively small number of Sierran glaciations compared to the number of ice-sheet maxima that occurred in the past 200,000 years was due to obliterative overlap of older moraines by younger moraines and by the limited temporal resolution of relative dating methods.

J.L. Bischoff, K.M. Menking, and G.I. Smith discussed ongoing studies of sediment core OL-92 taken from the Owens Lake Basin, California. Bischoff reported that total inorganic carbon (TIC) and total organic carbon (TOC) indicate down-core variation. Both parameters have minima between 25,000 and 15,000 yr B.P., corresponding to maximum glacial conditions. Older minima in TIC and TOC indicate older examples of extreme glacial conditions ranging back to 760,000 yr B.P. Bischoff concluded that during interglacials, Owens Lake was saline, alkaline, and highly productive and that, during interglacials, the lake was overflowing with fresh water and was relatively unproductive.

G.I. Smith and K.M. Menking, in their discussions of the sedimentology and mineralogy of core OL-92, indicated that poorly sorted coarse-grained sediments predominate between 195 m and the base of the core (323 m). They concluded that sediments from the lower part of the core (more than 490,000 years old) may represent shallow lake or fluvial conditions. Fine-grained sediments recovered in the top 195 m of core OL-92 were interpreted to indicate deep-water conditions. The presence of quartz, K-feldspar, and plagioclase feldspar in the clay-sized sediment fraction was interpreted by these authors to result from abrasion of Sierra Nevadan bedrock during times of glaciation.

Holocene Warm/Dry Period

The existence and timing of an especially warm/dry period during the Holocene have been argued for many years. Six speakers (E.M. Hattori, R. Karlin, P.F. McDowell, M. Rose, G.I. Smith, and P.E. Wigand) presented evidence for such an event. G.I. Smith suggested that the presence of a 4.2-m-thick oolite bed in the top of the Owens Lake core OL-92 indicates that the climate of the past 9,000 years was more arid than any time in the past 800,000 years. Based on a study of tree stumps submerged in the Lake Tahoe area, Rose argued that a large-scale change in climate caused simultaneous recessions of lake levels between 6,300 and 5,000 yr B.P. in the northern Sierra Nevada. Winnemucca Basin, which received water when Pyramid Lake spilled, seems to have been dry or nearly dry between 9,500 and 8,400 yr B.P. and between 7,800 and 4,000 yr B.P., according to archaeological data presented by E.M. Hattori. P.E. Wigand reported that the climate was too dry for woodrats to flourish on the western edge of the Lahontan Basin between 8,000 and 4,000 yr B.P., also supporting Hattori's data. Based on a summary of data from eastern Oregon, P.F. McDowell concluded that eolian sands dating to equal or less than 6,800 yr B.P. cover lake deposits in the Fort Rock Basin and that dunes in the Harney Lake Basin began to form just after 5,000 yr B.P. The pollen record from Diamond Pond also indicates that dry conditions existed between 6,000 and 5,400 yr B.P.

ABSTRACTS OF PRESENTATIONS

Long Quaternary Records from the Northwestern Basin and Range Province, California and Oregon

David P. Adam, U.S. Geological Survey

The Great Basin of the Western United States is the result of an extensional tectonic regime that has profoundly affected the physiography and drainage history of the region since Miocene time. Many existing and former sedimentary basins preserve long-duration records of past events. Plate-tectonic effects along the western margin of North America resulted in extensive volcanism in the Sierra Nevada and in the Cascade Range, and the passage of the Yellowstone Hot Spot eastward has produced more volcanism. This extensive eruptive activity produced many identifiable tephra layers that can be used to correlate among basins.

The northwestern Great Basin is particularly well suited to long-term stratigraphic studies because of the proximity of the Cascade Range (many tephra) and the relatively recent inception of the extensional tectonic regime (most sedimentary basins have not yet been breached). Ongoing studies by the U.S. Geological Survey have recovered numerous sediment cores from the Upper Klamath Basin and vicinity in northern California and south-central Oregon (table 1). These cores span intervals ranging in age from more than 50,000 yr (Wocus Marsh, Oregon, and Grass Lake, California) to more than 3,000,000 yr (Tule Lake, California). Vegetation at the sites now ranges from fir forests at the crest of the Cascade Range (Buck Lake, Oregon) to sagebrush steppe in the rain shadow to the east (Summer Lake, Oregon, and Tule Lake). Basin sizes range from about 1.5 km across (Grass Lake) to as much as tens of kilometers. Correlations between cores are based primarily on tephra and paleomagnetic polarity changes. Not all records are continuous. However, when many cores

Table 1. General summary of results from core sites in the Upper Klamath Basin and vicinity that were sampled during 1991–92

Site name	Core length (meters)	Recovery	Description
Butte Valley, California	102	Excellent	Both perennial and ephemeral lake deposits; numerous tephra; carbonate often preserved.
Grass Lake, California	29	Excellent	Basin dammed by lava flow; lake and marsh deposits at top of core, changing to volcanic debris flow at base; numerous tephra.
Buck Lake, Oregon	41.3	Excellent	Top 30 meters, lacustrine clay containing numerous tephra; bottom11 meters, volcanic debris flow underlain by basalt.
Round Lake, Oregon	50	Good	Top 25 meters, very tight clays that contain no visible tephra; probably representing intermittent lake.
		Excellent	Bottom 25 meters, open-water lake muds containing several tephra.
Wocus Marsh, Oregon	52	Excellent	Interbedded peats, organic muds, inorganic muds (basaltic rock flour?), and tephra. Probably includes glacial sediment from Mount Mazama.
Caledonia Marsh, Oregon	15.3	Good	Very soupy lacustrine muds containing several tephra. Drilling stopped by artesian conditions. Upper 12 meters sampled with good recovery using manual equipment in 1992.

are available from varying depositional environments in a relatively small area, hiatuses in one core may be bridged by the record in another core, and the hiatuses may be interpretable. Multiple cores also offer the opportunity to distinguish local (for example, tectonic) from regional (for example, climatic) changes.

Ongoing work on these cores includes detailed work on magnetic properties, pollen, diatoms, ostracods, and geochemistry. Combined pollen and magnetic work on the Buck Lake record indicates that magnetic susceptibility may function as an easily accessible proxy record of climate in this region. Pollen work combined with tephrochronologic time control indicates that the Buck Lake core includes a continuous climatic record that extends from the end of deep-sea oxygen-isotope Stage 9 or 11 back through most or all of Stage 12 or 14. The Butte Valley core record extends back more than a million years; the Brunhes-Matuyama boundary is present at a depth of 65 m in the 102-m core. Pollen and ostracod studies of that core are nearly complete.

Multiple long core records can only be retrieved from regions in which suitable deposits are widely distributed. The Basin and Range province, long characterized by an extensional tectonic regime, is one of the best places in the world to develop this multiple-core approach; suitable basins are present over a wide range of latitudes, climates, and ecological settings, and climatic changes have caused spectacular alterations in the landscapes of the Great Basin through time. Several long records are already available; these have provided key insights into the climatic evolution of the Earth, but such records have generally been studied independently of each other. It seems appropriate to view further studies in terms of a large and integrated study plan. Although such a plan may take many decades to complete, it can be accomplished by degrees. However, such a plan cannot achieve its full potential without a broad initial vision that includes both nested studies of multiple cores in small regions and transects of regional records across natural gradients. The potential of the Great Basin and adjacent regions for such studies is outstanding, and the studies described in this abstract and in the rest of this volume could be the beginning of a promising future.

Last 200,000 Years of Glacial History at Bishop Creek, Eastern Sierra Nevada, California

Andrew Bach, Ronald Dorn, and Tanzhuo Liu, Geography Department, Arizona State University
F.M. Phillips and Marek Zreda, Geosciences Department, New Mexico Institute of Mining and Technology
David Elmore, Purdue Rare Isotope Measurement Laboratory, Purdue University
Deborah Elliott-Fisk, Geography Department, University of California
James Clark, Chemistry Department, Arizona State University

Difficulties in accurate age determination and correlation of Quaternary age and geomorphic surfaces have been a long-standing problem in Quaternary research (Porter and others, 1983). The lack of materials appropriate for absolute age determination has caused workers to use various relative dating techniques to correlate deposits of Quaternary age (for example, Burke and Birkeland, 1979). The assumption used in relative dating is that lithology, climate, topographic position, and vegetation factors remain constant among sites so that the weathering of each deposit indicates the time that has elapsed since deposition (and the initiation of weathering processes). However, because lithologic and climatic factors vary between canyons of the east-central Sierra Nevada, conflicting conclusions by different workers at the same canyons and between canyons have resulted (Fullerton, 1986).

According to the classical chronology of Blackwelder (1931), the Sierra Nevada underwent only two major glaciations during the middle and late Pleistocene. Blackwelder's chronology was derived from the observation that most drainages in the eastern Sierra Nevada have only two or three easily distinguished sets of moraines. The basis for this chronology was examined using cosmogenic chlorine-36 (^{36}Cl) buildup in rocks, rock varnish ^{14}C, varnish cation ratios, and varnish microstratigraphy to establish a numerical chronology for the terminal-moraine complex at Bishop Creek, where the number of remaining moraines is unusually large. These dating methods indicated that there have been at least seven major glacial advances during the past 200,000 yr.

The timing of these advances seems to correlate well with intervals of maximum global ice volume that are inferred from marine oxygen-isotope records. However, unlike variation in global ice volume, the magnitude (equilibrium-line altitude decreases) of the various ice advances at Bishop Creek were nearly identical. Thus, the apparently small number of glaciations in the classical Sierra Nevada chronology is probably accounted for by a combination of obliterative overlap of older moraines by younger ones and by the limited temporal resolution of relative dating methods (Gibbons and others, 1984).

The Bishop Creek study area is in east-central California about midway between Reno, Nevada, and Los Angeles, California, near the town of Bishop, California. Bateman (1965, pl. 2) mapped the moraines using Blackwelder's (1931) classic stratigraphic nomenclature identifying Tioga, Tahoe, and Sherwin moraines. Additionally, a younger Tioga readvance and an older Tahoe moraine were identified. Detailed work on the lateral moraines confirmed Bateman's mapping and indicated that the older Tahoe moraine is a distinct pre-Tahoe deposit, rather than an early phase of the Tahoe deposit (Berry, 1991).

Although four lateral moraines are present along the course of Bishop Creek, at least 22 distinct pre-Tioga moraines can be mapped in the end-moraine complex. Previous studies have assigned all of these moraines to the Tahoe period (Bateman, 1965; Berry, 1991). Preservation of these moraines is probably due to the tectonic setting of the piedmont at Bishop Creek. Whereas other Pleistocene glacial deposits in the Sierra Nevada are cut perpendicularly by range-front faults, the lateral moraines are in a region of warping. Warping provides the steepest possible hydraulic gradient down the piedmont. Therefore, the glacier almost always follows the same path. When moraines are down-dropped by faulting, subsequent glacial advances easily cover older deposits, obliterating surface evidence of their existence (Gibbons and others, 1984). At Bishop Creek where faulting has not occurred on a large scale and the glaciers entered the piedmont at approximately a 45-degree angle to the piedmont slope, there is no tectonic mechanism to continually create a steep gradient. The glaciers behaved similarly to streams on an alluvial fan, flowing down one channel where material (till) accumulated until the flow gradient began to flatten, while steepening the gradient perpendicular to the axis of deposition. Given access to an alternative flow direction by ice overtopping a lateral moraine or some other mechanism, the glacier flowed down the steeper gradient, creating a topographic unconformity (cross-cut of moraines) (Bach and others, 1992). The final result of these changes of the course of these glaciers is that extensive later glaciations bypass moraines deposited by smaller earlier glaciation, and a large number are preserved.

Although these processes can occur in shorter time intervals (for example, surging), dating of the deposits indicates that they span at least 180,000 yr. The moraines have been grouped into seven age classifications (youngest to oldest, I to VII) based on analysis of rock varnish and the accumulation of cosmogenic ^{36}Cl in morainal boulders. Although analysis is not complete, preliminary age assignments are listed in table 1.

Deposits I through VI consist of multiple moraine crests that can be further subdivided into stadial events. Analysis of soil development on these moraines distinguishes four age groups (I, II, III–V, VI) that may correlate with soils on four lateral moraines up canyon (Berry, 1991). Soils formed on moraines in age group III–V vary in morphologic features, indicating either that the soils were of different ages or that the soil-forming factors varied in intensity between sites (most notably eolian additions), resulting in different soil development.

Numerical surface-exposure dating supports the hypothesis, based on morphologic mapping, of at least seven major glacial episodes at Bishop Creek. This finding strongly supports the obliterative overlap hypothesis of Gibbons and others (1984). Although Pleistocene glacial deposits in other canyons of the Sierra Nevada only contain deposits of two or three glacial stages, the fortuitous changes in glacial course at Bishop Creek allowed deposits to be preserved. The combination of obliterative overlap and of the limited temporal resolution of relative dating methods (Burke and Birkeland, 1979) has resulted in the apparently small number of glaciations identified in the Sierra Nevada.

Table 1. Estimated ages, rock-varnish microlamination, and solum depth of morained sequences, Bishop Creek, eastern Sierra Nevada

Deposit	Estimated age[1] (10^3 year)	Rock-varnish microlaminations	Solum depth (centimeters)	Relative age assignment [Bateman, 1965; Berry, 1991]
I	20–25	2	38	Tioga
II	31–37	4	56	Tioga
III	65–77	6	92	Tahoe
IV	95–115	8	101	Tahoe
V	155–170	10	110	Tahoe
VI	170–180	12–16	138	Pre-Tahoe
VII	>180	Eroded	79	Sherwin

[1]Based on chlorine-36 data and rock-varnish analysis.

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Thermoluminescence Tephrochronometry—Principles, Examples, and Potential

Glenn W. Berger, Quaternary Sciences Center, Desert Research Institute

Quaternary tephras are important for several reasons (Sarna-Wojcicki and Davis, 1991), particularly as a time-stratigraphic marker horizon; yet, until recently, there have been no means for the direct dating of tephra beds in the range 30,000 to 150,000 yr B.P. Although isothermal-plateau fission-track dating has recently decreased the upper range boundary to 70,000 to 100,000 yr B.P. (Chesner and others, 1991), there remains a considerable age gap for direct-dating methods and a need for an additional, widely applicable tephrochronometer for dating the middle and late Pleistocene. Fine-silt-size rhyodacitic glass shards (4 to 11 µm) in ash beds have been reported to be suitable for thermoluminescence (TL) dating (Berger, 1985), using the additive-dose procedure (Aitken, 1985; Berger, 1988) originally developed for fired pottery; however, only recently have sample purification and other appropriate procedures been applied to a suite of independently dated tephra (Berger, 1991, 1992). Although few independently well-dated samples were tested, the results (fig. 1) demonstrate that such glass can be an accurate TL clock for tephra to as much as about 450,000 yr.

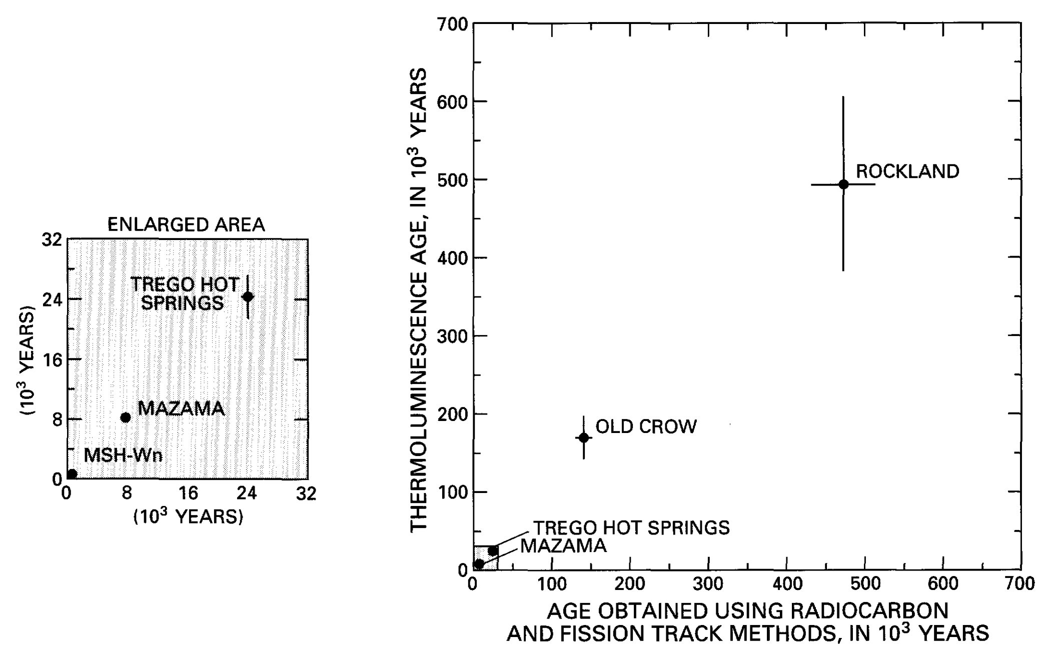

Figure 1. Relation between the thermoluminescence age determined by additive-dose procedure and carbon-14 or fission-track-based ages for five tephra samples (modified from Berger, 1985, 1991). The solid lines represent errors (±1 sigma). New, independent age estimates for Trego Hot Springs tephra (Oregon) (Negrini and Davis, 1992) [corrected to 24,000 calendar years (Mazaud and others, 1991)] and for Rockland tephra (California) (Alloway and others, 1992) are incorporated. Heavy-liquid separation was not applied to samples MSH-Wn (Washington) and Mazama (British Columbia). The natural 4 to 11 µm fraction of the Mazama sample contained 95 to 98 percent glass shards, but the proportion of glass in this fraction of sample MSH-Wn may be lower.

Principles

Volcanic glass shards can function as natural radiation dosimeters and are capable of storing a part of the absorbed energy from low-level environmental ionizing radiations (ambient and exogenous alpha-, beta-, gamma-, and cosmic-ray radiations). The stored energy can be read out as a luminescence signal that is proportional to the accumulated absorbed dose since last heating. Heating empties all electron traps (thermally stable lattice-defect sites). After burial, ionizing radiations repopulate emptied traps. In the laboratory, traps are again emptied, by heating, when electron-hole recombination produces luminescence. When the luminescence signal is related to an equivalent dose [(D_E) determined with the additive-dose procedure] by use of calibrated laboratory radiation sources, and an effective radiation dose rate is measured in separate experiments, then a thermoluminescence age equals D_E divided by effective dose rate. The objective of the additive-dose procedure is to obtain a plateau in D_E values that represents stable electron traps. Accurate measurement of a plateau D_E value has been the single greatest practical difficulty in the development of this new dating tool for tephra. Dose-rate measurements have been more straightforward.

The physical nature of glass shards and their inherent TL properties present practical difficulties. Shards larger than about 50 µm sometimes contain microlites and crystallites of feldspar minerals (Fisher and Schmincke, 1984). TL from such inclusions, which has recently been directly observed (Kanemaki and others, 1991), dominates the TL from glass. TL from volcanic feldspars has long been known to be very unstable (Wintle, 1973; Aitken, 1985) and needs to be avoided. This is the main reason that 4- to 11-µm glass shards were originally preferred by Berger and Huntley (1982). Separation in quantity of minerals in the fine-silt-size fraction traditionally has been difficult; however, Berger (1984) developed an effective heavy-liquid centrifugation procedure, employed for tephra for the first time by Berger (1987). Presently (1993), purification of the preferred 4- to 11-µm grain-size fraction to 95 to 98 percent glass seems the best that can be attained with a single repeat centrifugation.

Removal of laboratory-induced unstable TL also presents problems. As with unheated sediments, failure to minimize such unstable TL would generate age underestimates, other factors being equal. Berger (1991) tested different pre-readout heat treatments for removal of unstable TL and reported that heating sediments above 75°C for a few days can produce inaccurate D_E values, for reasons not yet understood.

Examples

Several previously undated ash beds from Summer Lake, Oregon, provided TL glass ages of as much as 200,000 yr (Berger and Davis, 1992), which were useful in determining the paleoenvironmental history of the enclosing pluvial-lake sediments. A TL glass age of approximately 50,000 yr for a Summer Lake tephra bed (ash bed 12) that was correlated to Mount St. Helens Set Cy tephra beds helps clarify the history of giant paleoflooding in eastern Washington (Berger and Busacca, 1991). In eastern Washington, another TL glass age of about 250,000 yr (Berger, 1991) for a previously undated ash bed in loess above a flood-cut unconformity clearly establishes that cataclysmic flooding in this region also occurred within the middle Pleistocene. Other examples of applications are described by Berger (1991). Some results need confirming by TL dating of additional material. One example is that of the Alaskan Sheep Creek tephra bed, whose first TL glass age was younger than expected from stratigraphic correlations based on geochemical fingerprinting of the tephra beds. This first age estimate is being confirmed using re-collected ash from more than one site.

Potential

The testing and application of TL need to be extended to glass of composition different from rhyodacite. Extension to a wider range of glass composition could provide a greater variety of dating targets, resulting in benefits for several fields of Quaternary geology. Additionally, possible extension to deposits older than about 500,000 yr needs testing. Finally, the encouraging result for the 500-yr MSH-Wn sample (fig. 1) indicates that TL tephrochronometry could be useful in studies of recent volcanism, such as in volcanic-hazards evaluation.

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Geochemical Variations in Sediments and Pore Waters of OL-92, a 323-meter Drill Hole from Owens Lake, California, and Their Probable Climatic Significance

James L. Bischoff, George I. Smith, and Jeffrey P. Fitts, U.S. Geological Survey
Kirsten M. Menking and Hannah M. Musler, University of California, Santa Cruz

During extreme dry periods (interglacials), Owens Lake was saline, alkaline, and biologically highly productive. In contrast, during extreme wet periods (glacials), the lake was flushed, overflowing with fresh water, and relatively unproductive. The transitions between these phases are indicated in the sedimentary column by marked variations in a number of geochemical parameters. Sediments from a 323-m drill hole (OL-92) were analyzed for radiocarbon, pore-water content and composition, $CaCO_3$, organic carbon, and cation-exchange capacity to determine these climatic changes and place them in a temporal context.

Point samples (50 cm^3) for water content and pore-water chemistry were obtained during the drilling operations at 2- to 3-m intervals for the entire length of the core (120 samples). Water content varied erratically down the core, generally decreasing from about 60 wet-weight (wt) percent at the top to about 20 wt percent at 240 m, but water content sharply increased at levels below 240 m to between 40 and 60 wt percent, indicating substantial undercompaction for the lowermost 100 m of the core. The pore waters were alkaline (pH 8–10) and had anionic compositions of HCO_3-CO_3>Cl>SO_4, which were similar to the chemistry of the modern lake before diversion of the source water. The pore-water salinity ranged from 0.4 to 5 wt percent dissolved solids. Salinity varied with depth in a smooth pattern; the minimum was at 30 m, gradually increasing to a single broad maximum at about 150 m, and sharply declining thereafter to steady low values at 210 m and below. The salinity-depth profile was drastically smeared by post-depositional diffusion of dissolved salts, indicating that the present (1993) pore-water composition has little relation to pore-water composition from the past climate.

Point samples were obtained from the top 24 m for accelerator-mass-spectrometer (AMS) radiocarbon analyses of extracted humates (15 samples). These results indicate a strong linear relation between age and depth (83 cm/10^3 yr). The average mass-accumulation rate of the Bishop Ash, as calculated from average pore-water content, average solid bulk density, and apparent accumulation rate (40.0 cm/10^3 yr) is 51.4 (g/cm^2)/10^3 yr. A similar calculation for the top 24 m, using radiocarbon control, results in 52.4 (g/cm^2)/10^3 yr, which is remarkably close to the calculated rate for the entire core. An age-depth curve for the entire core was determined using radiocarbon control from the top of the core and extrapolating to the Bishop Ash (759,000 yr at 309-m depth) after correcting for compaction using the pore-water data. The curve is based on the apparent constancy of the rate of sediment-mass accumulation during the extrapolation.

For determination of sediment composition, composite strip samples were obtained to represent the entire core (total of 91 samples). About 3.5-m segments of the core were sampled in a continuous strip or channel, each representing about 7,000 yr of deposition. These samples were individually homogenized, flushed of their interstitial salts, suspended in an aqueous CsCl solution, rinsed using distilled water, and dried. The CsCl treatment replaced all exchangeable cations (overwhelmingly Na) in the clay minerals with Cs. Aliquots were analyzed for percent $CaCO_3$, percent organic carbon, percent Cs, and grain-size distribution (sand/silt/clay).

The percent $CaCO_3$, percent organic carbon, and normalized Cs (percent Cs/percent clay) in the aliquots indicate cyclic variation down the core. All three parameters indicate conspicuous and sharp minima at 15,000 to 25,000 yr, corresponding to maximum glacial conditions. These results indicate that, during such conditions, the lake was nonproductive and that the clay fraction of the sediment (the major size component of the sediment) was dominated by glacial rock flour, which lacked cation-exchange capacity. Minima in these components individually (and more clearly in a normalized and summed composite) locate a series of older extreme glacial conditions extending back to 759,000 yr. Conversely, maxima locate depths of interglacial conditions. Summed composites indicate that Termination II occurred at 127,000 ± 10,000 yr.

Uranium-Series Dating of Sediments from Walker and Pyramid Lake Basins, Nevada

Charles A. Bush, Larry V. Benson, and John N. Rosholt, U.S. Geological Survey

For several years, studies have been made near many lakes in the Great Basin in an attempt to reconstruct regional paleoclimatic history from lacustrine sediments. Radiocarbon and uranium-series (U-series) methods have been used to determine ages of shoreline carbonates (tufas, oolites, and gastropods) to reconstruct the chronology of the rise and fall of the lakes in the Lahontan Basin (Lao and Benson, 1988; Benson, 1991). Organic and inorganic fractions of young lake sediment from the Walker and Pyramid Lake Basins in Nevada have been dated by ^{14}C methods (Yang, 1989; Benson, 1991). Dating by ^{14}C methods, however, can only be applied to sediments younger than approximately 40,000 yr B.P. Cored sediments from the Walker Lake and Pyramid Lake Basins extend beyond the range of ^{14}C dating; therefore, attempts have been made to apply U-series methods of age determination to these older core sediments.

The U-series-dating method, as applied to these lacustrine sediments, assumes that when $CaCO_3$ and organic materials are deposited in a subaqueous environment, dissolved uranium is coprecipitated with the $CaCO_3$ and organics and that the carbonates will contain negligible amounts of thorium. If the carbonate and organic fractions form a closed system for uranium and its daughter products, all ^{230}Th measured in the $CaCO_3$ and organic fraction is produced by radioactive decay of uranium after incorporation in the material.

Because the sediments also incorporate silicate-mineral detritus that contains significant concentrations of ^{230}Th and ^{232}Th, a correction for this initial or inherited ^{230}Th needs to be applied. Bischoff and Fitzpatrick (1991) have described a method of dating impure carbonates using an isochron-mixing line correction for inherited detrital ^{230}Th where the activity ratio of $^{230}Th/^{232}Th$ is plotted as a function of $^{234}U/^{232}Th$, and $^{234}U/^{232}Th$ is plotted as a function of $^{238}U/^{232}Th$. The slopes of the regression lines (Ludwig, 1991) of these plots provide the $^{230}Th/^{234}U$ and $^{234}U/^{238}U$ ratios used to calculate the age of the $CaCO_3$ and organic components in the sediment. The approach used here was to plot the measured ratios resulting from the total dissolution of several samples collected within a short interval in the sediment column. If the ratios of the samples have sufficient spread and little scatter about the regression line, slopes of the regression may be used to calculate an average age. This approach avoids the spurious fractionation of uranium and thorium that often occurs during preferential leaching of the detrital component (Bischoff and Fitzpatrick, 1991).

Two- to three m sections of core from 3 parts of Walker Lake core WLC84-4 (Benson, 1991), and 2 sections of less than 1 m from a Pyramid Lake core, PLC87-2, were each divided into 10 samples weighing 2 or 3 g. Each sample was analyzed by isotope-dilution alpha-spectrometry to determine activities of ^{234}U, ^{238}U, ^{230}Th, and ^{232}Th and $^{234}U/^{238}U$ and $^{230}Th/^{234}U$ ratios were calculated for each segment (table 1).

Table 1. Summary of sample data and calculated ages of sections of core from Walker Lake and Pyramid Lake

[Isotope-activity ratios with 1 standard deviation errors, and ages are calculated from those ratios; no age for the 27.4-meter depth of core PLC87-2 because errors were greater than the age; --, no data.]

Depth (meters)	Samples	Activity ratios		Age (10^3 yr B.P.)
		$^{234}U/^{238}U$	$^{230}Th/^{234}U$	
Core WLC84-4				
47.2	1–10	1.170 ± 0.028	0.391 ± 0.044	53 + 7/-7
71.6	1–10	1.278 ± .014	.534 ± .041	80 + 9/-9
95.5	1–7	1.134 ± .057	.749 ± .170	144 + 96/-52
95.5	1–10	1.228 ± .018	.825 ± .052	172 +28/-23
Core PLC87-2				
27.4	1–10	2.100 ± .549	.960 ± .164	--
58.1	1–10	1.338 ± .107	.726 ± .093	130 + 34/-27

Analyses of Section 20 of core WLC84-4 yielded an age of 53,000 ± 7,000 yr (1σ). Analyses of Section 28, also from Walker Lake, yielded an age of 80,000 ± 9,000 yr (1σ). When all 10 samples of Section 36 from Walker Lake were used, an age of 172,000 + 28,000/-23,000 yr (1σ) was calculated. The uranium concentration and ^{230}Th activity of the bottom three samples of Section 36 were substantially higher than of the upper seven samples in the section. However, the three samples are from radioactive growth slopes similar to growth slopes of the top seven samples. Regardless, they have been excluded from the age calculations because of their very different uranium concentrations. With the exclusion of the three samples, the calculated age becomes 144,000 yr and the 1-sigma error increases to +96,000/ -52,000 yr. Sarna-Wojcicki and others (1991) have estimated the age of an ash bed near the base of core WLC84-4 (145 m) to be 160,000 ± 25,000 yr. All ages estimated by U-series and tephrochronology are in correct stratigraphic order and seem to roughly follow the trend of sedimentation rate estimated by ^{14}C analyses done on sections from higher in the core (fig. 1).

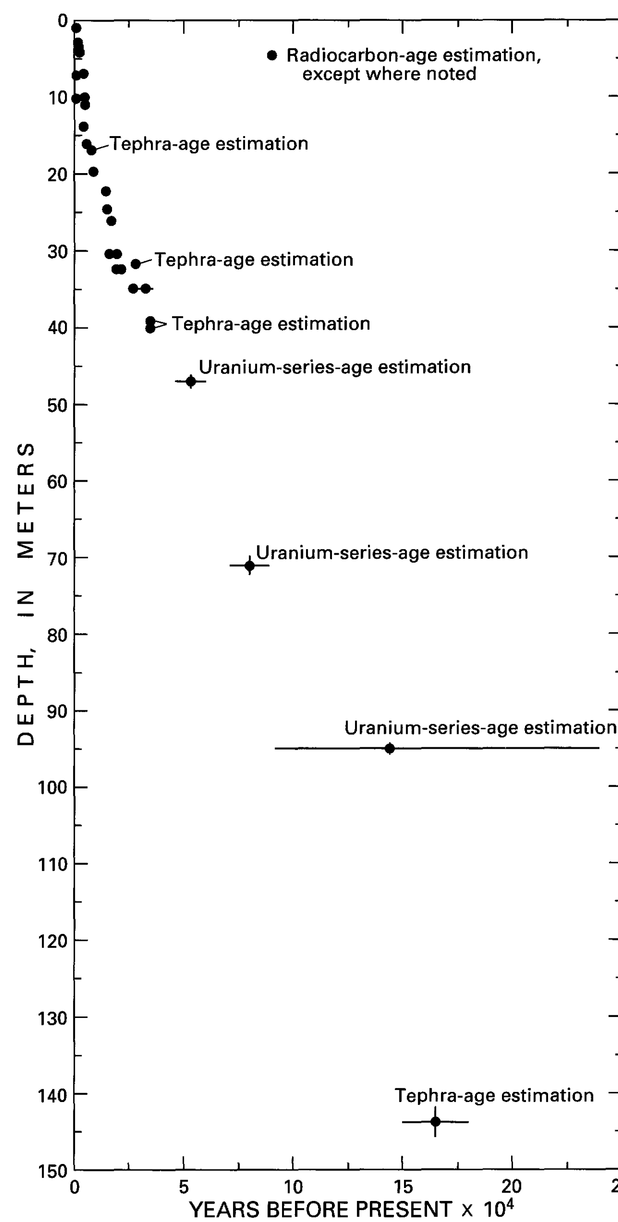

Figure 1. Age-depth plot of carbon-14 (not labeled), tephrochronology, and uranium-series dated segments of Walker Lake cores WLC84-4, WLC84-5, and WLC84-8.

The U-series-age estimates for lake sediments cannot account for any discontinuities; thus, deposition rates calculated from the U-series-age estimates are minimum values. Prior to 80,000 yr B.P., the rate of deposition was approximately 0.3 to 0.4 m/10^3 yr and after 80,000 yr B.P., the rate seems to have increased to 0.9 m/10^3 yr (fig. 1).

Results for the Pyramid Lake core are mixed. Although isochron lines may be fitted to the data from Section 19A of core PLC87-2, there is little extension along the line and much scatter away from the line. A calculated U-series age yields errors greater than the age itself. Examination of the core indicated the presence of a buried soil in the lower one-half of the section that had a wave-reworked sand in the upper one-half, so failure of the method is not surprising. However, analytical results for Section 31E of PLC87-2 had good extension and fit to the regression line and provided an age of 130,000 + 34,000/-27,000 yr. This age would indicate a minimum deposition rate of 0.45 m/10^3 yr.

This pilot study of five sections of core from Walker and Pyramid Lakes has indicated that closed systems for uranium and its daughters are sometimes formed within limited sections of sediment. Four of these sections provided reasonable U-series dates when compared with additional parts of sections within the core that were dated by other methods.

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Application of Magnetic Susceptibility and Grain Size in a Lithostratigraphic Study of Middle to Late Pleistocene Lacustrine Sediments Near Summer Lake, Oregon

Daniel B. Erbes and Robert M. Negrini, Department of Physics and Geology, California State University, Bakersfield

Positive results for intrabasinal correlation of outcrop and core from a lacustrine sequence of the Great Basin sites have been obtained using volume magnetic susceptibility and sedimentological data. The known sedimentary record from pluvial Lake Chewaucan in south-central Oregon contains more than 70 laterally continuous and distinct tephra layers, which enable high-resolution intrabasinal correlation of outcrop and core. This phenomenon has enabled testing of the effectiveness of volume magnetic-susceptibility data and sedimentological records as correlation tools. Preliminary results from this study indicate that magnetic-susceptibility records can be correlated throughout the entire sampled interval to within a few centimeters of stratigraphic depth between outcrops separated by 1 km. Similar results have been obtained correlating laterally distinct sands, pebble lags, and carbonate layers. Grain-size data are being collected using standard pipette and sedigraph methods to investigate their relation to these correlations.

Paleoclimatic Implications of Late Quaternary Lake-Level Variations in South-Central Oregon

Dorothy E. Freidel, Department of Geography, University of Oregon

Lake-level chronologies of three paleolakes in south-central Oregon indicated that these lakes reached their most recent highstands between 18,000 and 17,000 yr B.P., 3,000 to 4,000 yr earlier than the highest levels attained by Lakes Bonneville and Lahontan to the south. This study tests several climate scenarios that could have caused the lakes to rise to their highest levels at 18,000 yr B.P.

Paleobotanical data indicated that, at 18,000 yr B.P., conditions in the Pacific Northwest were colder and drier than at present (1993) (Thompson and others, in press; Barnosky and others, 1987; Barnosky 1985), while in the northern Great Basin, colder and somewhat moister conditions prevailed (Thompson and others, in press; Thompson, 1990). Few data have been collected for the area between these two regions where the Oregon lakes were located. However, based on estimates from these studies, air temperatures in south-central Oregon at 18,000 yr B.P. were about 7°C lower than at present.

Lake-ice simulations have indicated that a 7°C decrease in mean annual air temperature could have caused the paleolakes in Nevada to be ice covered during the winter months, accompanied by a 60-percent reduction in evaporation rates during that period (Hostetler, 1991). Water-balance modeling of the Oregon paleolakes indicates that they respond substantially to such a decrease in evaporation.

General atmospheric-circulation simulations have indicated that the mean position of the jet stream over western North America has shifted in response to changes in the large-scale controls of climate during the past 18,000 yr (Kutzbach and Guetter, 1986). Although coarse in resolution, the simulations indicate that, at 18,000 yr B.P., the mean winter position of the jet-stream core was at about the latitude of southern California, and the summer position intercepted the West Coast at about mid-Oregon. These atmospheric patterns would have brought south-central Oregon cold, dry winters; and cool, cloudy summers; and some increase in summer precipitation.

Thompson and others (in press) produced estimates of surface temperature, precipitation, and effective moisture anomalies for six regions of western North America, based on comparison of compiled paleoenvironmental data with simulations of a general atmospheric circulation model. The boundary between the northwestern and western regions of North America bisects Oregon, yet the model boundaries between climatic regions are too coarse to be interpreted precisely. At 18,000 yr B.P., the south-central Oregon paleolakes could have been affected by the cold, dry, glacial anticyclone prevalent in the Northwest in winter, and the cool, moist, westerly circulation of the Great Basin (in the western region) in summer (Thompson and others, in press). The Oregon paleolakes seem to have been in a transition zone between the two atmospheric regimes.

Several climate scenarios for the Oregon paleolakes were tested using a water-balance model. One scenario was based on Hostetler and Benson's (1990) hypothesis that historic precipitation and runoff maxima, plus evaporation, when decreased to 58 percent of present, were sufficient to cause Lake Lahontan to rise to its pre-highstand level. This scenario, using Oregon maxima (1.6 times present precipitation, 2.4 times present inflow) and 58 percent of present evaporation, caused the Oregon paleolakes to reach and maintain levels approximating their highstand shorelines.

A second set of scenarios based on Cooperative Holocene Mapping Program (COHMAP) simulations (COHMAP members, 1993; Thompson and others, in press) tests whether the Oregon highstands could have developed under the cold, dry conditions of the Pacific Northwest at 18,000 yr B.P. In these simulations, precipitation was decreased by 10 to 20 percent of present, evaporation was decreased by 50 percent [based on evaporation rates from Hostetler's (1991) lake ice-cover modeling], and runoff was estimated using Mifflin and Wheat's (1979) modifications of Langbein and others (1949) and Schumm's (1965) runoff curves, which were based on temperature and precipitation. One scenario, using 95 percent of present precipitation, 50 percent of present evaporation, 7°C colder than present temperature, and

3.35 times present runoff, also comes close to simulating highstand levels in the three Oregon basins. The runoff curves produced very rough estimates, and it is not known how accurately runoff was represented by these curves at low temperatures and low levels of precipitation. Comparison with maps of modern temperature, precipitation, and runoff distributions indicated that the combination of conditions represented in the second set of scenarios does not occur in modern environments. The level of runoff at 3.35 times present appears to be unrealistically high. Of the two scenarios that produce high lake levels, the scenario in which precipitation is increased seems more realistic.

In conclusion, paleoenvironmental data and climate modeling for the Pacific Northwest and the Great Basin and water-balancing modeling of three paleolakes in south-central Oregon indicated that, at 18,000 yr. B.P., cold, slightly moister than present climatic conditions caused the lakes to rise to their most recent highstands. The water-balance model scenarios that simulated conditions colder and drier than present could not simulate the highstand lake levels without using unrealistically high rates of inflow. The scenarios that simulated very cold conditions with no change in precipitation relative to present also required inflow rates that were quite high but within a plausible range. The modeling results are consistent with the hypothesis that at 18,000 yr B.P., south-central Oregon was within a transition zone between the very cold, dry glacial anticyclone to the north and the cold, moist, westerly circulation to the south. Cold winters, in which precipitation was somewhat lower than or the same as the present, and cool, cloudy summers, in which precipitation was higher than at present, were sufficient to cause the Oregon paleolakes to rise to their highest late Quaternary levels.

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Paleomagnetism of Owens Lake Sediments—Constraints on the Timing of Climate Fluctuations in the Quaternary

Jonathan M. Glen, Robert S. Coe, and Shannon S. Boughn,
Institute of Tectonics and Earth Sciences Department, University of California, Santa Cruz
George I. Smith, U.S. Geological Survey

Rock and paleomagnetic studies on three cores obtained from Owens Lake, California, by the U.S. Geological Survey to investigate late Pleistocene climate in the Western United States are described. Owens Lake, located just east of the Sierra Nevada, has probably received drainage (and accumulated sediments) by way of the Owens River and from local runoff from the Sierra Nevada for more than 3,000,000 yr.

The three cores, combined as a composite, span the top 323 m of lake fill consisting mostly of fine- to medium-grained sediments deposited during the past 800,000 yr. The cores, therefore, have the potential of providing excellent records of climate and paleomagnetic field changes within the Quaternary.

Paleomagnetic and rock magnetic studies are continuing on a suite of more than 500 discrete samples obtained from the composite section to provide magnetostratigraphy and to study in detail the Matuyama/Brunhes (M/B) transition and within-Brunhes excursions and secular variation (SV).

Because the cores were rotary drilled, individual core segments were rotated, resulting in a loss of absolute declinations. Therefore, field behavior was primarily interpreted on the basis of inclination, with additional aid from natural-remanent-magnetization (NRM) intensity and reconstructed declination data. In addition to determining the remanence directions, low field magnetic-susceptibility (MS) measurements were made.

NRM intensities of the samples varied by three orders of magnitude (10^{-7} to 10^{-4} emu/cm^3), the bulk of the range being due to variations in susceptibility, with a large fraction of sample NRM's being in the mid-10^{-7} range. Susceptibility (MS) depends on the amount and mineralogy of magnetic material present in a sample. These factors were potentially controlled (through a variety of mechanisms) by climate conditions. Therefore, variations in MS may be a proxy for climate change. The remanent magnetization, however, depends on the intensity of the Earth's magnetic field in addition to the concentration and types of magnetic grains in the sample. A first-order estimate of relative paleointensity of the magnetic field was, therefore, obtained by normalizing the NRM's to MS.

Results from the inclination record indicated the presence of several low inclination excursions (or possibly short polarity events) within the Brunhes and a 3-m section of transitional directions, interpreted as the M/B reversal, at the base of the composite core. Recent $^{40}Ar/^{39}Ar$ dates yield an age of 783,000 yr B.P. for the reversal. The depth and age of the reversal yield an average sedimentation rate of 40 cm/1,000 yr.

Shallow inclination anomalies seem to be a common feature within the Brunhes polarity chronology. These features may indicate global field excursions or short or aborted reversals. If so, they should correlate with features seen at globally distributed sites. With this in mind, the features identified in the Owens Lake cores were compared with those seen in other records. Because more events are contained in the Owens core than are present in the other records and because of the range of estimated excursion ages, unique correlations are difficult to establish.

A refinement of the correlations currently being made will be needed to establish the details (in direction and intensity) of the excursions. Even if such features are not global, regional records (within distances of 3,000 km) should be correlative. For this reason, events at sites in the Western United States [for example, the Mono Lake (28,000 yr B.P.) and Pringle Falls (150,000 yr B.P.) excursions] are of particular interest.

Prehistoric Human Occupation and Changing Lake Levels at Pyramid and Winnemucca Lakes, Nevada

Eugene M. Hattori, Nevada Historic Preservation Office and The California Academy of Sciences
Donald R. Tuohy, Nevada State Museum

Radiocarbon dates from Pyramid Lake and Winnemucca Lake archaeological sites in Nevada provide evidence for shifts in Holocene lake levels (table 1, fig. 1). Winnemucca Lake is especially sensitive to inflow variation because of its broad, shallow lakebed and the bedrock-controlled sill for its feeder channel, the Mud Lake Slough. In the past, Winnemucca Lake received water only when the Truckee River or Pyramid Lake rose above the slough's 1,177-m sill elevation. Prehistoric occupation or activity sites below the 1,177-m threshold elevation at Pyramid Lake most likely indicates that Winnemucca Lake was receiving no inflow and, in fact, that a playa had replaced the easily desiccated, shallow lake.

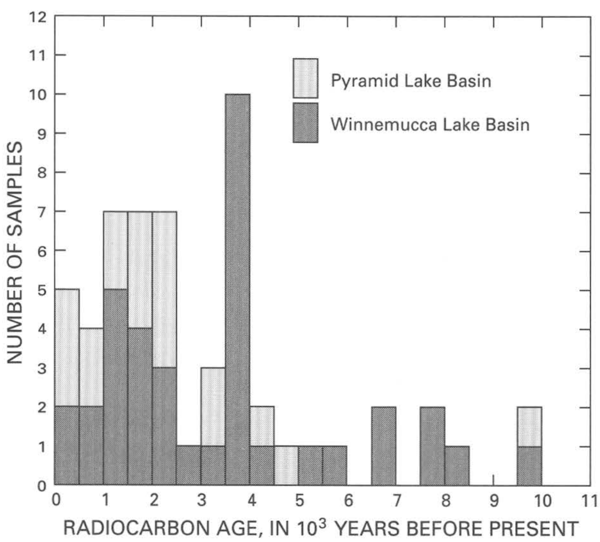

Figure 1. Radiocarbon ages of human occupation in the Pyramid and Winnemucca Lake areas.

Although more than 1,000 archaeological sites are located in the Pyramid Lake and Winnemucca Lake Basins, human occupation of the principal Winnemucca Lake sites are particularly sensitive to changing lake levels. The Falcon Hill and the East Shore archaeological areas at the northern end of the Winnemucca Lake Basin are several kilometers from springs and even some distance from the playa surface. Use of these sites was dependent on a water body existing in the basin. The contents of these sites also demonstrated the occupants' dependence on marsh and lake resources for sustenance and for raw materials for tools. All dates are presented as ^{14}C yr B.P.

The earliest radiocarbon dated artifact was a 9,660 ± 170 yr B.P. (laboratory number GX-13744) date on fishing line recovered near the Pinnacles at Pyramid Lake (table 1). At Winnemucca Lake, open-twined basketry from a shelter at Shinners Site-A (26Wa198) was dated to 9,540 ± 120 yr B.P. (UCLA-675). Following these dates, there are no cultural dates between 9,500 and 8,400 yr B.P. Then, a series of three dates spans the interval between 8,380 ± 120 (UCLA-672) and 7,830 ± 350 yr B.P. (laboratory number L-289) (table 1). During each of these dated occupations, Pyramid Lake was most likely above the threshold elevation, and Winnemucca Lake held water. Artifacts associated with these early Holocene dates include an atl-atl; edge-ground, stemmed projectile points; barbed bone points; plain-twined basketry; Great Basin crescents; and fish netting.

Between 7,800 and 4,100 yr B.P., there are only six dates. Although people inhabited the area during this interval, conditions apparently were not conducive to intensive site use within the Pyramid and Winnemucca Lake Basins. The 1,153-m elevation for the mano dated to 4,600 ± 60 yr B.P. (laboratory number LL-CM-26A21) from Pyramid Lake is accepted with caution because the mano is small and portable. Between 4,100 and 3,200 yr B.P., 13 dates indicate an interval of increased intensity of occupation and increased use of the marsh and lacustrine resources of Winnemucca Lake. This occupation is marked by increased development of material culture and increased contact with prehistoric cultures in coastal and central California. Marine shell ornaments, decorated plain-twined basketry, Humboldt and Gatecliff (Pinto)-type projectile points, western pond turtle-shell ornaments, and juniper seed beads are characteristic artifacts. In addition to turtles, the Winnemucca Lake Basin also supported populations of water birds, cutthroat trout, *cui-ui*, and marmot—animals no longer present in this lake basin. Coiled basketry containers also make their first appearance during this interval.

Lowered lake levels probably occurred between about 3,300 and 2,600 yr B.P given the limited number of archaeological sites that date to this interval (table 1). A brief interval between 2,600 and 2,400 yr B.P. yielded six dates; one of these was of particular interest. Two burials from a 1,154-m beach were dated at 2,480 ± 120 yr B.P. (laboratory number GaK-2386) from an associated piece of sagebrush (*Artemisia tridentata*) (table 1). This age, associated with a very low Pyramid Lake level, is similar to dates on basketry from lake-dependent sites at the extreme northern end of the Winnemucca Lake Basin. This potential discrepancy of a depressed Pyramid Lake level and an elevated Winnemucca Lake level may be due to the occurrence of a mesic interval that was punctuated by a rather severe drought or to diversion of the Truckee River solely to the Winnemucca Lake Basin. During this interval, between 2,600 and 2,400 yr B.P., coiled basketry greatly increased in frequency; twining, however, still persisted.

After an interval of some 400 yr between about 2,400 and 2,000 yr B.P. (with only two dates around 2,150 yr B.P.), there is a 900-yr interval between 2,000 and 1,100 yr B.P. that had 14 dates from 10 Winnemucca Lake and Pyramid Lake archaeological sites. A lake, or marsh, or both, was present in the Winnemucca Lake Basin, and Pyramid Lake seems to have been above the 1,177-m threshold elevation for most of this time. During this late Holocene interval, Lovelock wickerware plaiting, the hallmark of the Lovelock Culture, appeared. The other basketry technologies, plain-twining and coiling, still persisted. Projectile point styles included the Elko-type points. California trade items, especially marine shell ornaments, again appeared in large quantities.

The last 1,100 yr are represented by nine dates and by generally depressed lake levels. A cluster of four dates between 600 and 400 yr B.P. indicated the presence of water in Winnemucca Lake at that time. Lovelock wickerware, twining, and coiling persisted.

At 400 yr B.P., a major change in basketry technology appeared. A twill-twined water bottle from Pyramid Lake dates to 380 ± 100 yr B.P. (laboratory number GaK-2385) (table 1). This basket-weaving technique is ethnographically associated with the Northern Paiute Indians, the modern inhabitants of the area. This find was the first recognized appearance of the group in the local archaeological record. Other Pyramid Lake dates also associated with the Northern Paiutes include a 290 ± 90 yr B.P. (laboratory number WSU-1960) date on a house floor (Smithsonian number 26Wa1014-I) and, from the same site, a 240 ± 100 yr B.P. (laboratory number GaK-2389) date on an arrowshaft tipped with a desert-side-notched-type projectile point (table 1). There are no post-380 yr B.P. dates from the Winnemucca Lake Basin, although historic artifacts and ethnographies indicate that the basin continued to be occupied by the Northern Paiute, but possibly with reduced intensity.

When J.C. Fremont entered the region in 1844, Winnemucca Lake may have been dry, and Pyramid Lake stood at the Mud Lake Slough threshold elevation of 1,177 m, which was probably the beginning of the historic highstand. By the 1880's, Winnemucca Lake reached its maximum depth of 26 m; lacustrine biota had recovered enough to support a Chinese fishing camp along the western shore of the lake. A lake and marsh persisted in the Winnemucca Lake Basin until the historic water diversions of the 1900's.

Table 1. Radiocarbon dates for archaeological materials from Pyramid Lake and Winnemucca Lake

Lake	Carbon-14 age	Laboratory number	Site (Smithsonian number)	Specimen dated
Pyramid	240 ± 100	GaK-2389	Marble Bluff Dam (26Wa1014-I)	Arrow w/desert-side-notched point
Pyramid	290 ± 90	WSU-1960	Marble Bluff Dam (26Wa1014-I)	House floor (Paiute?)
Pyramid	380 ± 100	GaK-2385	Water Bottle Cave (26Wa372c)	Twill-twined basketry
Winnemucca	390 ± 80	UCLA-982	Shinners Site-I (26Wa205)	Grass w/burial
Winnemucca	400 ± 80	UCLA-985	Shinners Site-I (26Wa205)	Basketry
Winnemucca	580 ± 100	UCLA-677	Shinners Site-F (26Wa202)	Basketry
Winnemucca	595 ± 80	UCLA-986	Shinners Site-A (26Wa198)	Matting
Pyramid	620 ± 80	GaK-2810	Warrior Pt. Cave (26Wa729T)	Basketry
Pyramid	860 ± 90	SI-3430	Firepit III (26Wa1020)	Carbon
Winnemucca	1,150 ± 100	UCLA-668	Shinners Site-I (26Wa205)	Basketry
Winnemucca	1,190 ± 80	UCLA-673	Shinners Site-A (26Wa198)	Basketry
Winnemucca	1,240 ± 80	UCLA-906	Shinners Site-C (26Wa200)	Basketry
Pyramid	1,275 ± 160	La-73-113	Mixon Cave (26Wa528)	Digging stick
Pyramid	1,340 ± 100	GaK-2809	Mule Ears Cave (26Wa528)	Basketry
Winnemucca	1,400 ± 150	WSU-269	Empire Cave (26Wa197)	Basketry
Winnemucca	1,480 ± 155	WSU-268	Empire Cave (26Wa197)	Basketry
Winnemucca	1,510 ± 200	M-346	Crypt Cave (26Pe3a)	Fur robe
Winnemucca	1,610 ± 80	UCLA-671	Empire Cave (26Wa197)	Basketry
Pyramid	1,660 ± 60	LL-289B9	Pinnacles Beach (26Wa218B)	Tufa on percussor (1,179 m)
Winnemucca	1,725 ± 120	UCLA-933	Shinners Site-A (26Wa198)	Basketry
Pyramid	1,820 ± 180	I-2846	Marble Bluff Dam (26Wa1016)	Bone collagen
Winnemucca	1,860 ± 70	UCLA-124	Shinners Site-C (26Wa200)	Basketry
Pyramid	1,950 ± 100	GaK-2804	Desiccation Cave (26Wa291)	Matting
Pyramid	2,140 ± 110	GaK-2390	Mogoose Cave (26Wa275)	Basketry
Winnemucca	2,175 ± 80	UCLA-904	Shinners Site-A (26Wa198)	Basketry
Winnemucca	2,400 ± 200	L-289 II	Crypt Cave (26Pe3a)	Basketry
Pyramid	2,410 ± 90	GaK-3361	Blazing Star Cave (26Wa275)	Basketry
Pyramid	2,430 ± 100	GaK-2388	Mogoose Cave (26Wa275)	Basketry
Winnemucca	2,440 ± 100	UCLA-674	Shinners Site-C(26Wa200)	Basketry
Pyramid	2,480 ± 120	GaK-2386	Burial site (26Wa404)	Sagebrush
Winnemucca	2,590 ± 80	UCLA-692	Chimney Cave (26Pe3b)	Juniper bark mat
Pyramid	3,040 ± 115	SI-3429	Pit House & Burials (26Wa1016)	Carbon
Pyramid	3,270 ± 180	GaK-2805	Coiled Jug Cave (26Wa315)	Basketry
Winnemucca	3,325 ± 90	UCLA-931	Shinners Site-F (26Wa202)	Basketry
Winnemucca	3,620 ± 80	UCLA-976	Kramer Cave (26Wa196)	Basketry

Table 1. Pyramid Lake and Winnemucca Lake, radiocarbon dates from archaeological materials—Continued

Lake	Carbon-14 age	Laboratory number	Site (Smithsonian number)	Specimen dated
Winnemucca	3,660 ± 80	UCLA-905	Kramer Cave (26Wa196)	Basketry
Winnemucca	3,660 ± 100	UCLA-979	Kramer Cave (26Wa196)	Basketry
Winnemucca	3,700 ± 80	UCLA-984	Kramer Cave (26Wa196)	Basketry
Winnemucca	3,720 ± 100	UCLA-122	Kramer Cave (26Wa196)	Dart foreshaft
Winnemucca	3,745 ± 90	UCLA-932	Kramer Cave (26Wa196)	Basketry
Winnemucca	3,760 ± 80	UCLA-980	Kramer Cave (26Wa196)	Basketry
Winnemucca	3,830 ± 110	GaK-2387	Kramer Cave (26Wa196)	Dart foreshaft
Winnemucca	3,850 ± 100	UCLA-983	Kramer Cave (26Wa196)	Matting
Winnemucca	3,900 ± 100	UCLA-670	Kramer Cave (26Wa196)	Basketry
Winnemucca	4,030 ± 85	UCLA-978	Shinners Site-C (26Wa200)	Cordage
Pyramid	4,470 ± 110	GaK-2808	Blazing Star Cave (26Wa525)	Digging sticks
Pyramid	4,600 ± 60	LL-CM-26A21	Wizard's Beach	Tufa on mano (elev 1,153 m)
Winnemucca	5,100 ± 150	WSU-270	Shinners Site-D (26Wa201)	Matting
Winnemucca	5,670 ± 150	L-289 FF	Cowbone Cave (26Pe3c)	Juniper bark mat
Winnemucca	6,500 ± 150	L-596	Guano Cave (26Pe3d)	Occupation floor detritis
Winnemucca	6,730 ± 90	UCLA-981	Shinners Site-I (26Wa205)	Grass in cache pit
Winnemucca	7,830 ± 350	L-289	Fishbone Cave (26Pe3e)	Fish netting
Winnemucca	7,980 ± 610	I-6873	Nicolarsen Site (NV-Wa-197)	Basketry
Winnemucca	8,380 ± 120	UCLA-672	Shinners Site-A (26Wa198)	Matting
Winnemucca	9,540 ± 120	UCLA-675	Shinners Site-A (26Wa198)	Basketry
Pyramid	9,660 ± 170	GX-13744	Pinnacles Area (26Wa217A)	Fishing line

High-Resolution Simulations of Climate of the Western United States at 18,000 Years B.P.

Steve W. Hostetler, U.S. Geological Survey,
F. Giorgi and Gary T. Bates, National Center for Atmospheric Research, and
Patrick J. Bartlein, Department of Geography, University of Oregon

Simulations of perpetual January and July climate at 18,000 yr B.P. over the Western United States have been conducted using a high-resolution (mesoscale) atmospheric model (RegCM). The RegCM, the climate version of the NCAR/PSU mesoscale model MM4, was run using a grid spacing of 60 km, a resolution that captures regional atmospheric circulations and terrestrial features (such as coastlines, mountains, and lakes) that exert strong forcings on the climate of the West. The model includes a surface-physics package (Biosphere Atmosphere Transfer Scheme) that couples the biosphere with the atmosphere and includes a fully interactive lake model to simulate lake-atmosphere feedbacks associated with Lakes Bonneville and Lahontan.

A series of 90-day, 18,000-yr- and 0-yr-B.P. simulations were conducted to make estimates of the general climate and of the effect of feedbacks between the lakes and the atmosphere. Initial and lateral boundary conditions (sea surface temperatures and vertical profiles of temperature, humidity, and wind velocity) for the simulations were obtained from output from the paleosimulations of Kutzbach and Guetter (1986). The surface of RegCM included estimated distributions of vegetation, montane glaciers, the continental ice sheet, and the areas and depths of Lakes Bonneville and Lahontan at 18,000 yr B.P.

Results indicate that throughout the West, January air temperatures at 18,000 yr B.P. were about 2°C colder than the 0-yr-B.P. control, whereas July temperatures were more than 5°C colder than the 0-yr-B.P. July control run. January precipitation throughout the region was substantially greater at 18,000 yr B.P. However, July precipitation at 18,000 yr B.P. was not much different than that at 0 yr B.P., except around Lakes Bonneville and Lahontan.

Analyses indicate that the hydrologic signal associated with the position and strength of the jet stream at 18,000 yr B.P. dominates the hydrologic budgets of the Bonneville and Lahontan Basins. Over the Bonneville Basin, lake-effect precipitation was a substantial component of the hydrologic budget in January and July of 18,000 yr B.P. However, such feedbacks were minor components of the hydrologic budget of Lake Lahontan.

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Initial Results from Paleoclimatic and Geophysical Studies of Washoe Lake, Nevada

Robert E. Karlin, James T. Trexler, and Ronald Petersen,
Department of Geological Sciences, Mackay School of Mines, University of Nevada-Reno

A lacustrine sediment sequence from Washoe Lake in western Nevada was cored in 1992 for paleoclimatic and paleohydrologic studies. In late June 1991, the lake desiccated for the first time since 1933–34 and for only the second time in recorded history, offering a rare opportunity to obtain long continuous sections from a quasi-permanent lake in a climatically sensitive region.

Washoe Lake, located between Carson City and Reno, Nevada, is in the high-Sierran semiarid desert environment near the present boundary between the pinyon pine/mountain mahogany and the sagebrush ecotomes. In late spring 1993, the lake was approximately 31 km^2 in area and averaged 3 to 4 m in depth. During the Pleistocene, the water depth was about 13 to 17 m, and the lake occupied a much more extensive area, as evidenced by strand lines above the lake margins. Geologic studies of the basin sides indicated that lacustrine sediments have been deposited for at least the last 2,500,000 yr.

Gravity, magnetic, and electrical surveys were conducted to characterize the geometry and subsurface structure of the basin. The lake is in an asymmetric fault-bounded half graben surrounded by the Carson and Virginia Ranges, the Truckee Meadows, and the Carson Basin. Gravity modeling of the basin indicates that the lake is underlain by sediment thicknesses of almost 600 m; maximum accumulations are in the western end. A major change in sedimentary facies from coarse alluvial sands in the west to fine-grained lacustrine sediments in the east coincides with a deep basement offset identified from magnetic anomaly modeling. The deepest part of the present (1993) lake also is east of the zone of thickest sediment accumulation. These relations indicate that ancient and modern sedimentation patterns in the lake were and are tectonically controlled.

Two sites on the dry lake bed were cored to 25 m, and undisturbed sediments were obtained, resulting in 85-percent recovery. Two other sites yielded about 13 m of section until penetration was terminated because of fluidized sands. In addition, two 4-m trenches were sampled to obtain recent sedimentation history. The drill holes were geophysically logged to allow the sections to be reoriented in situ. Magnetic-susceptibility profiles of all sections were measured to characterize the stratigraphy of the lake. To date, sediment chemistry, textural analyses, and X-ray diffraction have been performed on sediments from the core site and on sediments from one trench.

Near the depositional center of the lake basin, the top 12 m of sediment consists of alternating intervals of lake sediments, peaty-bog deposits, and soils. The lower part of the drill core contains mostly lake sediments probably deposited within the last 60,000 yr. The lake sediments are characterized by elevated $CaCO_3$ concentrations, low susceptibilities, and relatively low abundances of terrigenous elements, such as Al, K, and Na, which are usually associated with detrital aluminosilicates. The carbonate minerals are calcite and aragonite that were authigenically precipitated during alkaline and probably saline lake stages. The soils show an inverse relation of $CaCO_3$ to terrigenous elemental abundances. Peaty intervals show relatively high organic carbon, low $CaCO_3$, and intermediate Al_2O_3 concentrations.

Magnetic-susceptibility profiles provided a sensitive, continuous record of lithologic change. Susceptibility maxima mostly indicated intermittent sand layers that were deposited in different environments. Textural variations were used to map transport pathways and depositional environments. Sands present in different depth intervals were derived from turbidites and subaerial sheet-flow deposits that formed during lake stages. Discontinuous channelized fluvial sands were deposited during the peaty-bog stages, and deflationary lag deposits formed during arid-soil stages.

The sediment record indicated that climate varied substantially during the Holocene. The top 3 m of sediment contained several silt/clay layers, some of which contained clam zones and peaty zones and soils, indicating that the lake had partially desiccated several times during the Holocene. Radiocarbon dating of three clam layers in the top 1 m and the presence of the Mount Mazama ash allowed calculation of sedimentation rates of 23 cm/10^3 yr from 2,600 to 0 yr B.P. and 11 cm/10^2 yr from 6,850 to 2,600 yr B.P. In the last 2,600 yr, the climate in the study area has been relatively wet, and Washoe Lake has been fully developed, except for three brief desiccation episodes. Earlier in the Holocene, the climate was generally drier and more variable, causing deposition of soils and marshy peat horizons. The results from Washoe Lake are consistent with a warm period from about 6,000 to 4,000 yr ago that apparently dropped the level of Lake Tahoe 10 to 20 m (Rose and Lindstrom, this volume). Further radiocarbon dating of lake, peat, and soil zones would enable determination of the timing, frequency, and duration of previous episodes of drought.

Detailed Sedimentary Record of Changes in the Level of Lake Lahontan from the Hot Springs Mountains, Nevada

Nicholas Lancaster, Quaternary Sciences Center, Desert Research Institute

Understanding the nature of late Pleistocene lake-level fluctuations requires detailed stratigraphic and sedimentological studies of cores and of exposed sections of lacustrine sediments. Initial results from investigations of nearshore lacustrine and eolian sediments from the Hot Springs Mountains, Nevada, are described in this abstract.

A series of sand ramps (coalesced climbing dunes) of late Holocene age cover the western flanks of the Hot Springs Mountains, located about 20 km east of Fernley, Nevada. In places, the eolian deposits have been trenched by ephemeral streams to expose the underlying beach, lacustrine, colluvial, and eolian deposits that were laid down at or near the shoreline of Lake Lahontan. These deposits could provide a detailed record of lake-level changes and associated depositional environments.

Investigations of the best exposure indicated the following sequence of deposits. Unit I forms the base of the exposed section and consists of rounded beach gravels, cobbles, and boulders that fine upward to fine gravels and shelly sands. Unit I thickens shoreward from 2.5 to about 9 m and is capped by a 20- to 50-cm boulder colluvium that has thick desert varnish on the clasts. Unit II, which is as much as 20 m thick, is a series of horizontally bedded, very fine sands to silty sands and occasional thin silt layers. In the lower 10 m of the unit, there are two thin fluvial subunits containing reworked tufa and beach gravels. Unit III overlies Unit II at an erosional contact and consists of a set of large-scale eolian cross beds.

A preliminary interpretation of the sequence of deposits is that Unit I represents a high-energy beach deposited in a transgression of Lake Lahontan to an elevation of 1,277 m. The lake then fell to less than 1,260 m, allowing colluvium to extend over the eroded surface of the beach deposits. A renewed transgression to an elevation of at least 1,293 m deposited Unit II as a series of nearshore sands. Following deposition of Unit II, the lake fell again to some lower elevation, causing a brief hiatus in sediment deposition around 1,272 m. There was extensive subaerial erosion and gullying of the lacustrine deposits prior to deposition of the eolian sands of Unit III.

Age control of the deposits is currently (1993) poor. Unit I and the lower part of Unit II contain abundant small gastropod shells. Samples submitted for radiocarbon dating provided an age of 12,660 ± 80 yr B.P. for Unit I and an age range between 11,920 ± 110 to 12,780 ± 90 yr B.P. for Unit II. The dates for Unit II are not in stratigraphic order and seem to be inconsistent with other data from the Lahontan Basin, which indicated that lake levels were below 1,250 m around 11,000 yr B.P. Contamination of the shell material is, therefore, suspected. A speculative correlation with other areas indicated that the horizontally bedded lacustrine deposits represented the latest Sehoo transgression, dated elsewhere to be around 13,000 yr B.P.

The Hot Springs Mountains also have an extensive series of wave-cut benches and other shoreline features. Eight shorelines at elevations between 1,277 and 1,328 m have been identified adjacent to the main section described above. Elsewhere in the area, 11 shorelines have been recognized at elevations that range from 1,279 to 1,332 m.

To further refine age control of the deposits and shorelines in the Hot Springs Mountains area, detailed mapping and sedimentological studies are needed. These studies are now in progress.

Changes in Pyramid Lake Sediment Composition in Relation to Fluctuations in Lake Level and Land-Use Practices

Martin E. Lebo and John E. Reuter, Division of Environmental Studies, University of California, Davis
Phillip A. Meyers, Department of Geological Sciences, University of Michigan

Sediment cores were collected from the shallow and deep basins of Pyramid Lake, Nevada, to analyze changes in the elemental composition of sedimentary material during the past 600 to 1,000 yr. Changes in carbon (C), nitrogen (N), and phosphorus (P) content of particles and pore-water nutrient concentrations are associated with variations in lake volume, water supply, and overall productivity of the lake. General relations of human activities in the Truckee River Basin to changes in sediment composition were determined for the historical record, and then those relations were used to analyze deeper parts of the cores.

Several relations of human activities or climate variations to sediment composition were identified. First, the $CaCO_3$ and particulate C content of sediment in the shallow basin decreased when river inflow to the lake was large. Conversely, $CaCO_3$ content of sediments throughout the lake decreased when river loading of calcium (Ca) was low during decreased inflow. Finally, total P content of sediments increased with decreasing lake volume. This unexpected relation was due to higher inorganic P content of particles as lake level decreased and may have been part of a mechanism maintaining constant P concentration in Pyramid Lake despite fluctuations in lake volume and P loading over time.

The observed changes in the composition of sediments during the past 600 to 1,000 yr indicated the productivity of Pyramid Lake may have increased to the current mesotrophic conditions in two stages. P.A. Meyers and others (1980), in an earlier study of Pyramid Lake sediments, suggested that productivity increased from oligotrophic to mesotrophic conditions at around A.D. 1300 to 1400. The compositional data did not provide strong support for an increase in productivity at that time. Changes in elemental composition of sediments provided much stronger evidence for higher productivity in Pyramid Lake beginning in A.D. 1600. In sediments of the deep basin, organic P (PP_{org}) content of sediments increased abruptly from A.D. 1600 to 1700 and from A.D. 1850 to 1935. Increase in productivity in the latter period also was supported by an increase in particulate N (PN) content of sediment. A comparison of phosphate and ammonium concentrations in pore waters indicated that the remineralization of organic matter over time could not account for the magnitude of observed increases in PP_{org} and PN content for these two periods. The observed increases, therefore, could be interpreted as increases in lake productivity.

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Paleomagnetic Dating and Correlation of Pre-Holocene Lake Sequences in the Northern Great Basin

Joseph C. Liddicoat, Department of Environmental Science, Barnard College, Columbia University

The paleomagnetic field behavior recorded in sediment deposited in Pleistocene lakes in the northern Great Basin has been studied in sufficient detail that dating and correlation of stratigraphic sequences may be possible within and between basins that contained some of the larger lakes (Lahontan, Bonneville, Russell). For the most part, however, dating and correlation are not possible unless a sequence contains a reversal of polarity or indicates unusual field behavior that functions as a stratigraphic marker. Two such markers are (1) a path of the Virtual Geomagnetic Poles that loops through the southern Pacific Ocean that is recorded in sediment on Paoha Island in the Moño Basin (California) and in Long Valley (California) and is assigned an age of about 150,000 yr, and (2) the Mono Excursion that is present in the Mono and Lahontan Basins in sediment dated by ^{14}C and tephrochronology to be between 28,000 and 26,000 yr old. Short- and long-term fluctuations in the declination and inclination recorded in the Sehoo and Eetza Formations in the Truckee River Canyon between Wadsworth and Nixon, Nevada, currently (1993) are being compared to time-equivalent (?) records for the Bonneville and Alpine Formations in the Bonneville Basin near Delta, Utah, to help determine the effectiveness of dating and correlation of spatially separated lake sequences.

Design of Drive Samplers for Cohesive Sediments

Daniel A. Livingstone, Departments of Zoology and Geology, Duke University

Most drive samples for cohesive sediments are crude and only suitable for taking cores a meter or less in length. Adherence to several design criteria would increase the length of core that could be taken in a single drive beyond the 30 m sometimes achieved at sea. In particular, it is important to minimize the force needed to drive the sampler and maximize the length of core barrel that can be driven without buckling.

All other things being equal, the force required to core cohesive sediment increases with the kerf ratio (that is, the ratio of wall area to sample area) of the sample tube:

$$\text{kerf ratio} = \frac{(\text{outside diameter})^2 (\text{inside diameter})^2}{(\text{inside diameter})^2} \quad (1)$$

Apart from the Shelby tube and some rod-driven piston samplers, all common core samplers used by limnologists and oceanographers have a very high kerf ratio. This high kerf ratio limits the length of sample that can be taken in a single drive in deep water. The high kerf ratio increases the weight and cost of the driving apparatus in any depth of water, and it increases the bow wave, tending to blow away soft sediment at the mud-water interface.

In addition to needing a low kerf ratio, an effective sampler needs a sharp cutting shoe that has a fine taper on the outside. The length of a single drive must not exceed a safe sampling distance, which depends on the ambient hydrostatic plus atmospheric pressure. The feedback loop between normal pressure and frictional resistance that develops as a sample tube is driven into the sediment needs to be controlled in some way. On the inside wall of the tube, feedback is decreased by a tight-fitting piston or by foils paid out from the cutting shoe. Inside feedback also can be decreased by foils on the outside, but there is no way of accommodating so much foil in a thin-walled shoe. Outside feedback also can be decreased by using a cutting shoe and couplings of an outside diameter slightly greater than that of the sample tube or by using a tube so short—for example, a Shelby tube—that the frictional forces do not exceed manageable levels.

In most samplers, the kerf ratio is high because of excessive outside clearance or a thick plastic liner, or both. In free-fall samplers, for use in deep water, there has been a tendency to increase the stiffness of the core barrel by increasing the wall thickness, but increasing the wall thickness is not effective because the heavy wall increases proportionately the load that needs to be imposed on the core barrel to drive it. A more effective method is to increase core length by using a thin-walled core barrel of large diameter that has a fine cutting shoe, a small outside clearance (2–3 percent), and no liner. Mooring the vessel from which the sampler is operated will eliminate the bending movement due to lateral motion of the vessel and the samples. Mooring will change the core barrel from something close to an Eulerian slender column with one fixed end and one free end to something much closer to a column with two fixed ends. This increases the amount of force that can be used to drive the core barrel by a factor approaching 16 and the length of core barrel that can be driven without buckling by a factor approaching 4.

If the free-fall sampler is used for minimally disturbed sampling of the mud-water interface instead of maximal depth of penetration, speed of penetration needs to be controlled with a hydraulic choke.

Late Quaternary Stratigraphy of Owens Lake, California

Steve P. Lund, Mark S. Newton, and Douglas E. Hammond, Department of Earth Sciences, University of Southern California
Owen K. Davis, Department of Geosciences, University of Arizona
J. Platt Bradbury, U.S. Geological Survey

Introduction

Owens Lake (36°30'N., 118°W.) is a large closed-basin lake located on the western margin of the Basin and Range province in east-central California. The lake is fed primarily by the southward-flowing Owens River. Owens Lake and Owens River are in a north-south-trending graben called the Owens Valley, which is bounded by the Sierra Nevada on the west and the Inyo Range on the east. Owens Lake is in the rain-shadow of the Sierra Nevada and currently (1993) receives less than 13 cm of rainfall per year, mostly from strong winter storms generated in the northeast Pacific Ocean. The drainage basin of the lake covers more than 6,500 km^2, including 180 km^2 of the Pleistocene-glaciated eastern Sierra Nevada margin. Prior to diversion of Owens River water for irrigation (1873), Owens Lake was a shallow (less than 10-m-deep) hypersaline lake that had a surface area of 290 km^2 (Gale, 1914). The site of the former lake is currently a salt-encrusted playa, 186 km^2 in area at an elevation of 1,081 m.

During the Pleistocene, Owens Lake intermittently overflowed its southern margin (sill elevation of 1,146 m) into a continuation of the Owens River that, in turn, fed China and Searles Lakes to the south (Gale, 1914; Smith, 1978). At times of overflow, the lake was as much as 75-m deep and had a surface area of 694 km^2 (Smith, 1978). A great deal of scientific study has been done at Searles Lake (Gale, 1914; Flint and Gale, 1958; Smith, 1962, 1978, 1979) and at Mono Lake (Russell, 1889; Lajoie, 1968; Newton, 1991), but little is known of the history of Owens Lake, except by inference from the history of Searles Lake (Smith, 1978). Matthes (1940) used early sediment core logs from Owens Lake to assess the timing of Pleistocene glaciation, and Smith and Pratt (1957) reported that Owens Lake contains more than 280 m of predominantly lacustrine sediments. The purpose of the current study is to better characterize the late Pleistocene and Holocene history of Owens Lake sediments.

Field and Laboratory Studies

Sediment cores were collected from the south-central part of Owens Lake on three occasions. In 1984 and 1987, a modified Livingston-type piston corer was used to collect one 12-m core (OWL84B in 1984) and six 4- to 5-m cores (OWL87A-F in 1987). In 1990, two 30-m cores (OWL90/1-2) were collected using a truck-mounted split-spoon corer. Core OWL84B was obtained closest to the center of the lake basin; all of the other cores were obtained about 2 to 3 km southwest of OWL84B and within 1.5 km of each other. Access to the lake playa for the coring operations was kindly provided by William McClung of Lake Minerals Corporation (Lone Pine, California).

All of the cores were stored in a refrigerated core-storage facility during subsequent laboratory analysis. All the cores were described physically and slabbed. Representative cores were X-rayed, and selected measurements of the organic/carbonate fraction (loss on ignition), grain size (Coulter counter), mineralogy (X-ray diffraction), and microfossils (diatoms and pollen) were made. Extensive rock-magnetic measurements (for example, magnetic susceptibility) also were made on these cores to help correlate them. All of these measurements form the basis for the core correlations and composite lake stratigraphy described in the next section. Five radiocarbon dates were obtained, and paleomagnetic secular-variation studies (Brandsma and others, 1989) were performed on selected cores to help determine a timeframe for the stratigraphy.

Lake Stratigraphy

All of the cores have a common stratigraphy, which broadly consists of four lithologic units (fig. 1). The units are, with increasing depth: Unit A - salt, Unit B - brown oolitic sand and silt, Unit C - brown muddy silt, and Unit D - black muddy silt. Below Unit D, the sediment sequence cycles between lithologies resembling Units C and D to a composite depth of 34 m below the surface (fig. 2) without reappearance of lithologies resembling Units A or B.

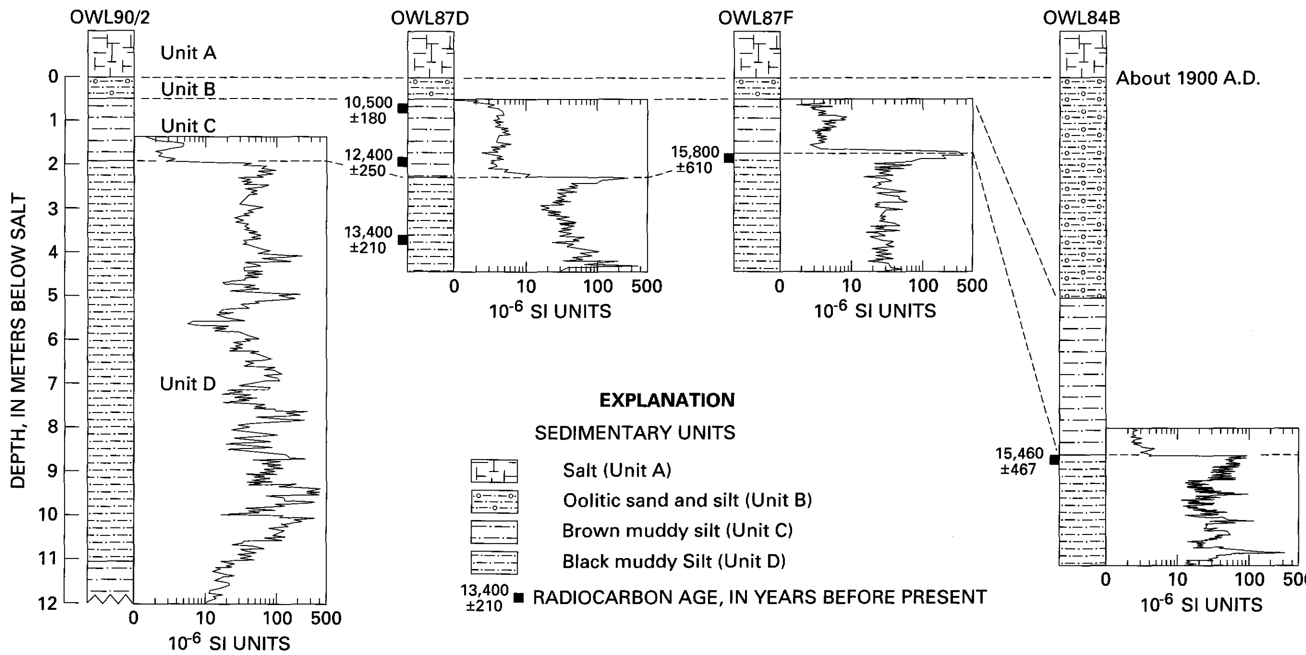


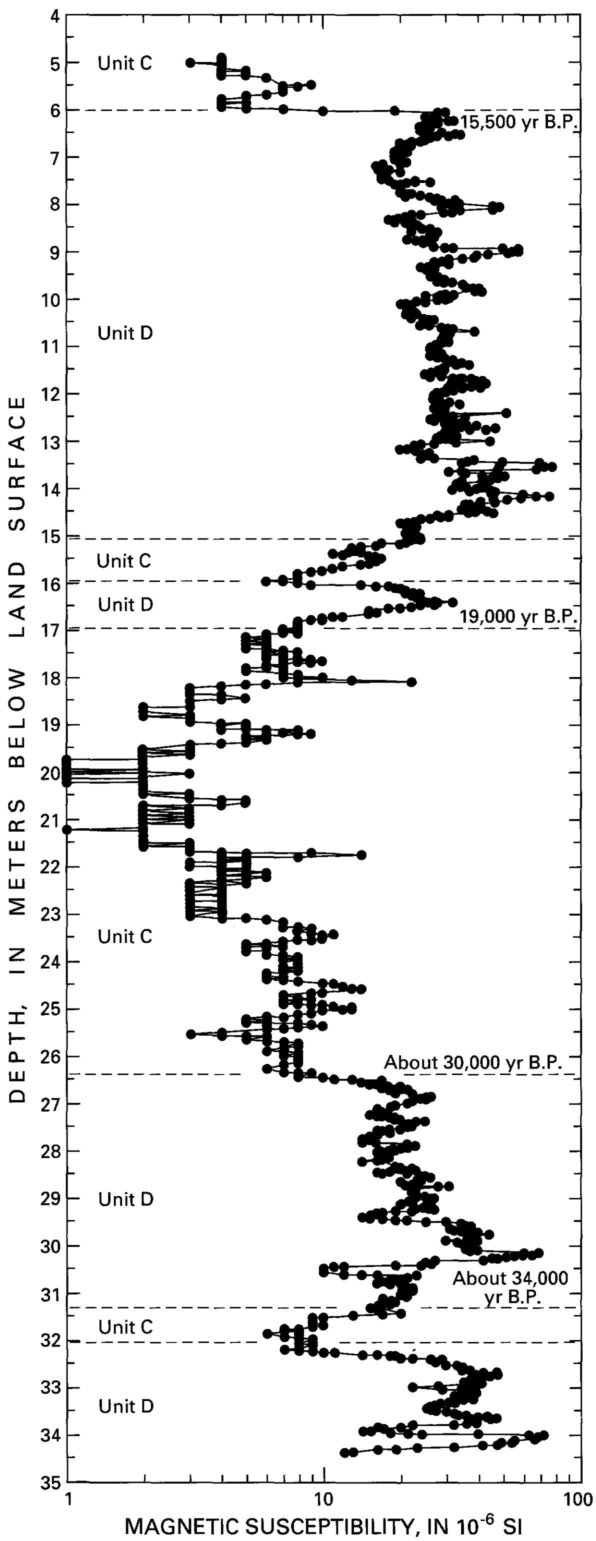
Figure 1. Correlation of four late Quaternary sediment cores collected from the south-central part of Owens Lake. Four lithologic units (A–D) have been described and correlated in all of the cores. Five radiocarbon age estimates (bulk organic fraction) have been recovered for initial age control. Magnetic-susceptibility profiles for the four cores are shown to the right of each lithostratigraphic column.

The uppermost lithologic unit (A - salt) is a crystalline layer of intergrown halite (NaCl), trona [$(Na_3(CO_3)(HCO_3)\cdot 2H_2O)$], and gaylussite [$(Na_2Ca(CO_3)_2\cdot 5H_2O)$] that contains smaller proportions of other common alkali carbonates (Smith and others, 1987). Unit A is generally 1 to 2 m thick in the area of the cores. The salt was deposited when Owens Lake desiccated after diversion of all Owens River water to the Los Angeles Aqueduct in 1913 (Gale, 1914).

Unit B (brown oolitic sand and silt) is characterized by alternating layers of coarse-grained (as much as 3-mm-thick) oolitic buff-tan sand; coarse- to medium-grained, well-sorted micaceous sand that contains tiny (0.5-mm) ooids; and dark-brown micaceous muddy silt. The large ooids have a dark amber core of aragonite and an outer white rind of high-Mg calcite (Newton, 1991). Unit B is generally less than or equal to 1 m thick in the cores, but is 5 m thick in core OWL84B, which is closest to the center of the lake basin. In core OWL84B, Unit B coarsens downward to a depth of 1.5 m below the salt; below this level, Unit B fines downward and contains intermittent fine-scale (millimeters) lamination and rare ostracods. Cross bedding is present, but is not an outstanding feature of this unit.

Unit C (brown muddy silt) is an olive-brown muddy silt (average grain size of 20 µm), which is rich in quartz and poor in clay minerals. Micritic calcite ($CaCO_3$) is detectable in thin section, and X-ray patterns indicate the presence of high-Mg calcite (Newton, 1991). The calcium carbonate consists of about 15 dry-weight percent of the sediment (fig. 3). Ostracods are rare in this unit and diatom variety is poor, except for numerous specimens of the diatom *Stephanodiscus niagarae*. Unit C generally is about 1.5 m thick in the cores, but is about 3 m thick in core OWL84B. The boundary between Units B and C seems to be gradational; grain size coarsens slightly at the top of Unit C. However, the boundary between Units C and D is quite different.

The base of Unit C in the cores is a sharp discontinuity characterized by five distinctive features: (1) A strong downward increase in magnetic susceptibility (figs. 1, 2); (2) an erosional (?) discontinuity that has as much as 1 cm of topography; (3) a 1- to 3-mm-thick lag deposit of coarse micaceous sand, containing purplish frosted quartz grains;



(4) evidence of mud cracks and rusty weathering penetrating downward into Unit D; and (5) substantially coarser average grain size in Unit C than in Unit D. This boundary is interpreted to be a hiatus caused by complete desiccation of Owens Lake. The duration of the event is not well known, but correlation of magnetic-susceptibility profiles and secular-variation records from all cores, indicated that the existing top of Unit D is essentially an isochron. Thus, any deflation associated with the desiccation most likely stripped away almost exactly the same amount of material from the lake surface for several square kilometers around the core sites. Alternatively, the isochronous top of Unit D, the thin lag deposits, and the minor evidence of erosion might indicate that the desiccation was relatively short-lived.

Unit D, the black muddy silt (pitch black on original core recovery with an H_2S odor), is a highly reduced, structureless muddy silt (average grain size of 10 μm) that quickly oxidized to light gray and contained orange speckles. The orange speckles are probably goethite (FeOOH), which replaced iron sulfide concretions when the silt was exposed to oxygen. The presence of greigite (FeS) in Unit D has been inferred from paleomagnetic studies, in which a strong magnetic overprint represented by a chemically unstable phase was observed. The mud fraction in Unit D contains some clay minerals but is dominated by quartz and is presumably rock flour derived from nearby Sierra Nevada glaciers, which fed the lake. Low-mg-calcite (Newton, 1991) averages about 7 dry-weight percent in the sediments. Unit D contains ostracods and a great abundance and diversity of diatoms typical of freshwater oligotrophic lakes. Pollen profiles in Unit D are notable for the presence of *Sequoiadendron* sp. pollen derived from Giant Redwood trees that now grow only west of the Sierra Nevada crest.

Below Unit D in cores OWL90/1 and OWL90/2, an oscillation between lithologies resembling Units C and D was reported to a depth of 34 m (fig. 2). Sharp boundaries in magnetic susceptibility below 20 m may indicate earlier desiccation events, but these intervals have not been studied well enough to assess this

Figure 2. Magnetic-susceptibility profile for core OWL90/2. High values of magnetic susceptibility can be associated with Unit D type lithology, whereas low values can be associated with Unit C type lithology. Preliminary age estimates, in years B.P., for the core are based on radiocarbon dating and paleomagnetic secular-variation studies.

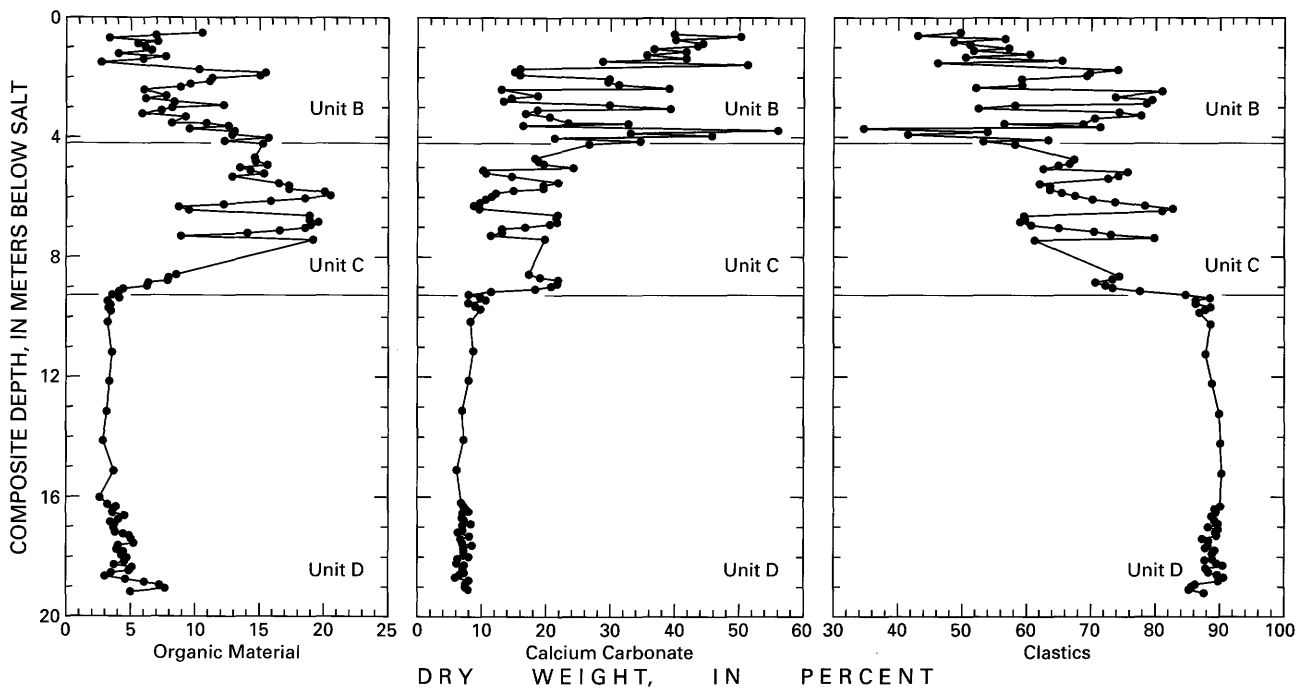


Figure 3. Summary plot of dry-weight percentages of organic material, calcium carbonate, and clastics in Units B through D from Owens Lake, California.

possibility. No evidence of lithologies resembling Units A or B below Unit D was observed in either of these cores.

Chronology

Five radiocarbon dates were recovered from bulk organic material in these cores (fig. 1). Two age determinations were originally made by Beta Laboratories, which returned almost identical ages for the top of Unit D in two different cores (15,460 yr B.P. for OWL84B; 15,800 yr B.P. for OWL87F). Three age determinations also were made on core OWL87D by the University of Arizona Radiocarbon Laboratory. However, their basal date within Unit D was 13,400 yr B.P., which is less than other Unit D age estimates. Two dates in Unit C were 10,500 yr B.P. near the Unit C/B boundary and 12,400 yr B.P. near the base of Unit C.

Limited paleomagnetic secular-variation (PSV) studies were done on Unit D, and the PSV records were correlated to other well-dated records from the Western United States (Brandsma and others, 1989; Lund and others, 1988). As discussed previously, the PSV records seemed to indicate that the top of Unit D is an isochron to within 100 to 200 yr everywhere in the study area. The age of this horizon is a bit ambiguous in that the PSV in this region broadly repeated its pattern every few thousand years (Lund and others, 1988). The PSV pattern at the top of Unit D could correlate with features dated elsewhere as either approximately 12,500 yr B.P. or approximately 15,000 yr B.P. Currently, the preferred PSV correlation indicated that the top of Unit D is approximately 15,000 yr B.P., and the base of Unit D in core OWL90/2 is approximately 19,500 yr B.P. The lower parts of the cores obtained in 1990 have not yet been studied.

Limited PSV studies on Unit C in core OWL84B were done. The preliminary results are consistent with the top of Unit C in this core being about 7,000 yr old. This age is substantially younger than the radiocarbon date in core OWL87D and could indicate some erosion near the top of Unit C (and in Unit B) in the 1987 and 1990 cores (which were obtained closer to the paleoshoreline).

Lake History

The presence of freshwater diatoms and the low concentration of $CaCO_3$ indicates that Unit D may have been deposited in a through-flowing permanent lake during late Pleistocene time (about 20,000 to 15,000 yr B.P.) that was associated with either steady-state glacial-maximum conditions or early retreat from glacial-maximum conditions. Sedimentation rate in the lake was quite high during this time interval (about 2 $m/10^3$ yr) due to the heavy flux of glacial outwash (rock flour) from the neighboring Sierra Nevada glaciers. The lake could have been holomictic because there is ample X-ray evidence of bioturbation in the bottom sediments even though the sediments were reduced below the sediment/water interface. This interval is broadly correlative to the Parting Mud interval in Searles Lake (Flint and Gale, 1958; Smith, 1962).

The shift to a desiccation surface at the top of Unit D is quite abrupt. However, the uppermost meter of Unit D does contain some evidence for a coarsening in average grain size and an increase in silt fraction, which might portend the desiccation. Unless there has been some uniform deflation of Unit D sediments, this desiccation may have occurred about 15,000 yr B.P. (The alternative, that the desiccation occurred about 12,500 yr B.P., requires that the replicate radiocarbon dates be incorrect.) This desiccation could be correlated to a substantial change in the Parting Mud sediments at Searles Lake at about 15,000 yr B.P. (Smith and Street-Perrott, 1983; Benson and others, 1990).

Unit C indicates the return of a mostly permanent lake after the desiccation and after recession of the main Sierra Nevada glaciers. The much coarser sediment, higher $CaCO_3$, lower sedimentation rates (?), and oxic sediment indicated that Owens Lake was a closed-basin lake that only intermittently rose to its sill and overflowed to the south. The radiocarbon dates indicated that this lake was present during the Pleistocene/Holocene transition (approximately 12,000 to 10,000 yr B.P.) and was correlative with the uppermost Parting Mud at Searles Lake. There is some evidence in core OWL84B that this lake persisted into early Holocene time (to about 7,000 yr B.P.?), which would be consistent with recent evidence from Mono Lake that a permanent, relatively deep lake existed at Owens Lake from 9,000 to 6,500 yr B.P.

Unit B represents the middle to late Holocene interval in Owens Lake history. The carbonate ooids, coarse-grained sediments, and intermittent cross stratification indicated that the lake was still a closed-basin lake, but its average lake level was probably much lower, and frequent desiccations (a true playa lake) occurred. The lake may have become saturated with various alkali minerals that were likely precipitated out during each desiccation and redissolved during the next wetting, until the alkali minerals were finally precipitated as Unit A in 1913.

The alternating pattern of Unit C and D type sedimentation below Unit D in the 1990 OWL cores (described previously) probably represented an alternation in the persistence of water flow in the Owens Lake drainage basin. This alternation could be related to temperature, precipitation variation, or the ebb and flow of continental glaciation. The study of these sediments is in the initial stages. If the sediment history at Owens Lake is similar to Searles Lake (Smith, 1962), Unit D sediments below 25 m in core OWL90/2 might correlate with mud stringers in the Lower Salt at Searles Lake. If the sediments do correlate, then the base of core OWL90/2 is probably about 34,000 yr old.

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Holocene Lake-Level Variations and Eolian Activity in the Oregon Great Basin

Patricia F. McDowell, Department of Geography, University of Oregon
Daniel P. Dugas, Intermountain Research, Silver City, Nevada

A preliminary chronology of lake-level variation was compiled, using the data available from the Fort Rock, Chewaucan, Alkali, and Harney Basins, Oregon. Drying of the large paleolakes that occupied the basins between approximately 20,000 and 13,000 yr B.P. (isotope stage 2) set the stage for Holocene events. The youngest lacustrine radiocarbon date for the Fort Rock Basin is 11,500 yr B.P. (McDowell and Benjamin, 1991). A radiocarbon date for camelid bone from the same area indicates that the Fort Rock Basin was dry or almost dry by 10,300 yr B.P. There are no direct dates on the drying of the large paleolakes in Alkali, Chewaucan, or Harney Basins. Archaeological evidence indicates that Alkali Basin was dry or almost dry by about 11,000 yr B.P. (Willig, 1988a).

There is no firm evidence from Oregon for a lake-level rise during the Younger Dryas (11,500 to 10,000 yr B.P.) determined for the Lahontan Basin, Nevada, and the Bonneville Basin, Utah (Currey, 1988). The Alkali, Chewaucan, and Harney Basins had moderate to high lake-level rises during the early Holocene. Dates from the Harney Basin indicate that, at 9,600 yr B.P., Harney and Malheur Lakes coalesced and stood at approximately 3 m below their highest shoreline, and at 8,900 yr B.P., the combined lake stood at its highest shoreline and overflowed (Gehr and Newman, 1978; Gehr, 1980). Dugas (1993) recently obtained dates indicating a moderate lake rise in the Harney Basin at about 9,500 to 9,300 yr B.P. In Alkali Basin, a date of 9,800 yr B.P. was obtained on ostracod shells. In addition, S.W. Robinson obtained a radiocarbon date of 9,400 yr B.P. from "tufa-like carbonates on beach gravels" that may represent a moderately low lake level in the Lake Abert subbasin of the Chewaucan Basin (cited in Gehr, 1980, p. 74–75). There is additional undated stratigraphic evidence indicating the presence of early Holocene lakes or marshes in Alkali and Chewaucan Basins. Based on archaeological evidence, Willig (1988a) proposed that shallow water bodies developed in Alkali Basin three times after 10,000 yr B.P. These lake rises are not radiometrically dated, but Willig (1988b) proposed that the first two rises occurred between 10,800 and 7,000 yr B.P. After the isotope stage 2 lake in the Chewaucan Basin declined to a low level, an extensive lake, called the ZX Lake, reformed (Allison, 1982). The ZX lake stand is not radiometrically dated, but its relation to archaeological sites (Oetting, 1989) indicates that it probably formed and then desiccated before 11,000 to 8,000 yr B.P.

The middle Holocene record indicates that dryness and eolian activity dominated. In the Fort Rock Basin, eolian sands overlying paleolake deposits dated from 6,800 yr B.P. and younger (McDowell and Benjamin, 1991). Dunes in the Harney Basin were actively building just after 5,000 yr B.P. (Dugas, 1993), and intermittently during the late Holocene. Sand dunes elsewhere in the basins generally had very weak soil development and appear to be no older than middle to late Holocene age. Evidence from archaeological sites supports a moister period in the late Holocene, called the neo-pluvial period by Allison (1979). Pettigrew (1985), Oetting (1988), and Willig (1988b) reported evidence in several sites on the east shore of Lake Abert of a lake rise to about 6 m above the modern average lake level. Allison (1979, 1982) reported low undated beach lines in several places around the Chewaucan and Fort Rock Basins that he attributed to the neo-pluvial period.

The evidence for a dry landscape and for eolian activity in the middle Holocene, and possibly wetter conditions in the late Holocene, is consistent with other Holocene records from this region. The pollen record from Diamond Pond indicates dry conditions from 6,000 to 5,400 yr B.P., moistest conditions at 4,000 to 2,000 yr B.P., and somewhat drier conditions with short-term fluctuations after 2,000 yr B.P. (Wigand, 1987). Sand dunes were active in the Catlow Basin in the middle Holocene, and several alternating episodes of eolian activity and soil development occurred during the last 3,400 yr B.P. (Mehringer and Wigand, 1986).

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Sedimentary Analyses from the Owens Lake Core and Their Implications for Climate Change

Kirsten M. Menking, Hannah M. Musler, and Robert S. Anderson, University of California
James L. Bischoff and Jeffrey P. Fitts, U.S. Geological Survey

Sedimentary analyses of a 323-m core from Owens Lake, in eastern California, indicated lithologic variations on many different time scales. Grain-size analysis, organic carbon content, $CaCO_3$ content, and water content were determined from point samples taken at 2- to 3-m intervals down the core. In addition, sand, silt, and clay percentages were measured on 3.5-m-long channel samples to supplement point-sample grain-size data. Clay mineralogy of these channel samples also was determined.

Grain-size analysis of point samples defined two distinctive depositional regimes (fig. 1). Except for a few thinly bedded coarse-grained layers at the top of the core, mean grain size fluctuated between 0.005 and 0.015 mm (clay- to silt-sized material) between 7 and 195 m in depth. In contrast, between 195 m and the base of the core at 323 m, mean grain size generally fluctuated between 0.01 mm (medium fine silt) and 0.05 to 0.1 mm (coarse silt to fine sand). Coarse-grained sediments tend to be poorly sorted, whereas silts and clays are usually much better sorted. Sand plus gravel, silt, and clay contents (measured as dry-weight percents) of 3.5-m-long channel samples broadly mimicked the point-sample trends; coarse-grained material was present in larger percentages between 190 and 323 m, whereas fine silts and clays predominated from the top of the core to 190 m.

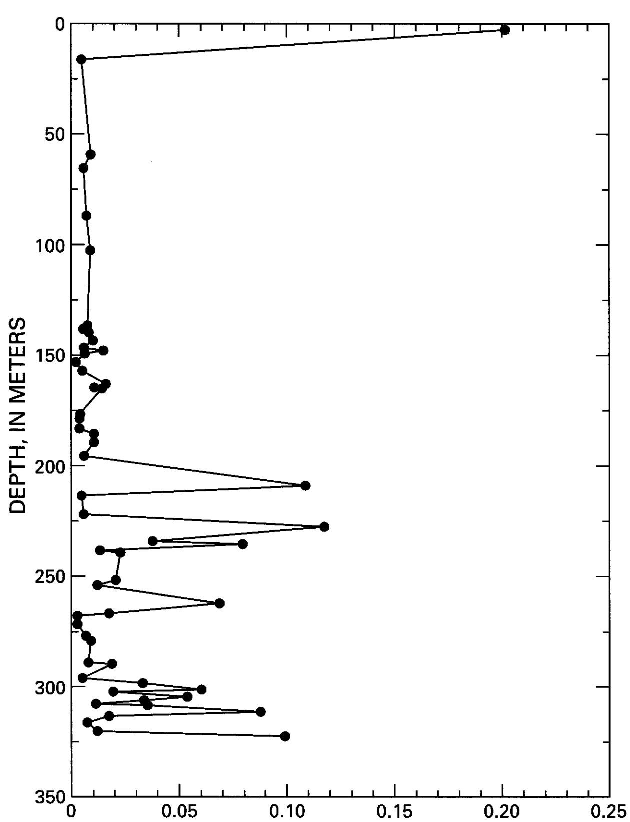

Figure 1. Mean grain size as a function of depth in core OL-92 from Owens Lake, California.

X-ray diffraction determinations of the clay mineralogy of channel samples indicated that illite, smectite, and chlorite were the primary clay minerals; some samples contained abundant clay-sized quartz, plagioclase, and potassium (K) feldspar. Peak areas of illite, smectite, chlorite, quartz, plagioclase, and K-feldspar were measured, and ratios between the minerals were determined for each channel sample. Ratios then were compared among samples. Smectite/quartz and plagioclase/K-feldspar varied inversely; smectite/quartz had a low value during the last full glacial maximum (20,000 yr B.P.), whereas plagioclase/K-feldspar was high. Smectite/quartz, illite/quartz, and chlorite/quartz ratios were similar; all ratios had high values at the same depths. Likewise, K-feldspar/quartz and plagioclase/quartz ratios varied similarly.

Carbonate content of point samples varied cyclically between 0 and 56 dry-weight percent and averaged 14 dry-weight percent between 0 and 225 m. From 225 to 275 m, percent carbonate remained at zero, whereas from 275 m to the base of the core, carbonate content again showed cyclic variations. Organic carbon content had very similar oscillations between 0 and 220 m; the organic carbon varied between 0 and 3 dry-weight percent. Between 220 and 260 m, organic carbon content was essentially 0 percent, and from 260 m to the base of the core, it oscillated between 6 and nearly 0 percent with an average value of 1 percent.

Five- to ten-gram subsamples of the grain-size point samples were weighed and then heated to drive off pore fluids. The dried samples were reweighed, and the water contents were calculated. With the exception of the top 10 m of the core, water content decreased from almost 65 percent at the top of the core to about 13 percent at 240 m. Water content increased to a high of 54 percent at 290 m and then decreased to 10 percent at the base of the core.

The various data sets bring up a number of questions. The change from interbedded sands and silts in the lower one-third of the core to sediments consisting primarily of silt and clay in the upper two-thirds of the core occurred at the 195-m depth. Given an average sedimentation rate of 0.4 mm/yr, this depth corresponds to an age of 488,000 yr. In comparison, Smith (1984) reported a major shift in paleohydrologic regime in Searles Lake from dry to intermediate shortly after 600,000 yr B.P. Although these two dates are quite different, the Owens Lake core currently (1993) has poor age control, making it difficult to determine whether the two lakes had fluctuations on the same time scales. If the shifts in paleohydrologic regime (in Searles Lake) and in grain size (in Owens Lake) were synchronous, the lower, sandy section of the Owens Lake core may represent a shallow lake or a fluvial environment that existed in a drier climate. However, the change in depositional style at 195 m also may have occurred as the result of tectonic factors, namely valley deepening or uplift of the spillover sill with respect to the top of the lake, causing Owens Lake to be volumetrically much larger before spillover after about 500,000 yr B.P. If not the result of ground-water-diffusive processes, the presence of the zero carbonate and organic carbon zones between 220 and 275 m may help to constrain which scenario was more likely. The lack of carbonate and organic carbon may indicate a very fresh, nutrient poor, unproductive lake, which implies that Owens Lake was spilling over into the China and Searles Lake Basins.

As does Newton (1991), we interpret the presence of nonclay minerals (quartz, plagioclase, and K-feldspar) in the clay-sized fraction to be the result of glacial abrasion in the Sierra Nevada. Of the true clay minerals, illite and chlorite may be recycled from metasedimentary roof pendants, but smectite is more likely produced from the weathering of feldspars during soil formation and may, therefore, be more indicative of climates conducive to chemical weathering. Smectite also is a common weathering product of volcanic ash, so care needs to be taken to distinguish between true climatic variations and volcanic events.

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Intrabasinal and Extrabasinal Correlations of Nonmarine High-Resolution Paleomagnetic Records of the Middle to Late Pleistocene from South-Central Oregon

Robert M. Negrini, Department of Physics and Geology, California State University, Bakersfield

High-resolution age control is a serious problem for those studying climate records associated with middle to late Pleistocene lake sediments. One possible solution is provided by paleomagnetic secular variation (PSV), the low-amplitude variation in the direction and intensity of the magnetic field of the Earth. If recorded accurately at or near the time of deposition of a sediment, PSV is a high-resolution correlation tool that can be used for intrabasinal and extrabasinal correlations, especially when complemented with independent methods of correlation (for example, tephrochronology). Theoretically, PSV could enable nonmarine stratigraphy to be correlated to the well-dated chronology common to marine sediment records. Examples of correlations based on this technique are described for ongoing paleomagnetic and lithostratigraphic studies of the sediments of Pleistocene Lakes Chewaucan and Lahontan and on previously published paleomagnetic records from Pleistocene Lake Russell, two unnamed Pleistocene lakes, and the Gulf of California.

Intrabasinal correlations in Lake Chewaucan support the theory that secular variation can be used to correlate outcrop exposures (three samples per horizon) on a depth scale of several centimeters, which probably corresponds to a temporal resolution of a few hundred years or less. Correlations between outcrop and core can still be made, but the resolution is somewhat lower. Extrabasinal nonmarine correlations are consistent with (1) low surface levels in Lakes Chewaucan and Lahontan between 27,000 and 24,000 yr B.P., (2) an age for the last Lahontan highstand of less than or equal to 15,500 yr B.P., and (3) the existence of three Great Basin lakes that are 100,000 to 200,000 yr old and that are separated by hundreds of kilometers. Preliminary high-resolution correlations with the marine record based on directional variations were developed for a period from approximately 50,000 to 20,000 yr B.P. Preliminary correlations with the marine record of lower resolution based on paleointensity variations were developed for a period from 100,000 to 50,000 yr B.P.

Paleoclimatic Inferences from Detailed Studies of Lake Bonneville Marl, Utah

Charles G. Oviatt, Department of Geology, Kansas State University

Marl was the littoral-to-pelagic facies deposited in Lake Bonneville at sites distant from sources of clastic input. Gilbert (1890) referred to the marl as the white marl and reported its presence throughout the lake basin as a thin blanket deposit. Certain features of the white marl are traceable throughout wide areas of the basin and provide a means of correlation and chronologic control. The base of the white marl is time-transgressive with elevation, and its age can be inferred accurately from the many radiocarbon ages that date the transgression of the lake in the basin (Oviatt and others, 1992). The age of the top of the marl can be inferred in a similar way, although the top is truncated to some extent in most sections. Within the white marl, a prominent stratigraphic marker bed that is characterized by an abrupt change from massive gray marl below to laminated sandy marl above is present in most exposed sections. This bed was interpreted as representing the drastic change in water depth caused by the Bonneville Flood, during which the lake dropped approximately 100 m in a few months and exposed thousands of square kilometers of former lake bottom to subaerial erosion and resedimentation in the lake. The flood is dated at about 14,500 yr B.P., and the flood bed, therefore, is essentially an isochronous stratigraphic unit that can be correlated basinwide. Ostracod biostratigraphy provided another means of correlating isolated marl outcrops; the ostracod zones are traceable basinwide and have been dated in a core from the bottom of Great Salt Lake (Thompson and others, 1990).

Basaltic volcanic ashes provided additional correlation and chronologic control. In the Sevier arm of Lake Bonneville, the Tabernacle Hill ash (14,300 yr B.P.) was erupted while the lake was overflowing at the Provo shoreline, and the Pahvant Butte ash (15,500 yr B.P.) marks a point in the lake history when the lake was within 15 m of the Bonneville shoreline (Oviatt and Nash, 1989). In the northern Bonneville Basin, the Thiokol basaltic ash, which was erupted from an unidentified source about 26,000 yr B.P., is widespread and is a valuable marker in the lower part of the marl.

The marl ranges in thickness from 2 to 3 m, depending on elevation and local controls on clastic input. Estimated average marl sedimentation rates during the transgressive phase of the lake ranged between 15 and 20 cm/10^3 yr. During the deepest water phase, deposition was as low as 1 to 2 cm/10^3 yr.

Measurements of the ratio of aragonite to calcite, total percent calcium carbonate, $\delta^{18}O$, and $\delta^{13}C$ in cores of the marl can be interpreted as records of lake-level change in the Bonneville Basin. These four variables covaried and responded in a similar way to changes in the ratio of evaporation to inflow in the lake. Collectively, they provided a record of water chemistry and, hence, of water-volume or lake-level change through time. A core of marl from the Old River Bed area, at the north end of the strait that connected the Sevier arm with the main body of Lake Bonneville, has yielded a detailed record of changes for about 5,000 yr of the transgressive phase of the lake (about 20,000 to 15,000 yr B.P.). Lake-level fluctuations that had probable amplitudes on the order of tens of meters are inferred from the carbonate record and are superimposed on the generally rising trend of the transgressive phase of the lake. Because $\delta^{18}O$ and $\delta^{13}C$ are highly correlated ($r^2 = 0.97$), there may not have been a major shift in isotopic composition of the lake water during this part of the transgressive phase of the lake. However, comparison with the isotopic data from carbonates of Holocene Great Salt Lake (Spencer and others, 1984; McKenzie, 1985) indicated a possible change in moisture source between the Pleistocene and Holocene.

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Marine-Terrestrial Correlation as a Means of Testing Global Climate Change

Milan J. Pavich, U.S. Geological Survey

Introduction—The Correlation Problem

Correlation is used as a fundamental approach to understanding the timing and magnitude of geologic change. Correlations, however, are often not rigorously tested, nor are they tied to an understanding of processes. One of the fundamental working hypotheses about climate change, the Milankovitch (1930) orbital-forcing model, relies on correlation of marine oxygen isotope records as a principal test. In general, the present (1993) state of testing used in marine geology is considered sufficient support for the Milankovitch model. However, questions arise as to the sufficiency of the marine-record correlations because few well-dated terrestrial records are available. As a result, it is argued that any model of global climate needs to be tested against the terrestrial record of climate change as well as the marine record. The correlation of marine and terrestrial records and the independent dating of terrestrial records cause doubt about the acceptance of orbital forcing as the fundamental cause of climate change. This study reviews some of the recent contributions to that doubt.

The interpretation of rates and magnitudes of future global changes needs to be put in the context of the geologic record. There are many possible records for different time scales to choose as the frame of reference. The widely accepted Milankovitch model of climatic cycles is based on orbital-forced variations in solar insolation. Change in the orbital geometry of the Earth is considered the fundamental forcing that is amplified by the subsequent changes in the albedo of the surface of the Earth and by changes in the concentration of greenhouse gases to produce climate variations (Berger and Loutre, 1992). However, as indicated recently by the Devils Hole $\delta^{18}O$ record (Winograd and others, 1992), serious problems result when independently dated terrestrial records are compared with marine records to test the astronomical/insolation models. Winograd and others (1992) also indicated that models that related isotopic variations in marine deposits to climatic forcing functions of regular periodicity, such as insolation, failed to account for other types of aperiodic, climatically related deposits. Thus, models and predictions using models need to be rigorously tested against observations of various types of stratigraphic systems.

The accuracy and precision of model predictions needed to satisfy policy questions related to the greenhouse gas/warming issue are not improved by assumed or forced correlations. The precision of predictions can be no better than the precision of measurement of paleo-parameters used to validate models. Presently, very few reliable, independently dated, high-resolution records of climate change are available. Without such records, the assumed correlations are of little objective value.

In the Western United States, marine and terrestrial Quaternary deposits that are in close proximity can be compared. Theoretically, careful stratigraphic analysis (discussed in numerous contributions in this volume) should yield records that allow for assessment of specific climate changes and climatic forcing functions. In practice, however, fundamental differences in the response of the marine and terrestrial systems to climate change create complications in the comparison of records.

Low-Resolution Marine Records

Marine sediments provide a repository of climate-related records. In areas of slow sedimentation and anoxic conditions, such as in deeper, isolated basins, rhythmites or varved sediments are preserved. The layering of these sediments indicates passive settling of both calcareous and siliceous components. The isotopic records in these sediments contain time series of climate change.

Many marine records seem to correlate, but lack, independent dating controls. The methods used by Imbrie and others (1984) and Shackelton and others (1990) relied on comparing various stacked records as a means of correlation and validation of the orbital-forcing hypothesis. If the same sequence was seen repeatedly, then the original hypothesis was accepted to be correct; see, for example, Shackelton and others (1990) who argued that the matching of stacked marine records, especially the discrete substages of Stage 5, verified the orbital forcing of variations of global ice volume. Shackelton and others (1990) further concluded that the marine record verified the hypothesis of orbital forcing of climate change described by Milankovitch (1930).

The ODP site 677 in the Pacific Ocean contains one of the few continuous, low-latitude records of climate change. Shackleton and others (1990) observed in the interval of 800,000 to 620,000 yr B.P. a transition during which the climate system changed from a predominant 41,000-yr periodicity attributed to obliquity forcing (Ruddiman and others, 1986) to a predominant 100,000-yr periodicity associated with the eccentricity cycle (Imbrie and others, 1984). Shackelton and others (1990) also developed a time scale back to 2,600,000 yr based on the assumption of a consistent phase relation between astronomical forcing and climatic (that is, ice volume) response. Shackelton and others (1990) and Hilgen (1991) concluded that time scales calculated from precession and eccentricity parameters provided improved dates of paleomagnetic time markers. For example, Imbrie and others (1984) had predicted the Brunhes-Matuyama boundary at about 730,000 yr B.P.; however, the revised calculations of Shackelton and others (1990) predicted the boundary to be closer to 780,000 yr B.P. Recent K/Ar dating of some paleomagnetic reversals (Baksi and others, 1992; Izett and Obradovich, 1992) supported the revised time scale derived using calculations from Shackleton and others (1990).

The apparent synchroneity of low-resolution, marine-sediment records with orbital cycles is strong evidence that such global-scale hydrodynamic and isotopic responses to solar insolation variation are real. The question remains, however, whether major climate shifts, such as glaciations, are initiated and driven by orbital variations. The Milankovitch theory presumes primary forcing by a low-magnitude variation of solar insolation. If correct, the climatic processes that respond to solar insolation variation (such as ice-volume change) seem to operate through various amplifying feedbacks after initially responding to a low-amplitude (or energy) forcing. In responding to low-amplitude forcing, the various subsystems that comprise the climatic system have minimal inertia across gradients of heat transport (from the sun to the atmosphere, from the atmosphere to the ocean, and from low to high latitude areas) that are all relatively simple in fluid media. This minimal inertia contrasts markedly with the various negative feedbacks that apparently operated in the atmosphere, hydrosphere, and biosphere to keep the pre-Quaternary climate of the Earth fairly constant during long periods (greater than 10^8 yr) of geologic time.

High-Resolution Marine Records

Results from marine-terrestrial comparisons also produced questions about the Milankovitch model. A simple determinative model would predict synchronous changes across climatic boundaries. Analysis of higher resolution records casts doubt, however, on the existence of such synchroneity. Analysis of higher resolution, marine-terrestrial records indicated problems inherent in comparing marine-terrestrial responses to climate change.

Gardner and others (1988) compared a nearshore continental slope record with the pollen record of Clear Lake, California. Comparison of these records indicated only a rough similarity in climate change during the past 20,000 yr. Correlation between the marine and terrestrial records required discarding 9 of 10 ^{14}C dates from the Clear Lake record. Attempts to force a correlation between the two records did not prove entirely satisfactory. Pollen data from the two cores indicated synchroneity between presumably coeval peaks; for example, a discrepancy of about 2,000 yr in the ages of initial increases in *Quercus* and decreases in *Pinus*. Despite marine diatom and organic carbon evidence for upwelling of cold coastal waters at 15,000 yr B.P., the terrestrial stratigraphy indicated no increase in *Sequoia*, a fog-dependent tree, until 12,000 yr B.P. At 12,000 yr B.P., there was an anomalous contrast of decreasing upwelling and increase in *Sequoia*. These unexplained phenomena indicate the problems of correlating even short distances of datable, high-resolution climate records. The same problem is well represented in comparisons of terrestrial records.

Terrestrial Records

Lacustrine

Terrestrial lacustrine sediments, by contrast with marine sediments, do not often continuously record changes in climate. There may be no single lacustrine record that matches continuous marine records. Therefore, records from several basins are needed to compile a quasi-continuous record of sedimentation that allows for paleoclimatic interpretation.

Bradbury (1992) published a climatic interpretation for the northern Great Basin (Tule Lake, California). He reported that the lower 100 m of the Tule Lake core (which was deposited between 3,000,000 and 2,500,000 yr B.P.) contained diatom concentrations that indicated a 41,000-yr cycle of diatom productivity. He also reported a 400,000-yr cyclicity, similar to the eccentricity cycle, between 2,900,000 and 1,400,000 yr B.P. During the past 10^6 yr, six major peaks of diatom concentration occurred that were indicative of 100,000-yr cycles. Bradbury (1992) reported, however, that correlation with the marine record was untested due to the lack of age control in both records.

Paleomagnetic measurements can provide an objective technique of testing correlations. Paleomagnetic measurements in the Tecopa Basin by E.E. Larson and others (University of Colorado, written commun., 1993) yielded evidence of regular cycles in lacustrine sediments. Larson and others observed a rhythmically spiked magnetic-susceptibility signal (dominated by nine 41,000-yr periodicities) for the Olduvai Subchron (1,700,000 to 1,860,000 yr B.P.) that is very similar in character to the jagged variations in marine oxygen-isotope values.

Above the Olduvai Subchron, E.E. Larson and others (written commun., 1993) found evidence for a strong 100,000-yr periodicity and lesser periods of 40,000, 22,000, 15,000 to 16,000, 12,000, 9,000, and 8,000 yr. Larson and others conclude that these periodicities are climatic and synchronous with astronomically forced processes. In addition, Larson and others compared the Tecopa Basin record with the record from Searles Lake (Smith, 1979, 1984; Jannik and others, 1991), stating that there was a strong similarity in the long-term trends in both basins. However, Larson and others also reported that, in these closely spaced basins, extended dry periods were not coeval; for example, Tecopa Basin was relatively dry from 1,600,000 to 1,000,000 yr B.P., but Searles Lake was dry from 2,000,000 to 1,300,000 yr B.P.

Jannik and others (1991) found some evidence of 100,000-yr cyclicity in Searles Lake sediments, but they also found evidence for lower frequency variations. They reported a regular alternation of evaporites and muds prior to about 700,000 yr B.P., during which the period of lake-level fluctuations was approximately 50,000 yr. During the shift from dominance by obliquity frequency to eccentricity frequency reported by Shackelton and others (1990), Searles Lake fluctuations changed from approximately 50,000 yr to a longer period. Jannik and others (1991) suggested that these fluctuations may correspond to the marine record, but there was not a one-to-one match between the Searles stratigraphy and the marine ^{18}O record. Moreover, Smith (1984) described a 400,000-yr period in the major hydrologic regimes of Searles Lake that was not strongly expressed in the marine record.

Devils Hole

Winograd and others (1992) discussed the results from a core (DH-11) of vein calcite from Devils Hole, Nevada. The 500,000-yr record of paleotemperature contained in that well-dated core indicated that major deglaciations, also recorded in marine sediments and ice cores, do not correlate, as previously hypothesized, with solar insolation peaks. Specifically, Winograd and others (1992) cited convincing evidence that the major insolation maximum at 128,000 yr B.P. did not trigger a major deglaciation. Furthermore, they argued that the DH-11 record indicated increasing durations of glacial periods since 500,000 yr B.P. They stated that the increase in the duration of glacial periods from about 130,000 to 80,000 yr (since the middle to late Pleistocene) indicated aperiodic climate shifts. The overall conclusion from their study was that, because the Milankovitch model failed to account for major terminations and the aperiodic duration of glacial periods, astronomical forcing has been over-interpreted as a control of climate change.

Future Research Needs

At present, the need for independent dating of high-resolution marine and terrestrial records cannot be overstressed. Process-based studies of lacustrine response to climate change at the scale of single drainage systems, such as the Owens River drainage studied by Jannik and others (1991), also are of high priority. Until basin response to climate change is understood, a broad picture of the various forces and feedbacks involved in regional- and global-scale climate cycles cannot be synthesized. The comparison of records will probably continue to indicate that the records are different in detail. Simplistically, these differences indicate that global-scale changes were not uniformly encoded in the geologic record and that establishing global records from lacustrine sequences will be limited. However, the differences in system responses will be most significant in understanding the processes that cause change in different systems and in different regions of the globe.

A promising new area of research that may provide more data about global geophysical process uses magnetic susceptibility and paleomagnetic field strength as a means of correlation. Mankinen and Champion (1993) described low-dipole field strength for volcanic rocks from Hawaii from 40,000 to 18,000 yr B.P., which was followed by a rapid rise to present strength values. Meynadier and others (1992) studied the intensity variations for the past 140,000 yr in the Somali Basin, where low intensities are recorded for long periods around 40,000 and 115,000 yr B.P., both of which were cold periods. Tric and others (1992) described a halving of average field intensity, relative to the present, between 142,000 and 80,000 yr B.P.—a period that contained a major cold (140,000 yr B.P.) period and a major warm (135,000–120,000 yr B.P.) period. Tric and others (1992) also noted a possibly stronger field intensity at 166,000 yr B.P., which was another cold period. Present data are not sufficient to determine whether the paleointensity records varied in phase with climate cycles. The present evidence does indicate, however, that even if the periodicities of variation are similar, the directions of change are not consistent.

Another intriguing, but enigmatic, observation from recently published paleomagnetic intensity measurements (Tric and others, 1992) indicates a quasi-24,000-yr cyclicity. This cyclicity cannot be directly attributed to solar insolation and may indicate some other geophysical effect, such as precessional torque that produced cycles that mimicked solar insolation cycles. If further study of the paleointensity record does reveal a quasi-24,000-yr cyclicity, it may indicate that there are nonclimatic geophysical parameters related to orbital precession that are covariant with climate cycles. If so, then the solar insolation cycles also may be noncausally covariant with climate.

Available information is insufficient to resolve the conflict between marine and terrestrial stratigraphy. Even if the marine stratigraphy were sufficient verification of the Milankovitch model, regional climatic models for land areas simply cannot be generated using present understanding. Many independently dated terrestrial records need to be studied to gain a better understanding of the processes that alter climate-related sediment characteristics. An example of the simplicity of misinterpretation is the early interpretations of the correlation between paleomagnetic intensity of sediments and isotopic composition of sediments (Oldfield and Robinson, 1985). Without more detailed analysis, researchers would have been satisfied with the conclusion that variations of magnetic field intensity were a cause of paleoclimate variability. When numerous parameters correlate, noncorrelative systems are needed to test the models and to test ideas about causality in complex feedback systems. The sometimes tedious process of recovering and analyzing numerous terrestrial records can provide critical tests of climate models.

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Calcium Carbonate Formation in Pyramid Lake, Nevada

Michael M. Reddy, U.S. Geological Survey

Calcium carbonate ($CaCO_3$) mineral formation occurs in numerous, temperate hardwater lakes of the United States. Carbonate minerals may serve as paleoclimate records if, as the minerals form, trace elements and certain isotopes are incorporated in amounts reflecting environmental conditions of the surrounding lake water. Well-studied lakes having this phenomenon include Fayetteville Green Lake and Onondaga Lake, both located near Syracuse, New York. Pyramid Lake in Nevada, in contrast, has unique physical and chemical characteristics that modify in-lake calcium carbonate formation processes from those observed in the lakes in New York (Galat and others, 1981; Benson and others, 1990; Benson, 1993). Some of these modifications include: (1) The occurrence of large carbonate mounds in the nearshore region of Pyramid Lake (Benson, 1993), which are not usually found in more humid environments; (2) the formation of a less stable form of calcium carbonate (aragonite) rather than the thermodynamically stable polymorph, calcite (Galat and Jacobsen, 1985); and (3) the occurrence of calcium carbonate precipitation episodes (termed "whitings") in the Pyramid Lake water column (Galat and Jacobsen, 1985).

These modifications indicate that calcium carbonate precipitation in Pyramid Lake is affected by chemical factors that control calcium carbonate supersaturation levels and carbonate mineral growth rates (Nancollas and Reddy, 1971; Reddy, 1975, 1978, 1986; Reddy and Nancollas, 1976; Reddy and Wang, 1980; Reddy and Gaillard, 1981; Reddy and others, 1981). Laboratory and in-lake calcium carbonate formation studies at Pyramid Lake are in progress to test this hypothesis. Field studies of calcium carbonate formation in Pyramid Lake were initiated in the summer of 1992; laboratory studies of calcium carbonate formation are continuing.

Calcium carbonate formation in a test tube or a lake is frequently initiated by crystal nucleation involving formation of unstable polymorphs and hydrates (Reddy and Nancollas, 1976; Reddy, 1986). Calcium carbonate nucleation rates, which differ among these polymorphs and hydrates, increase with increasing calcium carbonate supersaturation. Supersaturation is defined as the ratio of calcium carbonate ion-activity product to the calcium carbonate (calcite) solubility product (Reddy, 1986). The sensitivity of nucleation rates to calcium carbonate supersaturation confounds attempts to predict the onset and duration of calcium carbonate formation episodes in Pyramid Lake (Galat and Jacobsen, 1985).

At a fixed calcium carbonate supersaturation level, water temperature and concentration of certain ions (termed "inhibitors") can affect carbonate mineral polymorphic composition and formation rates (Reddy, 1977, 1978). Following nucleation, or in the presence of a suitable growth surface, calcium carbonate minerals may form from a supersaturated solution. Calcite formation under such conditions follows a parabolic growth-rate equation (Reddy, 1986). This equation relates the crystal growth rate to the product of three terms: the growth surface available for reaction, the crystal growth-rate constant, and the square of the reaction driving force.

Inhibitor ions reduce carbonate mineral growth rates and usually do not modify the reaction driving force (that is, the solution supersaturation). It is recognized that carbonate-crystal-growth inhibitors regulate mineral formation in Great Basin lakes such as Mono Lake and Pyramid Lake. Inhibitor-ion concentration levels also may have modified carbonate mineral formation in paleoenvironments.

Calcite crystal growth rates are reduced in the presence of millimole to micromole concentrations of magnesium and phosphate ions. In Pyramid Lake, concentration of these ions were sufficient to decrease calcite formation rates in the supersaturated lake water (Reddy, 1975). Magnesium-ion and phosphate-ion inhibition of calcite crystal growth was interpreted in terms of a Langmuir adsorption isotherm model. This model assumes that inhibitor ions adsorb at crystal growth sites on the calcite surface (Reddy, 1977, 1978, 1986).

A test of the proposed calcium carbonate formation mechanism in Pyramid Lake, using an in-lake crystal growth experiment, was begun in the summer of 1992. This test measured calcite and aragonite crystal growth rates for single crystals suspended in Pyramid Lake. Initial results from August and September 1992 (no growth in the supersaturated lake water) indicated the importance of crystal growth inhibitors, such as magnesium and phosphate ions, in calcium carbonate formation in Pyramid Lake.

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Climate of the Great Basin

Kelly Redmond, Western Regional Climate Center, Desert Research Institute, Reno, Nevada

The climate of the Great Basin is characterized primarily by aridity. The Great Basin is dry in two senses: precipitation is low, and the absolute and relative humidities of the air are low. The Great Basin is physically separated from potential oceanic moisture source regions by distance and mountains. In addition to mountains around the periphery, about 150 mountain ranges, separated by basins, occupy the interior of the basin. Because heavier and more frequent precipitation occurs in the vicinity of the mountains, most of the precipitation received in the basin falls over a relatively small fraction of the total horizontal area. Annual precipitation ranges from approximately 140 cm on mountain tops in the western part of the basin to 3 to 4 cm in Death Valley and vicinity.

The seasonal precipitation cycle indicates systematic variations with latitude/longitude position and with elevation. Houghton (1969) has identified the chief seasons that contribute most to the annual total precipitation at different points in the region. In the northern end of the basin, Hanson (1982) has used the dense network at Reynolds Creek near Boise, Idaho, to show that winter accounts for a successively greater fraction of annual precipitation as elevation increases. At that location, December–February precipitation accounts for about 43 percent of the annual total at 1,200 m and about 58 percent at 2,100 m.

The frequency of precipitation days (those days with at least 0.025 cm of precipitation) ranges from about 90 days per year at Boise and Salt Lake City to about 11 days at Lathrop Wells, Nevada. The drier the location, the greater the fraction of the annual precipitation that falls during the wettest few days. Despite the general aridity, water is a prime element in landscape alteration. Flash floods, although infrequent, leave their imprint for many decades. In many ways, climatic events lasting a few hours to a few days can be as important for long-term change as anomalies that persist for a month or longer.

The main influences on the annual cycle of precipitation amount and frequency are the seasonal expansion and contraction of the westerlies, the formation of low pressure systems in the Great Basin during the transition months, and the southwest monsoon in the summer. Specific locations have from one to four precipitation peaks within the year, each ascribable to a different cause. Each cause produces a certain typical component, so that the seasonal cycle at most locations can be described as a superposition of these main components, each with different weights. Houghton (1969) also used this notion in trying to account for the variety of annual patterns observed in the region.

The lack of precipitation measured at the surface is a reflection not only of a decreased frequency of water-bearing clouds, but also of the low atmospheric humidity. Convective clouds often have very high bases—also an effect of the low humidity—so that precipitation falls through thousands of meters of dry air, allowing ample opportunity for evaporation before reaching the valley floor. Precipitation falling onto mountains has a much smaller distance in which to evaporate, so a greater fraction of the precipitation leaving a cloud reaches the surface in mountainous locations.

In ecological and climatological terms, the numerous mountain ranges can be thought of as "islands" rising from the valley floors. Long-term climate measurements at elevated sites are extremely difficult to obtain, so that existing data sets tend to be dominated by data from inhabited locations, which severely impairs the ability to map climate elements adequately. The combination of modeling, remote sensing, and automated observations is increasingly being applied to the difficult problem of interpolation, which needs to be solved before mapping can proceed.

In the broadly elevated parts of the region, the dry soil and atmosphere result in large diurnal temperature ranges. Freezing temperatures are possible throughout the year in higher basins, and much of the region has a large number of freeze-thaw cycles. Somewhat protected by the Rocky Mountains and southerly enough to receive substantial solar energy in winter, the region seldom has extremely cold temperatures (-15°C or less) and almost never has low temperatures accompanied by wind.

Shallow temperature inversions are very common, and the Great Basin region would be very susceptible to air pollution if the population were larger. Persistent regimes of high pressure can result in extended periods of stagnant cold air in the cool season. The resulting decreases in visibility are easily noticeable because the air is among the cleanest in the United States. Visibilities of 150 km or more are frequent in eastern Nevada.

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Climatic Control of Late Quaternary Sedimentation on the Leidy Creek Fan in Fish Lake Valley, Nevada-California—Implications for Geomorphic Processes in Semiarid Climates

Marith C. Reheis and Constance K. Throckmorton, U.S. Geological Survey

A critical question in arid-land geomorphology has been debated for decades: Does sedimentation occur on alluvial fans when the climate is cool (effectively wet), in transition from cool to warm, or warm (effectively dry)? This question has never been adequately answered due to a general lack of datable material in alluvial-fan deposits. However, Slate (1992, this volume) has recently suggested that pulses of Holocene deposition on proximal fans in Fish Lake Valley commonly correspond to periods of warm climate or to transitions from cool to warm climate.

Extensive mapping and dating and some coring of proximal- and distal-fan deposits and adjacent marsh and peat deposits on the Leidy Creek fan in Fish Lake Valley in west-central Nevada and adjacent California provide unequivocal evidence that fan deposition in the Holocene was extensive and indicates that soils were forming on stable fan surfaces that had no debris-flow deposition during the late middle Wisconsin in this valley. Cores from the peat deposits, present in a narrow band from the location of core 1 to just north of the location of core 7 along the east edge of the marsh deposits (fig. 1), is yielding a detailed record of late Wisconsin and Holocene climate, based on analyses of pollen, diatoms, ostracods, and stable isotopes. This research has significant implications for geomorphic processes and regional climate patterns.

Proximal-fan deposits are well exposed in fanhead trenches and incised fault scarps along the White Mountains on the west side of Fish Lake Valley in the Leidy Creek drainage (fig. 1) and throughout Fish Lake Valley. The deposits consist mostly of superposed, matrix-supported, bouldery debris flows (Slate, 1992). Based on surface and soil characteristics, the deposits have been subdivided into units of late Pleistocene and early, middle, and late Holocene age (older deposits also are abundant elsewhere in the valley). The Holocene units are the most extensive and merge downfan with fine-grained distal-fan alluvium and valley fill. These units in figure 1 have been simplified from those in Reheis and others (1993, in press). The ages of the mapping units actually span only parts of the time intervals by which they are identified in figure 1 (for example, middle Holocene). The combined units show areas where deposits of distinctly different ages were combined for convenience, and the upper Holocene unit (fig. 1) includes two mapped units of different ages. The upper Pleistocene unit is only well exposed along the range front and is mostly buried by Holocene deposits to the east, but can locally be traced several kilometers downfan (fig. 1).

The exposed Holocene units have been dated in many parts of the valley by using ^{14}C analysis of charcoal and buried logs and by tephrochronology. These ages clustered into groups that indicated intervals of debris-flow deposition throughout the valley separated by intervals of nondeposition (Slate, 1992, this volume; Reheis and others, 1993, in press). Despite the abundance of datable material (more than 20 ^{14}C ages and more than 20 correlated tephra layers), no ages have been obtained from the proximal-fan deposits that are older than 6,850 yr B.P. (the Mazama ash). However, a few late Pleistocene ages (35,000–10,000 yr B.P.) have been obtained from alluvium that probably accumulated behind a landslide within the White Mountains.

The age of the upper Pleistocene unit is not so well determined, but based on a thermoluminescence age (Slate, 1992), a beryllium-10 age estimate (M.J. Pavich, U.S. Geological Survey, written commun., 1993), and identification of two tephra layers (A.M. Sarna-Wojcicki, U.S. Geological Survey, written commun., 1992), it is no younger than about 35,000 yr and probably no older than 100,000 yr.

Seven cores from 9 to 13 m long were obtained in 1991 from the distal part of the Leidy Creek fan, including 2 cores (cores 1 and 7) from a peat bog at the toe of the fan and another core (core 2) from adjacent marsh deposits (fig. 1). These cores were intended to provide a climatic and a stratigraphic record to investigate the relation of fan sedimentation to climate and to correlate to the sedimentation record provided

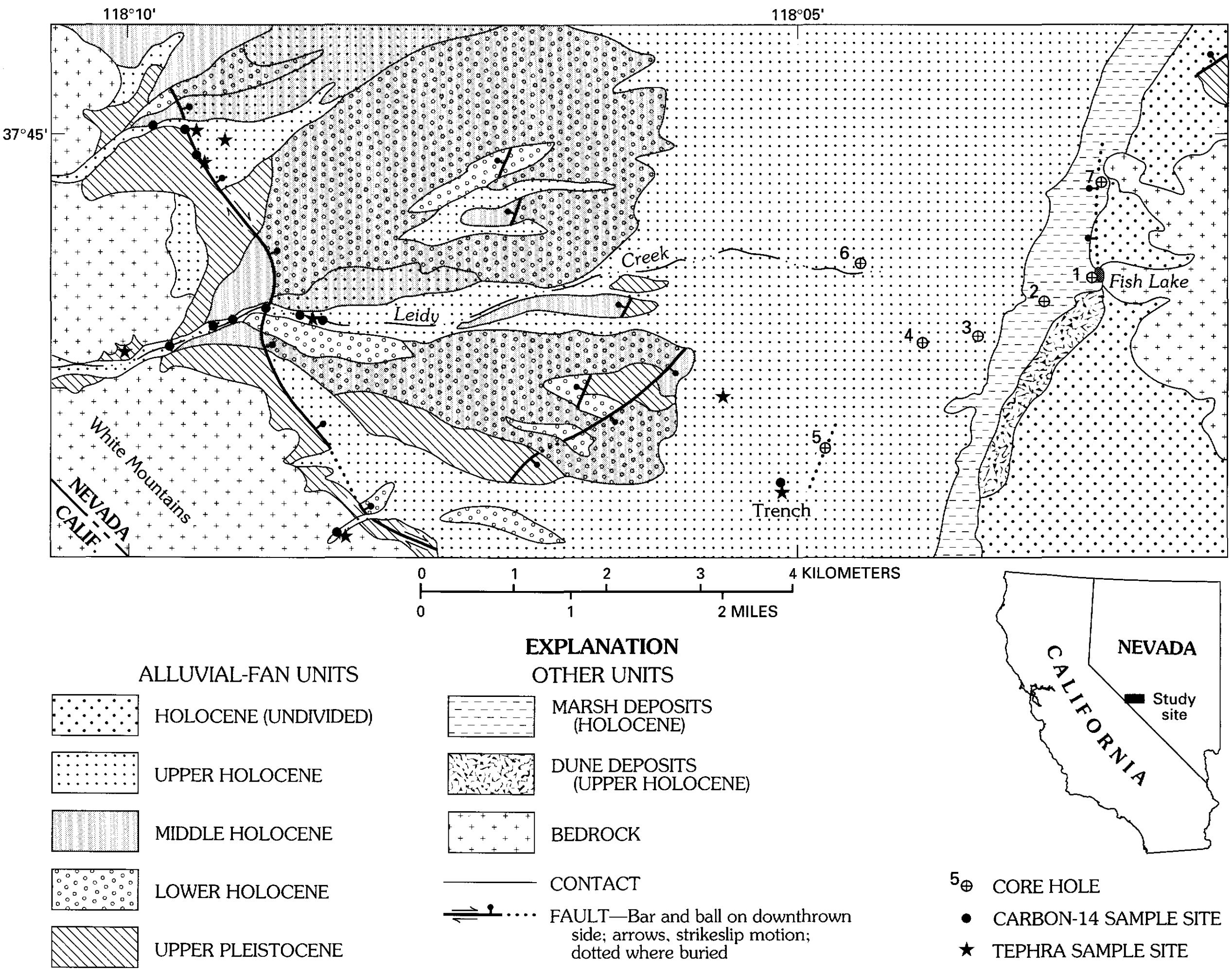

Figure 1. Generalized geologic map of Leidy Creek fan and vicinity in Fish Lake Valley showing locations of core holes and exposed age-control sites.

by the exposed proximal-fan deposits. Cores 1 and 7 are similar; both have about 5.5 m of peat and organic-rich silt and clay overlying about 2.0 m of fluvial, well-sorted, clean pebbly sand that, in turn, overlies fine-grained deposits. In core 1, these basal deposits are poorly sorted, pebbly sand and silt containing interbedded silty clay layers, whereas in core 7, the basal deposits are mostly stiff, green, silty clay. Core 2 consists mostly of locally organic-rich, calcareous silt and clay.

Cores 3 through 6, on the distal fan, indicate coarsening trends upward in each core and in an upfan direction. Each core contains three weak buried soils in the upper few meters; these buried soils are closely spaced in core 3, but are separated by sediment layers that thicken upfan. A backhoe trench, excavated 0.5 km southwest of core 5 to a depth of 5 m (fig. 1), also exposed three weak buried soils. In addition, cores 2 through 6 contain a strong buried soil characterized by an organic A horizon about 10 to 30 cm thick, a Bt or Bw horizon as much as 20 cm thick, and a Bk horizon at least 30 cm thick and containing 15 to 50 percent $CaCO_3$. The depth to the top of this buried soil gradually increases upfan from about 4.6 m in core 2 to about 11.5 m in core 6.

Eighteen ^{14}C and tephra ages on material from cores 1 and 7 indicate that organic-matter accumulation has been continuous since about 12,500 yr B.P. A single ^{14}C age of 24,240 ± 330 yr B.P. was obtained from the fine-grained sediment about 1 m below the base of the fluvial sand in core 1. Four ^{14}C ages from cores 2, 3, and 4 constrain the age of the strong buried soil. Three samples from the A horizon (mean-residence time ages) were dated at 9,630 ± 80, 8,900 ± 80, and 9,900 ± 80 yr B.P., respectively; the fourth sample was from organic-rich clay about 50 cm above the buried soil in core 2 and was dated at 8,810 ± 80 yr B.P.

Preliminary interpretation of the ^{14}C ages, tephra correlations, and stratigraphic relations yielded the following conclusions and speculations:

1. Organics accumulated relatively rapidly in cores 1 and 7 between about 11,000 to 10,000 and 8,000 yr B.P. and between about 4,000 and 2,000 yr B.P. and accumulated slowly between about 7,000 and 4,000 yr B.P.

2. Beginning at about 8,900 yr B.P., the fan was aggrading by debris-flow and distal-fan sedimentation; the thickness of these sediments decreases downfan from about 11.5 to 5.5 m. Thus, the Holocene has been a time of rapid fan sedimentation. However, the weak buried soils in the upper parts of the distal-fan cores indicate that deposition rates were slower in the middle(?) to late Holocene than in the early Holocene.

3. Prior to about 12,000 yr B.P., the climate was effectively wet enough that an axial stream (represented by the fluvial deposits beneath the peat in cores 1 and 7) flowed at the toe of the fan. Deposits of this age on the fan, however, are represented by the strong buried soil. Thus, debris-flow deposition was limited or absent, and slopes were stable, presumably because the climate was moist enough to increase the vegetative cover.

4. At and before 24,000 yr B.P., a small, shallow lake may have existed at the toe of the fan, based on the fine-grained deposits beneath the fluvial unit. If this lake existed, it was probably small because there are no preserved shorelines and because similar sediments do not occur at the same depth in core 3, located only about 1 km upfan from core 1.

The development of the strong buried soil present in most of the cores may have required several tens of thousands of years. If the interpretations are correct, then the Leidy Creek fan (and, by extension, other fans in Fish Lake Valley) was stable and had no deposition during the middle (?) and late Wisconsin, when the climate was relatively wet.

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Spatial and Temporal Variation in Nitrogen Fixation in Pyramid Lake, Nevada

Cathryn L. Rhodes, John E. Reuter, Martin E. Lebo, and Charles R. Goldman,
Division of Environmental Studies, University of California, Davis

Introduction

Although generally considered oligotrophic (Galat and others, 1981), Pyramid Lake in Nevada has large summer-fall blooms of the nitrogen-fixing cyanobacterium *Nodularia spumigena.* Near the bloom peak, surface chlorophyll *a* concentration increases from an average epilimnetic value of 3.3 mg/L to as much as 100 mg/L at the surface.

The size of the *Nodularia* bloom varies dramatically from year to year. Although *Nodularia* is always a major nitrogen source in Pyramid Lake (Horne and Galat, 1985), the mass of nitrogen fixed annually by the bloom varies by two orders of magnitude. During the very large 1992 bloom, approximately 1,000 metric tons of nitrogen were fixed, accounting for 94 percent of the total annual nitrogen budget. Even during the small 1991 bloom, approximately 90 metric tons of nitrogen were fixed, accounting for 60 percent of the annual nitrogen budget. Horne and Galat (1985) reported that 866 metric tons of nitrogen were fixed during the 1979 bloom, accounting for 81 percent of the annual nitrogen budget.

Spatial and temporal heterogeneity of the blooms complicates assessment of nitrogen fixation by *Nodularia.* The dense surface scums that characterize *Nodularia* blooms in Pyramid Lake may occur anytime between July and October (Galat and others, 1990). Horizontal patchiness of the blooms is reflected in the surface chlorophyll concentrations. On any date during a bloom, surface chlorophyll may vary spatially by two to five orders of magnitude. Because of the spatial and temporal heterogeneity characteristic of these blooms, a multiple synoptic-sampling approach provided the best assessment of nitrogen fixation. In addition to providing a good estimate of the nitrogen budget, the multiple synoptic approach allowed the examination of the spatial and temporal patterns of nitrogen fixation in Pyramid Lake.

Study Site

Pyramid Lake is located in the Great Basin Desert approximately 100 km north of Reno, Nevada. The lake is the terminus of the Lake Tahoe-Truckee River watershed. Pyramid Lake is deep, having a maximum depth of 105 m and an average depth of 59 m (Galat and others, 1981). The lake is nitrogen poor and phosphorus rich. Generally, epilimnetic dissolved-inorganic nitrogen averages less than 10 mg/L during the summer, whereas dissolved-inorganic phosphorous concentrations average greater than 60 mg/L. Pyramid Lake is hyposaline (3,500 mg/L) and alkaline (9.4 pH).

Materials and Methods

Surface synoptic sampling was conducted approximately weekly during the 1991 and 1992 *Nodularia* blooms. Acetylene reduction was used to assess nitrogenase activity and was assumed to be proportional to nitrogen fixation. Acetylene-reduced samples were incubated the day after the synoptic sampling for 2 hours across solar noon. Incubations were conducted in onshore flow-through incubators under ambient light conditions. At the end of the incubations, acetylene-reduction bottles were shaken, and the headspace gases were injected into Vacutainers. The gas samples were analyzed using a flame-ionization-detector-equipped gas chromatograph fitted with a Porpak N column. Chlorophyll samples were stored frozen until extraction using alkaline acetone and fluorometric determination.

Results

Surface acetylene-reduction rates and chlorophyll concentrations were spatially and temporally heterogeneous during 1991 and 1992. Acetylene reduction was first detected during the 2-yr study on August 28, 1991, and on July 27, 1992. The onset of the 1991 *Nodularia* bloom was associated with the decrease in wind speed in early August, and the bloom peaked around September 5, 1991. *Nodularia* appeared a month earlier in 1992 than in 1991. Although the beginning of the 1992 bloom was not associated with a drastic wind-speed change, the average wind speed during summer 1992 was similar to the average wind speed during the 1991 bloom. Acetylene-reduction rates and chlorophyll concentrations during the 1991 bloom were 2 to 2,000 (nmol/L)/hr and 0.5 to 13 μg/L, whereas during the 1992 bloom, these parameters ranged from 0.6 to 30,500 (nmol/L)/hr and 1 to 433 μg/L. The acetylene-reduction rates and the chlorophyll concentrations indicated the dramatic size difference between the 1991 and the 1992 *Nodularia* blooms. Collapse of the 1991 and the 1992 blooms coincided with an increase in wind speed.

In addition to the wide variation in acetylene-reduction rates and in chlorophyll concentrations, the chlorophyll-specific acetylene-reduction rate varied by five orders of magnitude. Although the variation was not associated with sampling date or with station, the variability in chlorophyll-specific acetylene reduction during 1992 was greater than in 1991. Most of the variability occurred at low chlorophyll values (<5 μg/L).

The spatial patterns in acetylene-reduction rates and in chlorophyll concentrations were analyzed using multidimensional scaling. Before analysis, all values were standardized within sampling date by subtracting the mean and dividing by the standard deviation. The station interrelations mapped in a two-dimensional ordination using multidimensional scaling indicated a north-south spatial pattern in acetylene-reduction rates. In contrast, no similar pattern was detected in the chlorophyll concentrations. Based on Landsat images of *Nodularia* blooms (Galat and others, 1990), a pattern indicting the concentration of algal material along the west-central shore of Pyramid Lake was expected. Neither the standardized acetylene-reduction rates nor the standardized chlorophyll had a pattern consistent with this nearshore distribution.

Conclusions

The timing and magnitude of the annual *Nodularia spumigena* bloom in Pyramid Lake is related to wind speed. The large, early-1992 bloom occurred during a low wind-speed year, whereas the small, late-1991 bloom occurred during a calm period in an otherwise windy year. The inverse relation of wind speed to bloom size implies that water-column stability affects the size of the yearly bloom. There is no consistent relation of the rate of nitrogen fixation to chlorophyll concentration. Because the chlorophyll-specific nitrogen-fixation rate is variable, estimates of nitrogen fixed by *Nodularia* based on chlorophyll concentration may be inaccurate. The station interrelations mapped in a two-dimensional ordination using multidimensional scaling indicated a north-south spatial pattern in acetylene-reduction rates. This pattern is consistent with wind-direction data from Sutcliffe, Nevada, that indicated prevailing north winds during the summers of 1991 and 1992. A similar analysis based on chlorophyll concentrations did not indicate a comparable spatial pattern.

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Middle Holocene Decrease in the Surface Level of Lake Tahoe, Nevada and California

Martin Rose, Biological Sciences Center, Desert Research Institute, Reno, Nevada
Susan Lindstrom, Archaeologist, Truckee, California

Submerged tree stumps in Lake Tahoe, dating 6,300 to 5,000 cal (calibrated) yr B.P., document an interval of lower lake level. Three hypotheses can be offered to account for this phenomenon. The first and probably most obvious hypothesis is that a period of aridity caused the lake level to drop. The forest surrounding the lake gradually followed the lake surface to a lower elevation. Approximately 5,000 yr B.P., a return to more mesic conditions caused the lake to rise and kill the trees, some of which are still preserved in the cold water. A second hypothesis is that sediments were deposited over the natural sill during a period of decreased outflow that is related to a regional drying trend. The sediment accumulation over the sill functioned as a dam, causing the lake level to rise, resulting in the death of trees that became submerged. A third hypothesis is that a structural rise in the sill or subsidence at the south end of the basin, or both, resulted in the death of trees that were submerged (Harding, 1965).

Lake Tahoe, averaging a depth of 302 m, is believed to be the 10th deepest lake in the world and the 3d deepest in the United States. Its maximum depth is about 503 m. The total drainage area that is tributary to the lake is 1,310 km^2; the lake itself occupies 497 km^2. With an average surface elevation of 1,897 m, it is the highest lake of its size in the United States. Lake Tahoe occupies a graben, a steep-sided valley formed when faulting caused a block-shaped area to drop relative to the surrounding terrain.

Lake level has fallen below the natural sill 11 times during the present century (Lindstrom, 1990). It is currently (1993) below the natural sill for the fourth consecutive year. During the drought of the late 1920's and early 1930's, the lake dropped below the natural sill for 6 straight years, reaching a recorded low of 1,896.4 m in December 1934 (Tahoe World, Incline Village, Nev., April 21, 1988).

When Lake Tahoe was at its lowest elevation in 1934, Harding (1965) observed 11 stumps near Tallac. Three samples from two stumps were radiocarbon dated. One stump, at an elevation of 1,896.7 m dated 4,250 ± 200 and 4,790 ± 200 yr B.P. The second stump, at an elevation of 1,896.7 m, dated 4,460 ± 200 yr B.P. These dates are about 700 yr less than the calendar age of the stumps because they have not been calibrated against the exactly dated bristlecone pine record. Six of the 11 stumps located by Harding (1965) were relocated by Lindstrom in 1985 (Lindstrom, 1990). Elevations for the six stumps ranged from 1,896.1 to 1,896.9 m. Lindstrom (1990) reported that two radiocarbon samples obtained from Harding's lowest reported stump dated 4,870 ± 60 and 4,520 ± 60 yr B.P. (uncalibrated dates) or 5,640 and 5,197 cal yr B.P. (calibrated). These dates indicated that the tree was at least 350 yr old. In 1990, some of these stumps were relocated, and several increment cores were obtained that showed the radial growth pattern. If the lake-level fluctuations were climatic in origin, the age of the tree indicated the lake dropped below 1,896.9 m for at least 350 yr.

Between 1989 and August 1992, 20 additional stumps were located at South Lake Tahoe between Stateline and Emerald Bay. All of these stumps are at or below the natural sill level. Twelve stumps are at Baldwin Beach, one is at the mouth of Emerald Bay, four are at Al Tahoe, and one is at El Dorado Beach in Stateline. Two stumps were recently found near the University of Nevada Extension facility and 4-H camp in Stateline.

The largest stump found to date (1993) is more than 2.5 m tall and about 1.07 m in diameter. In general, the deepest stumps are the tallest, and the shallowest stumps are the shortest. The shallow stumps have been repeatedly exposed during the past 5,000 to 6,000 yr and have been subjected to decay and abrasive wave action. Fourteen stumps have been radiocarbon dated; the ages were between 6,304 and 4,846 cal yr B.P. The oldest stumps are generally the deepest. They document a period about 1,300 yr long when the trees were exposed, but whether the land on which the trees are located was continuously exposed for the entire period cannot be answered with the limited data available. Higher frequency fluctuations in lake level may have occurred, but such fluctuations cannot be identified using the small sample of radiocarbon dates.

Until 1993, no stumps from Lake Tahoe were found outside the south-shore region. Because of the immense publicity the trees have received in newspaper articles and on television, additional stumps have been reported and verified elsewhere in Lake Tahoe. A large submerged tree in about 10 m of water has been independently reported by two divers several hundred meters off the pier at Tahoe Vista. A single tree has been reported off Dollar Point, and numerous additional stumps have been independently reported by several individuals from Emerald Bay.

The unique and potentially scientifically valuable reports are of submerged trees at depths of 21 to 23 m in Sand Harbor. One of the trees was recently located using a remotely operated submersible. The tree is about 10 m long and larger than 1.3 m in diameter. It is lying horizontal, located along a dropoff to deeper water; the trunk faces the lake, and the top points toward shore. Currently, trees that are rooted upright are being sought, a difficult and time-consuming task, at depths of 23 m. The Sand Harbor trees are definitely not sinkers or remnants of logging activity during the last century because they have not been cut.

Submerged tree stumps have been observed in other high-altitude Sierran lakes. Ten stumps were observed in Independence Lake, California, several years ago. Three samples from below the pre-dam level were radiocarbon dated from between 700 cal yr B.P. to within the past 150 yr. Samples for radiocarbon dating also have been obtained from several submerged tree stumps at pre-dam levels in Donner Lake, California (Lindstrom, 1990). Additional stumps at depths of about 21 to 24 m have been reported from Fallen Leaf Lake, California (John Kleppe, University of Nevada, oral commun., 1990). One of the deep stumps was verified by a diver in 1990. Finally, Lindstrom (1990) reported stumps 3 to 6 m below the surface of a lake west of Downieville, California. The widespread presence of submerged stumps of approximately similar ages would strongly indicate that a large-scale climate change caused simultaneous drops in the surfaces of several lakes.

Several hypotheses can be offered to account for the presence of the submerged trees below the natural sill of Lake Tahoe. A climatic argument can be proposed that postulates more arid conditions than in the present century. According to this hypothesis, a middle Holocene decline in lake level resulted from a drying trend in the Western United States. Such a trend is supported by geomorphological and palynological data. This evidence includes 5- to 6-m-high beach ridges at Kings Beach, drowned shorelines and sand-spit-dammed marshes along the south shore, submarine canyons on the east and west sides of the lake, buried lakeshore deposits on the south shore, pollen cores from within the Tahoe Basin, and sediment cores from the lake bottom. A modification of the hotter and drier hypothesis suggested by Davis and Elston (1972), was that the middle Holocene may have been more variable and not necessarily hotter and drier. A combination of the two hypotheses also is possible.

A subsidiary to the climatic hypothesis proposes that a rise in the level of Lake Tahoe could have resulted from sediment accumulation over the natural sill. Cores obtained from inside the outflow channel by the U.S. Department of Agricultural Forest Service (Hug, 1989) and the Bureau of Reclamation (Hawkins and others, 1986) revealed 4-m-thick sediments. Radiocarbon samples need to be obtained to date the sediments and provide a test for this subsidiary.

Another hypothesis that could account for the presence of the submerged stumps in Lake Tahoe is tectonic in origin (Harding, 1965). This hypothesis is that there was a structural rise in the elevation of the natural sill near Tahoe City or subsidence at the south end of the basin, or both. Although there has been considerable tectonic activity in the Tahoe Basin, it has not been demonstrated to be of Holocene age. Tectonic activity within the Tahoe Basin, and in the lake in particular, during the Holocene needs to be investigated.

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Sediment-Magnetic and Paleomagnetic Records of Middle Pleistocene Sediment from Buck Lake, Southern Oregon

Joseph G. Rosenbaum, Richard L. Reynolds, Priscilla Fitzmaurice, David P. Adam, and Andrei M. Sarna-Wojcicki, U.S. Geological Survey

Sediment-magnetic and paleomagnetic studies of lacustrine sediment from Buck Lake in southern Oregon are designed to contribute to understanding regional paleoclimatic change during the Pleistocene. The results could be used to determine intrabasin and interbasin correlations and sedimentation rates. These studies also may reveal variations in the concentration and character of detrital magnetic particles that record changes in environmental conditions in the source area, in depositional environments, or in alteration of the sediment. Magnetic-susceptibility (MS) measurements, using a whole-core, pass-through instrument, have been made for the greater than 41-m length of the core. Sediment-magnetic and paleomagnetic analyses also have been completed for more than 300 azimuthally unoriented specimens from depths of 5.2 to 24.1 m in a rotary-drill core.

The age of the sediment is controlled by several tephra layers correlative with the Loleta ash bed/Bend pumice (greater than 400,000 to 300,000 yr B.P., depth 9.5 to 9.6 m), the Rockland ash bed (less than 470,000 to 400,000 yr B.P., depth 19.9 m), the Dibekulewe ash bed [between 600,000 and 500,000 yr B.P. (?), depth 20.8 m], the Lava Creek B ash bed (greater than 665,000 yr B.P., depth 22.3 m), and the Rye Patch Dam ash bed [670,000 yr B.P. (?), depth 22.6 m]. Normal polarity inclinations were measured to 24.1 m, indicating deposition of the sediment during the Brunhes chron (less than 780,000 yr B.P.).

Comparisons of magnetic properties that reflect the abundance, mineral type, and grain size of magnetic minerals (fig. 1) strongly imply that the magnetization of the sediment is dominated by detrital magnetite. The magnetic properties indicate no obvious post-depositional alteration, such as the conversion of magnetite to ferric oxide or to nonmagnetic iron sulfide minerals, and the properties lack evidence for authigenesis of magnetic iron sulfide, such as greigite. Nevertheless, the requisite petrologic study to confirm or deny a detrital origin of magnetic grains needs to be undertaken.

The MS (primarily magnetite content, fig. 1) of fine-grained lacustrine sediment, the predominant lithology to 24 m in depth, ranges over about 2 orders of magnitude. Between 5- and 24-m depths, two zones, each 3 to 4 m thick, of relatively high MS are bounded by intervals of lower MS. Except for the Rockland ash bed, a 50-cm thick interval of basaltic cinders above the Rockland ash (greater than 19.4 to 19.7 m), and two thin tephra layers (12.23 and 14.92 m), each of which contains material having high MS, the variations in MS have no obvious correspondence to lithology.

The content of ferric oxide mineral, such as hematite [the hard isothermal remanent magnetization (HIRM) shown in fig. 1], correlates so closely with the content of magnetite that both mineral types must be part of a detrital heavy-mineral suite and are not related by in-situ alteration. The ratio of magnetite to ferric oxide mineral (indicated by the S parameter in fig. 1), however, does not remain constant. The greater the absolute amount of magnetite (high MS), the greater the relative amount of magnetite (high S values). This relation indicates that high magnetite content is associated with source rocks that have undergone little chemical weathering involving alteration of magnetite to ferric oxide. Conversely, low magnetite content may be associated with relatively more chemical weathering (low S values), possibly during soil formation, in the catchment. On a broad scale, large magnetic grain size (lower ratios of IRM/MS and ARM/MS indicate larger size, fig. 1) corresponds to high magnetite content (high MS), and small magnetic grain size (high-ratio values) corresponds generally to low magnetite content. Paradoxically, the magnetite grain size generally decreases (higher ARM/MS) as magnetite content increases (higher MS) for the MS spikes within both high MS zones,. Whatever erosional or depositional processes increased the magnetite content, causing the high MS spikes within the broad high MS zones, they also decreased the grain size of the magnetite. The possibility of mechanical weathering needs to be tested using detailed grain-size

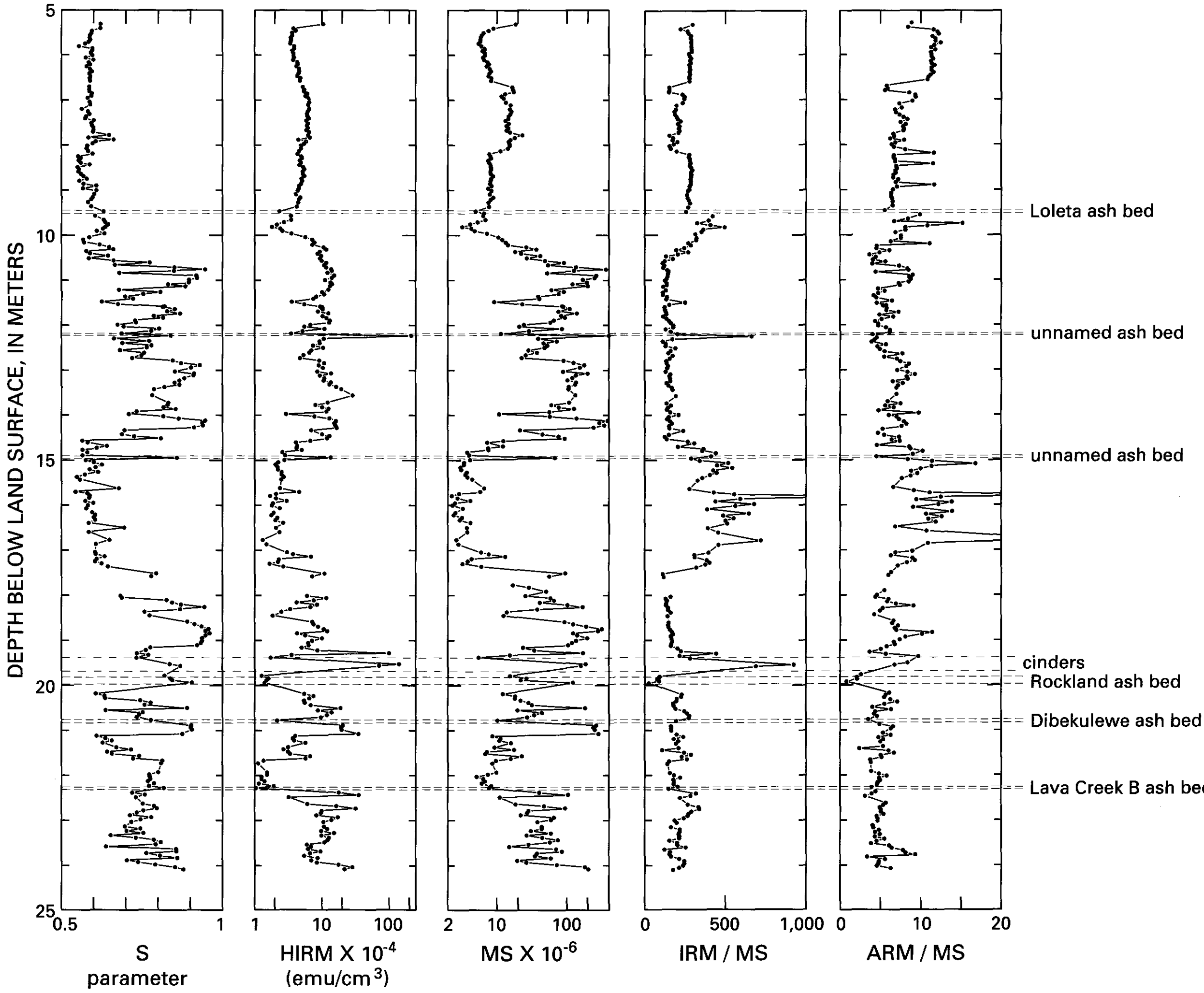

Figure 1. Plots of magnetic properties with depth, Buck Lake (southern Oregon) core 1. The S parameter shows relative amounts of magnetite and high-coercivity magnetic minerals (for example, ferric oxide minerals, such as hematite and goethite). Higher values indicate more magnetite. The S parameter is the ratio of the isothermal remanent magnetization (IRM) acquired in a backfield of 0.3 Tesla (T) to forward IRM acquired at 1.2 T: $-IRM_{-0.3T}/IRM_{1.2T}$. HIRM (hard IRM; $[IRM_{-0.3T} + IRM_{1.2T}]/2$) is a measure of the concentration of high coercivity magnetic minerals. Magnetic susceptibility (MS), in cgs-volume units, is primarily a measure of magnetite content. IRM/MS and ARM/MS (ARM=anhysteretic remanent magnetization) are measures of magnetic grain size. Lower values of these ratios indicate larger effective magnetic grain size.

analyses of bulk sediment across a number of the magnetic features.

Pollen and MS zonations correspond closely. Warm climates are indicated for the intervals 5.3 to 9.8 m and 14.5 to 16.5 m, which are both zones of mostly low MS. In contrast, cold and open environments of sagebrush steppe are characteristic of the intervals 9.8 to 14.5 m and 16.5 to about 20 m, which are both zones of predominantly high MS. The magnetic mineralogy thus seems to indicate changes in climate, at least for broad intervals of time. The shorter wavelength magnetic features also may record climate change—a possibility that could be tested by detailed palynologic and sedimentologic analyses of a few such features.

Tephrochronology in Studies of Middle to Late Quaternary Climate Change, Northern Great Basin

Andrei M. Sarna-Wojcicki, Charles E. Meyer, and Elmira Wan
U.S. Geological Survey

Tephrochronology, the study of the age and correlation of volcanic ash layers, is useful in studies of Quaternary climate change in the northern Great Basin because it makes it possible to identify spatial and temporal reference datum planes (Sarna-Wojcicki and Davis, 1991; Sarna-Wojcicki and others, 1991). Much of this region is downwind from two major volcanic areas that have been the sources of sporadic, but prolific, explosive volcanic activity throughout the middle and late Quaternary. These sources are the Cascade Range of Washington, Oregon, and northern California, and the Long Valley-Mono Craters area of east-central California (fig. 1). Isotopic and other numerical ages determined from widespread ash layers from these sources provide local and regional age control, whereas chemical and petrographic characterization of the layers allows unambiguous temporal correlation of associated sediments. Correlations of ash layers also provide a means of evaluating the accuracy of numerical ages obtained from the same layers. Tephrochronology has been particularly useful when applied to upper Quaternary sediments because the geologically young tephra is usually unaltered and because the method complements magnetostratigraphy, which may not be as effective within the Brunhes Normal Chron (the last approximately 775,000 yr).

Tephra layers that originated from the Cascade Range generally provide the largest number of unambiguously identifiable ash layers in the northern Great Basin. This is because the volcanic field is large and has several active sources of explosive activity (fig. 1) and also because the tephra is derived from small magma bodies that formed frequently throughout Quaternary time (table 1), providing compositional contrasts that can be readily determined with the electron microprobe.

Tephra layers that originated from the Long Valley and Mono Craters in east-central California are generally less distinctive and, consequently, more difficult to identify than tephra layers from the Cascade Range, probably because the former tephra layers were derived from mostly larger, longer-lived magma bodies. Nevertheless, sets of these layers can be separated on the basis of chemical characteristics and chronostratigraphic data into temporal sequences representing several past eruptive episodes. Furthermore, certain individual layers of the Long Valley and Mono Craters suite can be readily identified. Tephra layers that originated from the volumetrically smaller Inyo Craters and from Mammoth Mountain in Long Valley are generally distinctive.

All tephra layers become thinner, finer, and scarcer with distance from the two main source areas. Dispersal from the sources was multidirectional. Some tephra clouds, such as the Loleta ash, were carried to the west, and their layers are present in marine sediments of coastal northern California and in deep-sea cores of the Pacific. Layers from these eruptions provide vital chronostratigraphic units that allow correlation between land basins and the deep-ocean paleoclimatic record. Predominant direction of tephra transport, however, was toward the east. The Lahontan Basin of northwestern Nevada (including Pyramid Lake, the Humboldt River Basin, and the Carson Sink) is the area where tephra layers from the two volcanic sources overlap most frequently. Widespread tephra layers derived from the Yellowstone volcanic source in northwestern Wyoming and in eastern Idaho also are present within the northern Great Basin.

Generally, good numerical age control to about 35,000 yr B.P. from radiocarbon dating is obtained from Holocene and late Pleistocene tephra layers, although finite ages within the older end of this range may be suspect. Tephra layers in the age range of about 300,000 to 35,000 yr B.P. have been poorly dated until recently. New work using the $^{40}Ar/^{39}Ar$ method, combined with thermoluminescence and uranium-series methods, has begun to provide geologically realistic and stratigraphically consistent numerical ages within this range. Laser-fusion $^{40}Ar/^{39}Ar$ analysis of single crystals is providing some reliable new age data on both proximal and distal tephra layers older

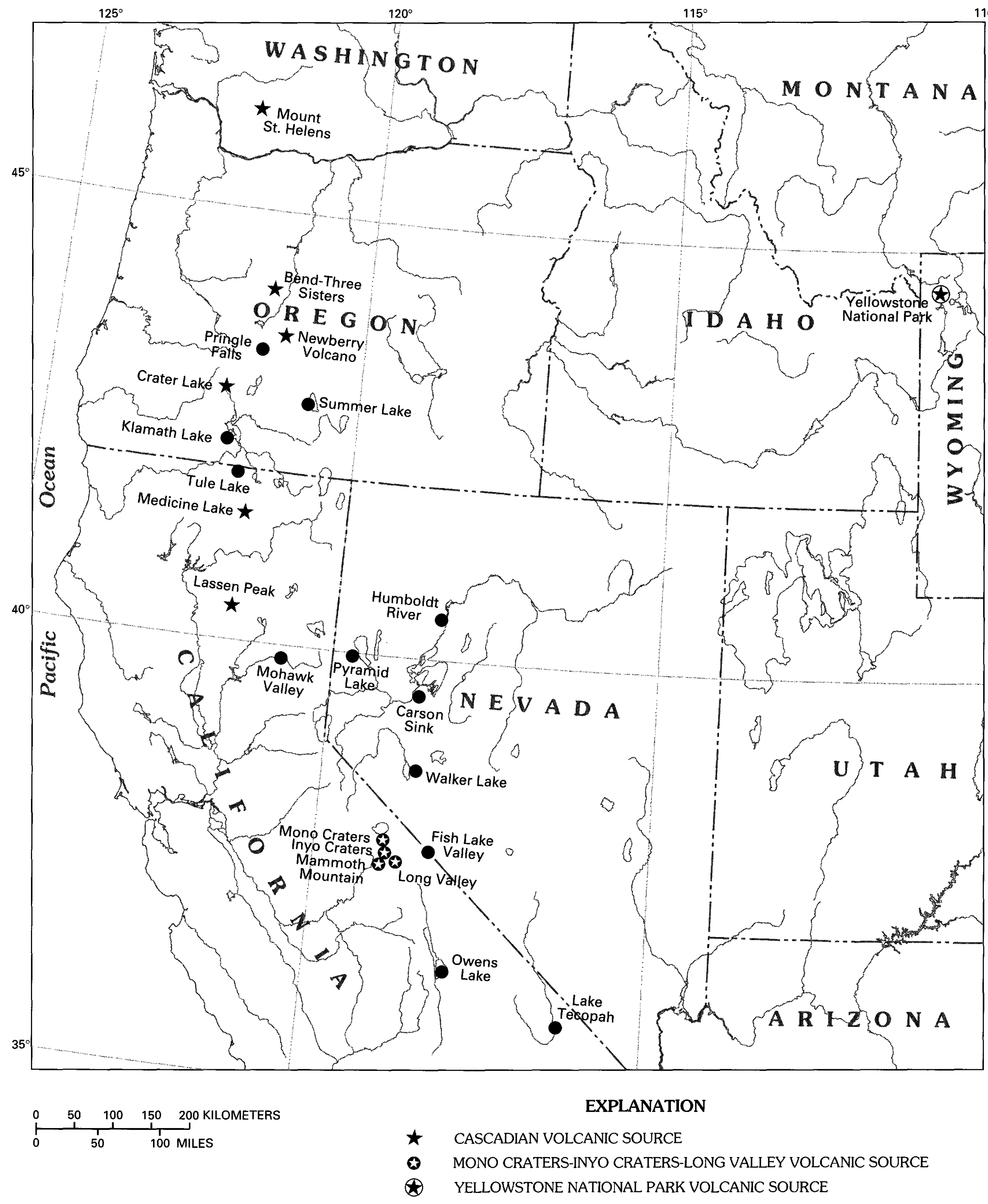

Figure 1. Location of middle to late Quaternary volcanic sources and sites where studies of Quaternary climate change are being conducted or have been recently completed.

Table 1. Tephra layers useful in correlation of upper Quaternary sediments, northern Great Basin

O-stage[1]	Age, (10^2 yr B.P.) (age range) [age event]	Name of layer (or tephra sequence) [name of magnetic event]	Layer, also known as (number of layers in sequence) [name of magnetic event]	Source of tephra
1	(~0.7-11)	(Mono-Inyo Craters)	(20 layers, minimum)	Mono, Inyo Craters
1	6.85 ± 0.15	Mazama[2]	Climatic airfall pumice	Crater Lake
1	7.02 ± 0.05	Tsoyowata	Llao Rock airfall pumice	Crater Lake
2	(~13-36)	(Wilson Creek ash beds)	(19 layers, minimum)	Mono Craters
2	20.5-23.4	Trego Hot Springs	Redcloud pumice	Crater Lake
2	~25	Wono		Crater Lake
2	~27	Tule Lake T2438	Summer Lake E1	unknown
2	~27	Wilson Creek ash 15		Mono Craters
2	[~27]	[Mono Lake Event]		
2-3	(33-36?)	(Marble Bluff)	(Mount St. Helens set C) (5 layers, minimum)	Mt. St. Helens
4?	55-75	Olema	Clear Lake Ash 4	Newberry (?)
3-5?	(50-100)	(Proto-Mono Craters and Mammoth Mountain)		Mono Craters and Mammoth Mountain
5	[112-114?]	[Blake Event]		
5	~120-130	Tule Lake T64[2]	DSDP 173-1-3, 4	Crater Lake (?)
6?	~140	Tule Lake T1193	Summer Lake V	Newberry (?)
6	~150	Summer Lake FF		Bend-Three Sisters (?)
6	~150	Summer Lake GG	Paoha Island 4	Bend-Three Sisters (?)
6	~155-160	Tule Lake T2080L	Wadsworth Bed (?)	Medicine Lake (?)
6	~160	Summer Lake JJ	Paoha Island 3	Bend-Three Sisters (?)
6	~160±25	Tule Lake T2023	Summer Lake KK	Medicine Lake
6?	[~160-180]	{Jamaica?]	[Biwa I Event?]	
6?	~180	Summer Lake NN	Paoha Island 2	Bend-Three Sisters (?)
10?	300-400	Loleta[2]	Bend pumice	Bend-Three Sisters (?)
11?	400-470	Rockland[2]	Rockland pumice tuff breccia; Maidu	Lassen Peak area
16?	~500-600?	Dibekulewe		unknown
16	665 ± 5	Lava Creek B[2]	Pearlette O	Yellowstone
16	670?	Rye Patch Dam	Desert Springs	Bend-Three Sisters
19	758 ± 2	Bishop B[2]	Ash of the Bishop Tuff	Long Valley Caldera
[19]	[~775]	[Brunhes/Matuyama Chron boundary]		

[1]Imbrie and others (1984).

[2]Tephra layers present, or magnetic events detected, in both marine and nonmarine sediments.

than about 100,000 yr when potassium-rich minerals, such as sanidine, are present. This mineral is present in tephra layers from the Long Valley-Mono Craters and Yellowstone sources, but is not present in tephra layers from the Cascade Range. For tephra layers from the latter sources, plagioclase concentrates of about 100 mg or more that are obtained from proximal exposures can be analyzed by the step-heating $^{40}Ar/^{39}Ar$ method. These methods minimize the problem of detrital or accidental contamination in the mineral concentrates and permit evaluation of results caused by alteration or other perturbations of the argon isotopic ratios that could result in spurious ages.

Recently completed or ongoing studies using tephrochronology, which are relevant to paleoclimatic studies of upper Quaternary sediments within the western Great Basin and closely adjoining regions, are within the Pringle Falls and Summer Lake area of central Oregon; the Tule Lake-Klamath area of south-central Oregon and northern California; the Mohawk Valley of northeastern California; Pyramid Lake, Humboldt River, Carson Sink, Walker Lake, and Fish Lake in western Nevada; and the Mono Basin (which contains the Mono Craters), Long Valley, Owens Valley, and Lake Tecopah in east-central and southern California (fig. 1).

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Effect of Holocene Climatic Fluctuations on Alluvial-Fan Deposition, Fish Lake Valley, Nevada-California

Janet L. Slate, Associated Western Universities–Northwest Division, Washington State University, Tri-Cities Campus

Many authors recognize similarity in deposits, morphology, and inferred processes in alluvial fans of the arid and semiarid Southwest United States (for example, in Nilsen, 1985), but interpretations of the origins and timing of deposition and entrenchment of modern fans are diverse. A combination of several factors may control fan formation and development; these factors include source-basin lithology, tectonics, and climatic change. Whether fan aggradation occurs during relatively humid climatic intervals (Lustig, 1965; Dorn and others, 1987) or during transitional intervals from relatively humid to arid periods (Bull and Schick, 1979; Wells and others, 1987) is controversial. Relating fan deposition to climatic change requires independent age control that is generally difficult to establish in desert basins because of the paucity of datable materials in such deposits (Wells and others, 1987).

Fish Lake Valley (FLV) in Nevada and California (fig. 1) is an ideal place to study the timing of alluvial-fan development. The Quaternary deposits contain abundant datable materials such as charcoal and wood fragments, tephra, and buried soil A (Ab) horizons (table 1). These materials help to constrain the ages of fan formation and can help to resolve the controversy of climatic controls on fan aggradation and entrenchment.

The fan deposits were dated wherever suitable materials were available to help determine whether deposits mapped in four different drainage settings are approximately coeval. Coeval deposits among the diverse drainage settings would indicate that a broad controlling factor, such as climatic change, affected alluvial-fan aggradation. If deposits are not coeval, however, other more local factors, such as faulting or factors related to the source-area setting or geomorphic history, could have been responsible for alluvial-fan aggradation.

Thirty-two samples consisting of charcoal and wood fragments, tephra, and buried soil A horizons from FLV fan deposits define eight depositional pulses within four Holocene surficial units (Q7, Q6, Q5, Q4) (table 1). Most of the dates of the depositional pulses represent the average of two or more ^{14}C dates and estimated tephra ages that overlap at one standard deviation. These subunits were interpreted as discrete, basinwide depositional pulses. Because the four areas studied have similar geomorphic and soil sequences (Slate, 1992) and the dates among the units from different drainages agree, deposition of the FLV alluvial fans may be controlled mostly by climate.

A comparison of depositional pulses recorded by dated Holocene alluvial-fan deposits in FLV to proxy paleoclimatic records from nearby areas (fig. 2) indicated an influence of Holocene climatic fluctuations on alluvial-fan deposition. The records of areas in the adjacent White Mountains (LaMarche, 1973), northern Eureka Valley just south of FLV (Spaulding, 1985), and two sites in the Sierra Nevada—100 km to the west (Curry, 1969, 1971; Burke and Birkeland, 1983) and 150 km to the southwest (Scuderi, 1987)—were used for this comparison. These data indicate that three pulses occurred during times of relative warmth, two pulses occurred during climatic transitions from cold to warm, two pulses occurred during cold or cooler climates, and one pulse had an uncertain correlation (table 2). Thus, Holocene fan deposition in FLV was associated with warm periods or transitions from cold to warm periods in 63 percent of the cases.

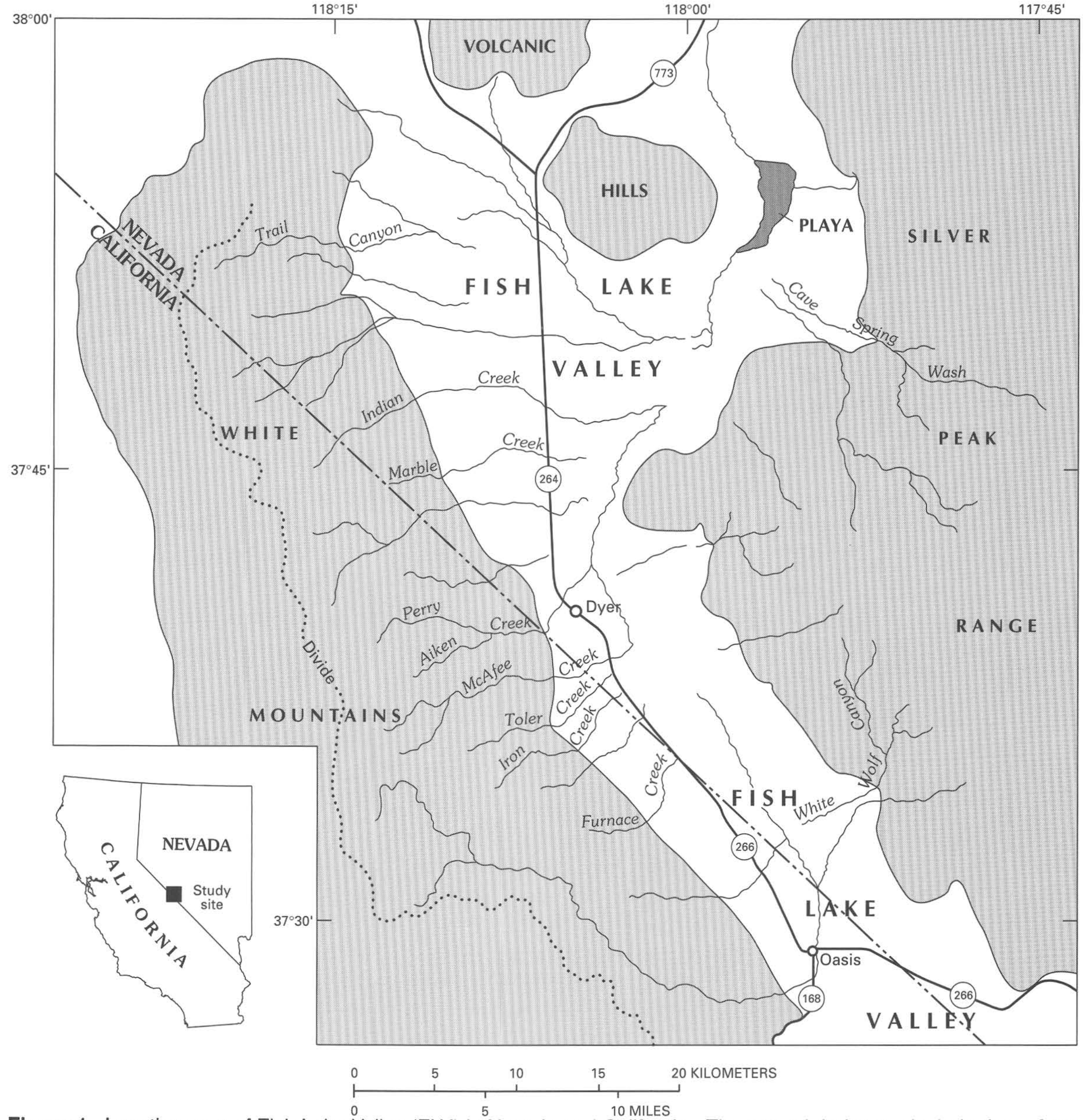

Figure 1. Location map of Fish Lake Valley (FLV) in Nevada and California. The named drainages include those from which dates were obtained.

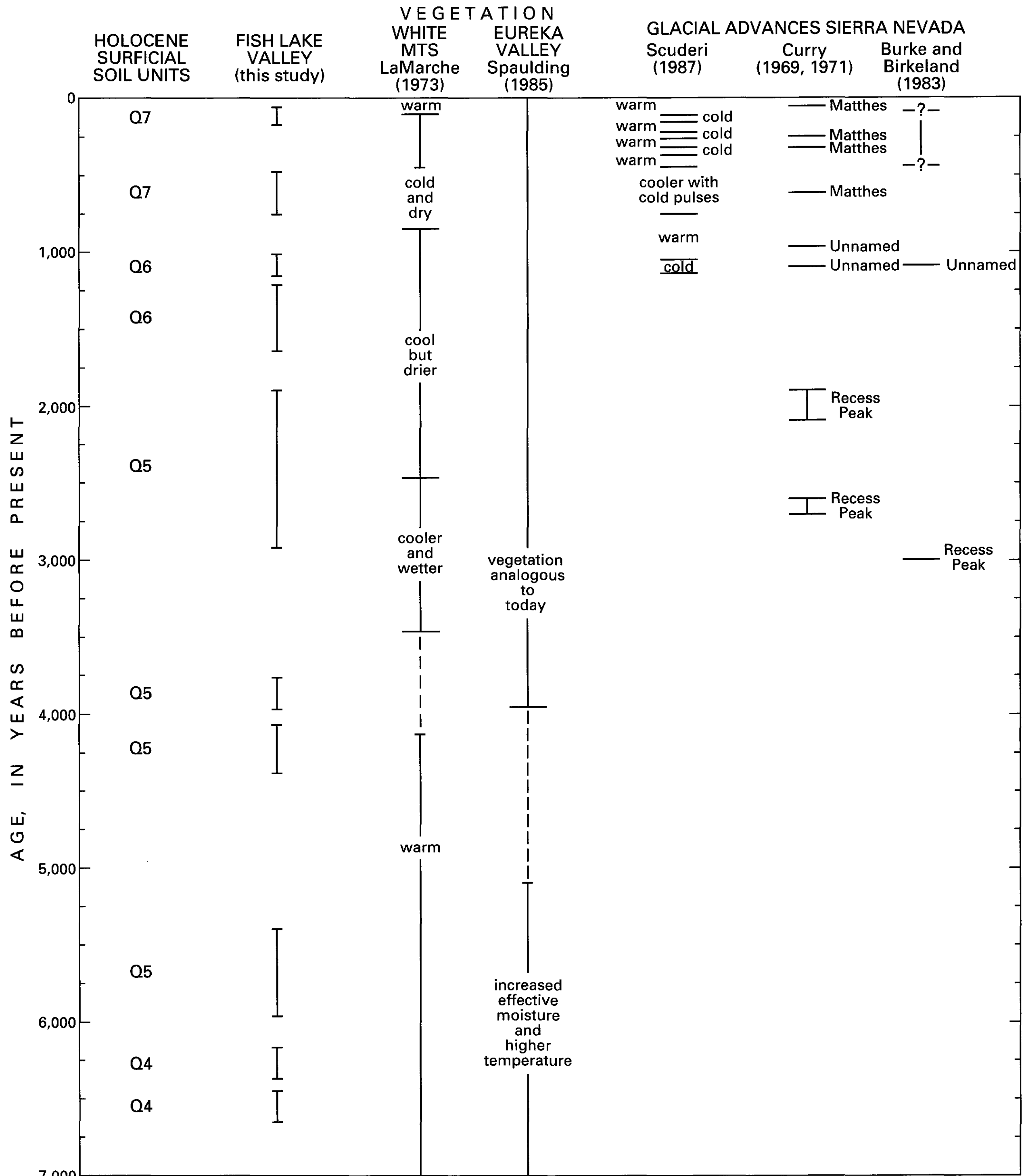

Figure 2. Chronology of Fish Lake Valley depositional pulses compared to proxy paleoclimatic records of the White Mountains, Eureka Valley, and Sierra Nevada during the last 7,000 yr B.P. Horizontal lines that terminate vertical lines in this figure indicate the time period over which a soil-forming, climate, vegetation, or glacial event occurred.

Table 1. Geochronologic controls on alluvial-fan-deposit ages

Surficial unit (age)	Date[1]	Technique (material)	Depth below surface (cm)	Area	Field number	Laboratory number[2]	Comments
Q7	110±60 yr B.P.	^{14}C (charcoal)	80	Trail Cyn	FLV-C-11	QL-4370	3.1 km upstream from soil 24-TC at 1,805 m[3]
(Late Holocene)	120±45 yr B.P.	^{14}C (wood)	5-26	Marble Cr	5-TS-1-51MC-I	Beta-26169	Site RC-1 (Sawyer, 1990)
	<150 yr B.P.	AMS-^{14}C (charcoal)	65	Trail Cyn	FLV-C-25	LL-858	Soil profile JH-6.5A[4]
	520±50 yr B.P.	AMS ^{14}C (charcoal)	65	Trail Cyn	FLV-C-15	TO-1667	Soil profile JH-6.5A
	640±40 yr B.P.	EMA (tephra)[5]	200	Marble Cr	5-TS-1-7MC-C	T145-1	Site RC-1 (Sawyer, 1990)
	660±50 yr B.P.	^{14}C (wood)	80	Marble Cr	5-TS-1-7MC	A-4765	Site RC-1 (Sawyer, 1990)
	680±65 yr B.P.	^{14}C (wood)	200	Marble Cr	5-TS-1-49-B	A-5068	Site RC-1 (Sawyer, 1990)
	1,090±50 yr B.P.	AMS ^{14}C (charcoal)	~200	Trail Cyn	FLV-C-16	TO-1668	Composite of soil sites 24-TC, JH-7
Q6	1,065±55 yr B.P.	^{14}C (wood)	40-50	Marble Cr	5-TS-1-52	Beta-26170	Site RC-4 (Sawyer, 1990)
(Late Holocene)	1,150±1.15 yr B.P.	EMA (tephra)	70	Trail Cyn	FLV-70-TC	T195-4	Surface disturbed; not true depth
	1,150±1.15 yr B.P.	EMA (tephra)	80	Toler Cr	FLV-67-MA	T195-1	Distal fan
	1,150±1.15 yr B.P.	EMA (tephra)	100	Cave Spg Wash	FLV-64-CS	T170-7	NE side of FLV; Rhyolite Ridge Quad
	1,150±1.15 yr B.P.	EMA (tephra)	~200	Iron Cr	FLV-68-IC	T195-2	Proximal fan
	1,000-2500 yr B.P.	EMA (tephra)	90	Furnace Cr	FLV-131-FC	T203-4	1.0 km downstream of soil 18-FC at 1,575 m
	1,410±200 yr B.P.	^{14}C (charcoal)	100	Indian Cr	FLV-C-6	W-6181	1-cm-thick reworked lens
	1,420±210 yr B.P.	^{14}C (charcoal)	N/A	Indian Cr	FLV-C-3	W-6168	Base of colluv. wedge overlying deposit
	1,555±45 yr B.P.	^{14}C (wood)	50	Marble Cr	5-TS-1-28FC	A-4768	Site RC-3 (Sawyer, 1990)
	6,845±50 yr B.P.	EMA (tephra)	490	Furnace Cr	FLV-71-FC	T-195-5	Southernmost Mt. Mazama ash locality
Q5	1,938±50 yr B.P.	AMS ^{14}C (charcoal)	~100	Indian Cr	FLV-C-5	AA-4886	In-situ burn buried by debris flow
(Middle	2,155±50 yr B.P.	^{14}C (wood)	220	Marble Cr	5-TS-1-22A	A-4767	Site RC-2 (Sawyer, 1990)
Holocene)	2,340±55 yr B.P.	^{14}C (wood)	~500	Marble Cr	5-TS-1-10MC	A-4766	Site RC-2 (Sawyer, 1990)
	2,400±500 yr B.P.	^{14}C (charcoal)	~200	Indian Cr	FLV-C-4	W-6179	In-situ burn, 1 m below FLV-C-5
	3,600±3.6 yr B.P.	EMA (tephra)	87-89	Trail Cyn	FLV-135-TC	T203-5	Soil JH-6A; 2.4 km dn fan of 141-TC
	3,600±3.6 yr B.P.	EMA (tephra)	~630	Trail Cyn	FLV-141-TC	T-203-7	Surface disturbed; depth uncertain
	3,600±3.6 yr B.P.	EMA (tephra)	~500	Trail Cyn	FLV-142-TC	T203-8	0.9 km down fan of 141-TC
	3,860±100 yr B.P.	AMS ^{14}C (charcoal)	~510	Trail Cyn	FLV-C-24	LL-859	10 cm below FLV-142-TC

Table 1. Geochronologic controls on alluvial-fan-deposit ages—Continued

Surficial unit (age)	Date[1]	Technique (material)	Depth below surface (cm)	Area	Field number	Laboratory number[2]	Comments
	4,213±157 yr B.P.	AMS ^{14}C (charcoal)	50	Indian Cr	FLV-C-7	AA-4907	In-situ burn buried by debris flow
	5,680±250 yr B.P.	^{14}C (charcoal)	220	Indian Cr	FLV-C-2	W-6166	1.7 m below FLV-C-7
Q4	6,260±100 yr B.P.	^{14}C (charcoal)	95	Trail Cyn	FLV-C-14	QL-4369	Soil profile JH-5C
(Pleistocene/	6,550±100 yr B.P.	^{14}C (charcoal)	330	Indian Cr	FLV-C-13	QL-4371	3-cm-thick reworked lens
Holocene (?) or	7,500±100 yr B.P.	TL (Ab horizon)[6]	133-140	Trail Cyn	FLV-L-6-TC	ITL-270	Buried by Q4 alluv; soil JH-5B
Early Holocene(?))	11,000±2,000 yr B.P.	TL (Ab horizon)	133-137	Trail Cyn	FLV-L-9-TC	ITL-282	Buried by Q4 alluv; soil JH-5B

[1]Carbon-14 and TL error ranges reported at 1σ level. ^{14}C dates provided by M.C. Reheis, USGS-Denver, Colorado; J.W. Harden, USGS—Menlo Park, California; and T.L. Sawyer, University of Nevada-Reno.

Dates of Sawyer (1990) are calibrated according to schemes of Stuiver and Pearson (1986).

[2]Key to numbers for ^{14}C Dates:

A-# = University of Arizona, Tucson

AA-# = Accelerator mass spectrometry (AMS) date, University of Arizona, Tucson.

Beta-# = Beta Analytic, Inc., Coral Gables, Florida.

LL-# = Lawrence Livermore Laboratory, Livermore, California

QL-# = University of Washington, Seattle

TO-# = University of Toronto, Ontario, Canada.

W-# = U.S. Geological Survey, Reston, Virginia

[3]Soil profiles with suffixes keyed to localities (FC=Furnace Creek, TC=Trail Canyon) were described by J.L. Slate.

[4]Soil profiles with JH prefix are at Trail Canyon, and were described by J.W. Harden or J.L. Slate, or both.

[5]Electron microprobe analyses (EMA) of volcanic glass done by C.E. Meyer, USGS-Menlo Park, California.

[6]Thermoluminescence (TL) analyses done by J.L. Slate at INSTAAR facility, University of Colorado-Boulder.

Table 2. Average age ranges of depositional pulses within Holocene surficial units (Q7, Q6, Q5, Q4) in Fish Lake Valley, number of carbon-14 and tephra samples providing age control, and inferred paleoclimate for those times based on records in the nearby White Mountains, Eureka Valley, and Sierra Nevada

[--, no samples provide age control for that depositional pulse.]

Unit	Age in years before present	Number of samples providing age control		Inferred paleoclimate
		^{14}C	Tephra	
Q7	110 ± 55	3	--	Transitional
	600 ± 50	3	1	Cooler or cold
Q6	1,100 ± 55	2	5	Transitional
	1,450 ± 205	3	--	Uncertain
Q5	2,200 ± 500	4	--	Cool
	3,900 ± 300	2	3	Warm
	5,700 ± 250	1	--	Warm
Q4	6,550 ± 200	2	1	Warm

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Preliminary Report on the Ostracod Record from Pyramid Lake, Nevada

Alison J. Smith, Department of Geology, Kent State University

A preliminary investigation of the ostracod record from Pyramid Lake sediments (trenches PLT87, PLT91-2; cores PLC87-2, PLC87-4, PLC91-1, and PLC91-2) indicated a faunal assemblage similar to that indicated by records from Walker Lake, Nevada (Bradbury and others, 1989), and the San Agustin Plains, New Mexico (Markgraf and others, 1983; Phillips and others, 1992). This similarity was offset by the unusually low abundance of ostracods in the Pyramid Lake samples as compared to samples from Walker Lake. Therefore, the ostracod record from Pyramid Lake is not as detailed as the records obtained from other previously studied saline systems.

An analysis of 150 samples obtained from the 64.5-m record of sediment cores indicated that the dominant species throughout most of the history of Pyramid Lake were alternately *Limnocythere ceriotuberosa* and *Limnocythere bradburyi* (fig. 1). These ostracods are benthic, euryhaline species that inhabited saline water in which calcium was depleted relative to bicarbonate (Forester, 1983, 1985, 1987; Smith, 1993). Both species are living today in calcium-depleted, bicarbonate-enriched saline lakes that have dissolved-solids concentrations ranging from 1,050 to 11,500 mg/L (Smith, 1993; R.M. Forester, U.S. Geological Survey, written commun., 1993). These species are not known to co-occur in the modern world. In the sediment cores from Pyramid Lake, as in the cores from Walker Lake and the San Agustin Plains, there was some overlap in the ostracod distributions that may be due to the sediment accumulation rates and sediment mixing. The species *L. bradburyi* is confined today to shallow saline lakes in the Southwest United States and Mexico, and its northernmost known occurrence is Rhoad Forks Playa, New Mexico (R.M. Forester, oral commun., 1993). The species *L. ceriotuberosa* is distributed in deep and shallow lakes north of the frost line and extends into the northern Midwest United States and northward into the Canadian Great Plains (Delorme, 1971; Smith, 1993).

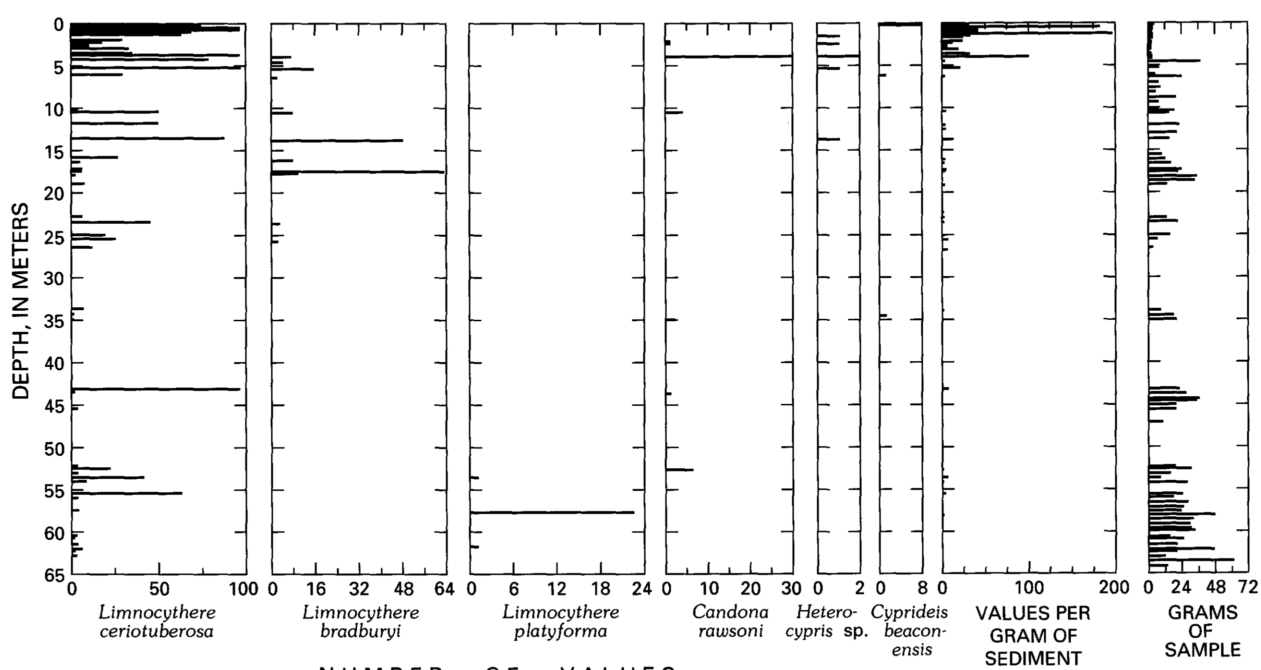

Figure 1. Fossil ostracod abundance with depth in Pyramid Lake cores.

These distributions are strongly indicative of a temperature control on the life cycles of the two species. The species *L. bradburyi* may be unable to reproduce and thrive in regions that have seasonal variation in temperature, whereas *L. ceriotuberosa* may be unable to reproduce and thrive in regions that do not have strong seasonal temperature fluctuations (Forester, 1987).

The oxygen isotope values from the Pyramid Lake carbonates can be compared with the distribution of the two ostracod species in the cores. When *L. bradburyi* is abundant, the associated carbonates are isotopically light, whereas when *L. ceriotuberosa* is abundant, the associated carbonates are isotopically heavier. Because both species inhabit lakes that have similar water composition and concentration and, yet, are geographically incompatible in the modern world, one interpretation is that these species are indicators of temperature changes. On the basis of that interpretation, the *L. bradburyi* peaks that coincide with the isotopically light values from the carbonates would indicate warmer air and water temperatures than the *L. ceriotuberosa* peaks that precede and follow *L. bradburyi*.

The presence of *L. platyforma* at the base of the core sequence is an additional point of interest. The species *L. platyforma* is a rare species in the modern world. Modern geographic locations include Canadian prairie lakes (Delorme, 1971); the species also has been reported in the Pleistocene records from the San Agustin Plains (Phillips and others, 1992). It is associated in the modern world with fresher water—more dilute water in which calcium is not depleted relative to bicarbonate—than *is L. ceriotuberosa.*

A multivariate statistical method known as the modern analog technique (Overpeck and others, 1985) was applied in this study to the ostracod record from Pyramid Lake. The modern distributions of 70 taxa and associated hydrochemistry from 250 lacustrine, spring, and wetland sites located throughout the Western United States, Mexico, midcontinent, and Alaska were used as the calibration data set. By comparing the fossil assemblage in each sample with the modern calibration data set using a distance measure [in this case, the presence-absence measure known as the Jaccard coefficient (Pielou, 1984)], and the Multivariate Statistical Package (MVSP) computer program (Kovach, 1990), a modern analog for each fossil assemblage was obtained.

The modern analogs for most of the Pyramid Lake ostracod record are modern Pyramid Lake and Walker Lake. However, the best modern analog for the *L. bradburyi* assemblages that are present in cores at 10.21, 17.15, and 17.57 m is Rhoad Forks Playa, New Mexico, which implies a time in which annual air temperature was higher, seasonality was decreased (warmer winters), and runoff or surface input decreased. In the past, *L. bradburyi* may have been present in lakes that were much deeper than the playas in which the species is present today and as represented in the modern data set. Large lakes that have an extensive pan-shaped littoral zone could provide a habitat suitable for *L. bradburyi*, as long as the temperature and seasonality limitations are met. There is no modern analog for the *L. platyforma* assemblage at the base of the core sequence because there is no *L. platyforma* in the modern data set. However, based on the limited information available on this species, it can be inferred that the lake basin probably was receiving more surface runoff or input, the water was fresher, and seasonality was enhanced (cooler winters).

These results correspond well with the oxygen isotope record in the cores. In general, the heavier isotope values are associated with *L. ceriotuberosa* and *L. platyforma*, indicating cooler annual temperatures and enhanced seasonality. The lighter isotope values are associated with *L. bradburyi*, indicating a change to warmer air temperatures and decreased seasonality.

One difficulty with the Pyramid Lake ostracod record is that it is relatively sparse compared with records from other saline lakes, such as Walker Lake and Mono Lake. This paucity of ostracods may be the result of location of the core sites away from a sediment focus point or of post-depositional dissolution of the ostracods. An additional problem with the ostracod record is the high abundance of reworked material, indicating that many of the shells have been transported from other depositional sites. Continued analysis of more samples from Pyramid Lake may produce samples that have higher ostracod abundances. Isotopic analysis of the ostracod shells could provide a direct link to the faunal-climate record.

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Lithologic Variations in Core OL92 from Owens Lake, California, and Their Probable Climatic Significance

George I. Smith and James L. Bischoff, U.S. Geological Survey
Jonathan Glen, University of California, Santa Cruz

Owens Lake, in southeast California, was 15 m deep and covered about 290 km^2 in 1872 when it was the terminus of the Owens River, which drains the east side of the high Sierra Nevada. Although it has been dry since about 1920 as a result of construction of the Owens Valley Aqueduct in the early 1900's, Owens Lake was the first in a chain of as many as five lakes that were created as a result of the increased volume of water received during the pluvial periods of the Pleistocene. To obtain a more complete record of variations in runoff during these pluvial periods, three 7.6-cm-diameter cores were obtained from adjacent sites on Owens Lake in late spring 1992. The three cores extend from 0.00 to 7.16 m (100 percent recovery), 5.49 to 61.37 m (80–85 percent recovery), and 61.26 to 322.86 m (75–79 percent recovery); and together, these cores represent about 800,000 yr of sedimentation. The core site is in the south-central part of the now-dry lake (SE corner of sec. 9, T. 18 S., R. 37 E., Mount Diablo Meridian) and about 4 km from the nearest pre-1900's shore.

Except for 2 to 3 m of salts on the surface, which formed during this century, all of the sediments appear to be perennial lake deposits. A 4.2-m-thick oolite bed underlies the salts, and a ^{14}C date on humus in the dark silt or clay immediately beneath the oolite layer indicates that its deposition began shortly after 8,930 ± 70 yr B.P. Below the oolite bed, clay, silt, and a little sand characterize the underlying 200± m of lake fill. Fine to coarse sand and less common silt and clay characterize the lowest 100± m of sediments. Notable concentrations of fish scales and bones, aquatic mollusk shells, and plant fragments and seeds occur in two 20-m-thick zones centered near 200- and 300-m depths. A thick bed of volcanic ash (303.5 to 309.1 m) has been identified as airfall and reworked Bishop Tuff (758,000 ± 2,000 yr B.P.); the top of the Brunhes-Matuyama magnetic polarity reversal is near 320 m.

Many of the fine sediments in the upper 200± m of core are moderately well bedded, with clast size or subtle color changes defining the beds. Many beds are massive or mottled (bioturbated?), and laminar bedding is rare. In this interval, beds of coarse to very coarse sand were noted near 67 m (0.25 m thick) and 95 m (1.25 m thick); the deeper bed contained methane under high pressure. The grain sizes of sediments in the lower 100 m of the core are more variable and include lacustrine(?) sediments ranging from clay to very coarse sand that may include a few granules or very small pebbles. Besides the Bishop Tuff, several additional beds tentatively identified as tephra are being studied.

The 200± m of fine-grained sediments underlying the near-surface oolite bed appear to represent deep-water perennial-lake deposition since about 500,000 yr B.P. (using the apparent overall perennial-lake sedimentation rate—2,500 yr/m—for estimating their age). The two coarse sand beds in this interval apparently represent nonclimatic events, because neither bed is underlain or overlain by sediments indicating a shallowing or deepening of the lake. This would be expected in a transitional stage before or after a brief, climatically caused shallow-water episode (study of the fossil content of these zones may change this interpretation). The lowest 100± m of coarser, more variable sediments, including four zones containing coarse to very coarse lacustrine sand, suggest a 300,000-yr period of fluctuating lake levels that included four or more shallow, but not dry, periods, although a flood-plain environment cannot be ruled out at this time. However, this 100± m interval also includes repeated zones of interbedded clay, silt, and very fine sand that show that periods when the lake was also deep and possibly overflowing recurred many times—an interpretation that is compatible with the record from Searles Lake, which was downstream from Owens Lake during pluvial periods.

In the late 1800's, Owens Lake waters had a salinity of 6 to 9 percent. The oolite bed near the surface is interpreted to represent a period of saline lakes that were (1) shallow (allowing wind energy to produce currents on the bottom) and (2) constantly saturated or supersaturated with $CaCO_3$ (so that carbonate deposition would overwhelm clastic deposition). Assuming that the lower 100 m of sediments are eventually determined to be lacustrine, the absence of similar oolite beds in the deeper parts of the OL92 core sequence is interpreted to mean that, in this area (and probably in a much larger area as well), the past 9,000 yr have been more arid than during any period in the past 800,000± yr and have been characterized by several high-latitude glacial-interglacial cycles.

Sedimentary Record of Climatically Induced Lake-Level Fluctuations, Pyramid Lake, Nevada

Joseph P. Smoot, U.S. Geological Survey

Pyramid Lake is a large, perennial, terminal lake that was part of the late Pleistocene Lake Lahontan. Shoreline deposits and outcrops of lacustrine material attest to dramatic changes in the depth and area of the lake during the past few thousand years that are attributed to fluctuations in regional rainfall and evaporation. A core (PLC87) obtained on the north side of the lake about 1.5 km from the lake shore penetrated 65 m of unconsolidated sediment overlying Tertiary volcanic rocks. The upper 3 m of sediment were not cored, but were sampled from a trench (PLT87). The sedimentary structures in this core were analyzed to determine their sensitivity to changes in lake depth that indicate changes in climate.

Most of the sedimentary variability in core PLC87 is within seven categories that are interpreted as indicators of different lake depths. In the first category, brecciated silts and clays are interpreted as lake sediments that were desiccated during subaerial exposure. Root casts and gypsum crystals are common accessories. In the second category, oscillatory rippled sand formed in relatively shallow water (less than 3 m in the modern lake) and thin clay drapes or flasers (in clay lenses), were present in the deeper formations. In the third category, sand lenses within bioturbated clay represent storm wave deposits in conditions where settling out of silt and clay was the norm. The sand lenses range from layers of oscillatory ripples like those formed in shallower water to low-amplitude pinch-and-swell laminae that represent rolling-grain oscillatory ripples in much deeper water. The bioturbated clay contains ostracods and sand-filled burrows. The depth range for the formation of this sedimentary category depends on the surface area of the lake and the scale of storms and probably varied through time. In the fourth category, graded silt laminae in bioturbated clay are interpreted as storm lag layers in a deeper water setting. The silt laminae consist mostly of skeletal material that was dispersed within the bioturbated clay, then concentrated in layers with no orientation, indicating transport as waves winnowed the finer matrix. As in the third category, the absolute depth for this wave activity depends on the size of the lake, but the water needed to be deeper than that for the pinch-and-swell laminae. In the fifth category, silty clay containing random burrows represents deposition caused by settling out below wave base. Skeletal material in these muds is generally smaller than that in the shallower water deposits, which were possibly dominated by pelagic organisms. The bioturbation indicates that infaunal organisms randomly burrowed the upper part of the sediment, indicating oxygenated conditions. This category of structure in core PLC87 is coincident with lighter carbon and oxygen isotopes measured by L.V. Benson (unpublished data). Benson interprets the isotope shift from heavy to light as a transition from closed- to open-basin conditions as the lake exceeded spillover height. The lowest spillover that would produce this change is 37 m higher than the elevation of the core site, which can be used as a minimum depth for this deposit. In the sixth category, clays dominated by horizontal burrows and partial preservation of laminae represent conditions below the wave base where oxygen is very depleted at the sediment-water interface. The burrows are smaller in diameter than those of the randomly burrowed muds, and the skeletal debris is totally dominated by pelagic organisms; both conditions were consistent with oxygen-depleted bottom water, probably due to greater lake depths. This change is interpreted as being depth controlled rather than controlled by salinity or seasonality. Because the deposits are only transitional to randomly burrowed mud and flat-laminated mud, they have only light isotopic signatures, and their thicknesses are too large to represent deposition for periods less than several tens of years. In the seventh category, flat, continuous laminae of graded silt or alternations of clay and biogenic or calcareous material are interpreted as deposits settling out on an anoxic lake bottom, as indicated by the absence of bioturbation. The graded silt laminae are associated with pebbles and cobbles that randomly penetrate the laminae in a manner consistent with drop stones. These deposits are interpreted as representing ice cover allowing saltation of coarser material over larger areas and transportation of cobbles and pebbles in ice floes. The highest shorelines around Pyramid Lake are 150 m above the elevation of the drill site, which probably indicates the upper range of the deepest lakes.

The seven sedimentary categories are present in systematic successions that indicate the rise and fall of lake level at a variety of time scales. Transitions from deepest water deposits (categories 6 and 7) to shallowest water or subaerial deposits (categories 1 and 2) are present over intervals of about 10 m. Smaller scale changes are present rhythmically at the scales of 3 m and decimeters. The changes in sedimentary categories are gradational, except for wave deposits that commonly have abrupt contacts with underlying deeper water deposits or subaerial deposits. The sharp contacts are interpreted as erosional surfaces with depths that cannot be assessed from the core. In all of the depth-change intervals, the transgressive deposit is characteristically thinner than the regressive deposit and may have resulted from sediment depletion during transgression as lake water flooded drainages. During regression, these sediments would be eroded and transported lakeward, producing a thicker sedimentary sequence.

Core PLC87 may, in part, be correlative to the Truckee River Canyon deposits that have the Wadsworth tephra in the middle and are capped by tufa dated at about 12,500 yr B.P. The lake must have been more than 100 m deep at the core site for deposition of the clay sequence in the canyon. Therefore, sedimentary categories 6 and 7 are the only likely correlative deposits in the core. If the lake dropped to levels below 60 m deep at the core site, the canyon sequence was subaerially exposed. The difference in sedimentation rates between the core site and the canyon make relative thicknesses a poor tool for comparison.

Uranium-Series Dating of Tufas from Pyramid Lake, Nevada

Barney J. Szabo and Larry V. Benson, U.S. Geological Survey

Pyramidal tufa mounds (as much as 100 m in height) border the present shore of Pyramid Lake in Nevada. The elevations of the tops of many of these mounds correspond to elevations of intersubbasin sills. Although the late Pleistocene highstands of Pyramid Lake are well studied, the relation of the tufa mounds to lake level has not been previously documented. A reconnaissance survey of the tufa mounds indicated a variety of morphologies. Varieties of tufas sampled for dating included dense laminated, branching, nodular, platy, and concentrically laminated tubular tufas. Precipitation probably occurred where the calcium-rich ground water mixed with carbonate-rich lake water.

The U-series ages of more than 30 samples collected at elevations between 1,160 and 1,220 m indicate that at least some, and possibly most, of the pyramidal mounds formed between about 40,000 and 13,000 yr B.P. These results are consistent with unpublished ^{14}C ages of the same tufa samples that were determined by M. Kashgarian (written commun., 1993). The water in Pyramid Lake stood above 1,177 m during ≤40,000 to 30,000 yr B.P., and the Pyramid Lake and Winnemucca Lake (Nevada) subbasins were usually connected. During approximately 23,000 to 16,000 yr B.P., Pyramid Lake stood at or above about 1,207 m, indicating that lakes in the Pyramid Lake, Winnemucca Lake, and Smoke Creek-Black Rock Desert subbasins had coalesced. The U-series ages of laminated tufas that line the walls of a small cave at Marble Bluff on the southeastern part of the Pyramid Lake subbasin at an elevation of 1,204 m indicate that lakes in the Pyramid Lake, Winnemucca Lake, and Smoke Creek-Black Rock Desert subbasins probably were connected between about 400,000 and 170,000 yr B.P.

Chemical analyses of the tufa samples indicate the presence of a small acid-insoluble residue having as much as 40 μg/g uranium, 3,700 μg/g thorium, and 4,700 percent excess ^{230}Th activity relative to ^{234}U. This residue probably was transported as a colloidal suspension in ground water, which discharged through the tubular structure present in the bases of tufa mounds.

Changes in the Pyramid Lake, Nevada, Drainage Basin as Indicated by the Organic Content of Recent Sediments

Gabrielle E. Tenzer and Philip Meyers, Department of Geological Sciences, University of Michigan

Organic-matter contents of lake sediments indicate the amounts and types of biota that have inhabited lake environments. Indicators of organic-matter sources can provide information about lake productivity (changes in the amount and type of aquatic-source material) and about changes in climate, river runoff, and land-use practices (changes in the amount and type of terrigenous source indicators). A number of organic-matter-source signals have been and are presently (1993) being analyzed in sediment cores from Pyramid Lake to determine changes in the lake system. Biomarkers from core 2, collected in 1976 just to the west of Anaho Island, were analyzed by Meyers and others (1980). First, a decrease in the n-C_{29} to n-C_{17} ratio at a depth of approximately 1 m was concluded to represent a change from oligotrophy to mesotrophy at around A.D. 1300 to 1400. Second, an increase in n-alkanoic acid concentrations in the top 30 cm, unaccompanied by any such increase in alkane concentrations, was interpreted as a preservational effect resulting from an increase in the rate of sedimentation. Finally, the various geolipids within the core were postulated to be primarily terrigenous in origin and supplemented by aquatic sources. In the absence of measured sedimentation rates, interpretations from the study by Meyers and others (1980) were based on the known history of the drainage basin and on sedimentation rates of nearby Walker Lake. The ^{210}Pb dating of Pyramid Lake cores 96A, 93A, and 93L collected in 1991 allow these previous interpretations to be examined within an improved chronological framework.

A combination of isotopic, elemental, and molecular determinations have been obtained for cores 96A, 93L, and for the very top part of core 93A (fig. 1). Dry-weight percent $CaCO_3$ was analyzed by the carbonate bomb method, and values were verified using a UIC/Coulometrics CO_2 coulometer. All

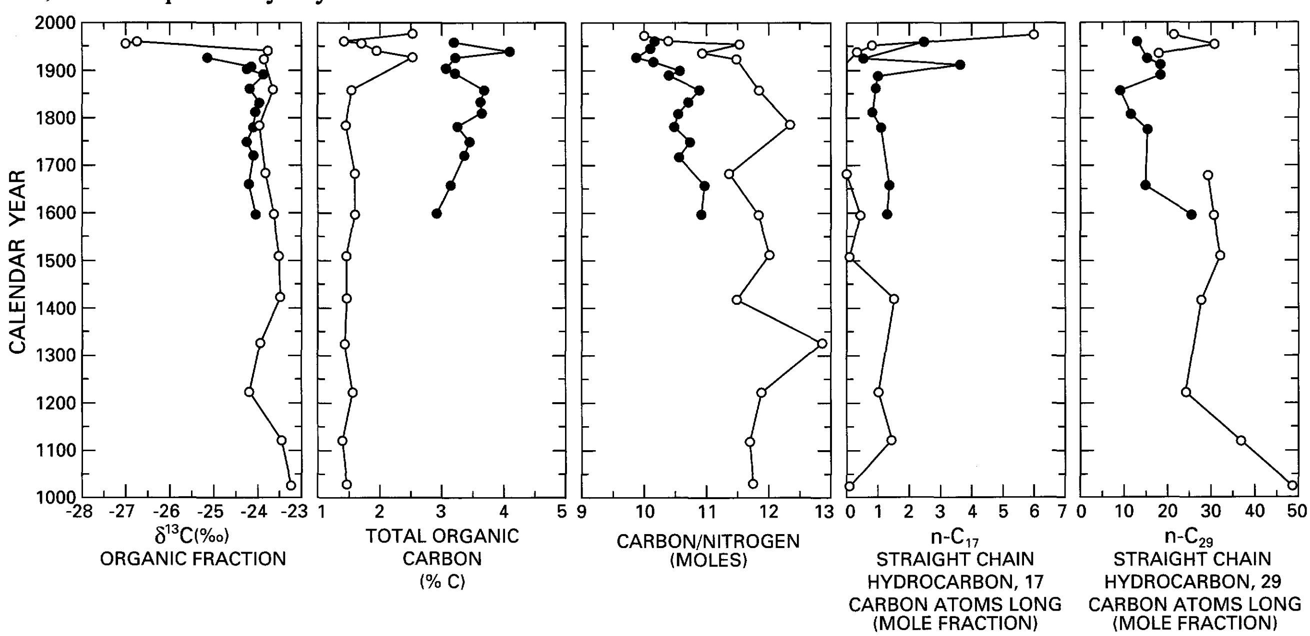

Figure 1. Selected chemical characteristics of shallow- and deep-basin cores in the Pyramid Lake drainage basin. Open circles represent cores 93A/L (shallow basin), and solid circles represent core 96A (deep basin). Isotopic values are relative to PDB; TOC values are total organic-carbon wet-weight percentages; C/N ratios are molar; and the hydrocarbon (HC) values are in 10^3 ng/g organic carbon.

inorganic carbon was assumed to be present as $CaCO_3$. Total organic-carbon (TOC) values and C/N ratios were obtained using a Perkin-Elmer 2400 CHN elemental analyzer after removal of carbonates using 3N HCl. These carbonate-free samples were analyzed for $\delta^{13}C_{org}$ using a Finnigan MAT Delta-S mass spectrometer. Organic compounds, including fatty acids and straight-chained hydrocarbons, were extracted by sonication using dichloromethane. Fatty acids were converted to their methyl esters after having been saponified using BF3. Unsaponified and saponified fractions were separated in petroleum ether and toluene, respectively, using alumina over silica gel columns. Geolipid fractions were analyzed using a gas chromatograph that had an automatic injector and a flame ionization detector.

The TOC values were higher in core 96A than in cores 93A/L. These higher values for core 96A were probably due to a finer grain-size distribution and a higher mass-accumulation rate at the deeper site. Core 96A is located farther from the Truckee River input, thereby decreasing any dilution of the TOC signal by river input, as may have been happening at the core 93A/L sites. In contrast, C/N ratios were higher in cores 93A/L, as was the long-chained hydrocarbon content (Σn-C_{27}, n-C_{29}, n-C_{31}) of cores 93A/L. Both of these signals indicated a larger amount of terrigenous input to the shallow basin of Pyramid Lake, which was to be expected considering the location of the Truckee River inlet with respect to the shallow and deep basins. On average, there is a higher aquatic-source-material signal in the deep basin as indicated by the short-chained hydrocarbon content of core 96A. A comparison of the short- and long-chained hydrocarbon contents of both cores indicated an overall domination of the organic sedimentary material of the lake by terrigenous sources, as was suggested by Meyers and others (1980). The C/N molar ratios of about 10–12 and an organic-carbon flux study for Pyramid Lake by Galat (1986), however, indicated the dominance of autochthonous sources. The predominance of long-chained hydrocarbons is probably due to the higher proportion of n-alkanes in land-derived organic material in comparison to lake algae (Cranwell and others, 1987; Goossens and others, 1989; Meyers and Ishiwatari, 1993).

Decreases in long-chained hydrocarbon content were interpreted as periods of decreasing river runoff and more arid climate conditions in the Pyramid Lake drainage basin. This interpretation is supported by some evidence from archaeological findings (Hattori and Tuohy, this volume). These periods of increasing dryness were indicated in core 96A from approximately A.D. 1600 to A.D. 1750, A.D. 1780 to 1860, and A.D. 1890 to 1960 (includes the Dust Bowl) and in cores 93A/L from approximately A.D. 1030 to 1130, A.D. 1230 to 1420, A.D. 1790 to 1945, and 1970 to about 1980 to the top of the core. A distinct increase in n-C_{29} in core 96A, which also may occur in cores 93A/L from approximately A.D. 1860 to 1920, corresponded to a period of logging and agricultural development in the Pyramid Lake drainage basin. The decrease in n-C_{29} in core 96A after 1920 may be a delayed effect of the Derby Diversion Dam, constructed in 1905, to provide irrigation in the Fallon-Carson Desert areas.

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Late Quaternary Vegetation and Stable-Isotope Record from the Lahontan Basin

Peter E. Wigand, Quaternary Sciences Center, Desert Research Institute, Reno, Nevada

Plant macrofossil and stable-isotope data from multiple strata of a single fossil woodrat (*Neotoma* spp.) den (midden) on the former western shore of pluvial Lake Lahontan extend the paleoclimatic proxy record for this area to 30,000 yr B.P. These data indicate Pleistocene depressions of distributions of plant species of as much as 1,100 m and horizontal extensions ranging as much as 30 km in northwestern Nevada.

Changes in plant community composition indicate three periods of effectively moister climate generally coincident with expansions of pluvial Lake Lahontan, approximately 24,000 yr B.P., between about 16,000 and 12,000 yr B.P., and about 10,700 yr B.P. The first two periods were typified by expansion, down to 1,380 m, of whitebark pine (*Pinus albicaulis*) into Utah juniper (*Juniperus osteosperma*) woodland. During the second period, approximately 16,000 to 12,000 yr B.P., increased importance of sagebrush (*Artemisia* sp.) in the understory was markedly greater than during the earlier period, and mountain mahogany (*Cercocarpus ledifolius*) was depressed as much as 330 m in elevation. The latest period, approximately 10,700 yr B.P., was typified by the maintenance of Utah juniper woodland in areas as much as 20 km beyond where it presently (1993) occurs.

Variations in the proportions of heavy isotopes of carbon and hydrogen in radiocarbon-dated Utah juniper from the same midden (and confirmed by stable isotopes from other middens in the region) recorded the corresponding three periods of effectively wetter climate. Coincident maximums in Utah juniper seed length probably indicated increased productivity resulting from greater effective precipitation. Greater woodrat-dung widths probably indicated increased woodrat size due to cooler temperatures. Together, these data indicate that the temperature decreased and precipitation increased at about 24,000 yr B.P. and that temperature decreased between about 14,000 and 12,000 yr B.P., but the precipitation increases were smaller than the increases that occurred at about 24,000 yr B.P. Apparent breaks in the woodrat midden record between 28,000 and 25,000 yr B.P. (last interstade), between 20,000 and 16,000 yr B.P. (glacial maximum), and between 8,000 and 4,000 yr B.P. (Altithermal) may indicate periods when environmental conditions were locally too dry, too cold, or too hot for woodrats to flourish.

Sedimentary Facies, Quaternary History, and Paleoclimatic Implications from Sediments in Mohawk Valley, California

James C. Yount, David S. Harwood, John P. Bradbury, Andrei M. Sarna-Wojcicki, and Charles E. Meyer, U.S. Geological Survey

Mohawk Valley is along the Middle Fork of the Feather River between the Sierra Valley and Quincy, California. The floor of Mohawk Valley is at approximately the 1,280-m elevation, at the intersection of the Feather River and Sulphur Creek. The crest of the Sierra Nevada is west of Mohawk Valley, and summit elevations reach 2,589 m. Typical elevations are in the range of 1,980 to 2,290 m. Although the Sierra Nevadan crest is west of Mohawk Valley, the actual drainage divide separating the westward-flowing Feather River from drainage to the east is at Beckworth Pass on the east edge of the Sierra Valley. This eastward position of the drainage divide indicates past capture of the Sierra Valley drainage by the Feather River.

Tertiary volcanic rocks, including thick sections of intermediate composition lahar, debris flow, and lava, surround Mohawk Valley on the north and east. Pre-Tertiary metavolcanic, metasedimentary, and plutonic rocks are west and south of the valley. These pre-Tertiary rocks underlie most of the area that was glaciated to the southwest of Mohawk Valley. The contrast in source-rock lithology across the valley aids in separating glacial outwash from fluvial Feather River deposits, particularly in the older part of the Quaternary section.

In the 1890's, Turner (1891) interpreted the sedimentary section at Mohawk Cliff as lacustrine and assigned the strata there the name Mohawk Lake Beds. This interpretation was continued by Durrell (1987), who postulated a lake filling Mohawk Valley whose shore followed the 1,585-m contour. The lake, impounded by a large landslide at Consignee Creek, may have had a water depth of as much as 150 m in its central part. The original outlet of the lake may have been located at Poplar Creek, a few kilometers southwest of the present (1993) exit of the Feather River out of Mohawk Valley. Durrell (1987) suggested that headward erosion by the Feather River through the Consignee Creek landslide may have been completed sometime during the late Pleistocene or Holocene and that this breach may have resulted in rapid lake emptying and subsequent erosion and exposure of the sedimentary fill in Mohawk Valley.

The studies discussed in this report confirmed that erosion has recently exposed the sediments of Mohawk Valley, but most of these sediments may have accumulated in meadow or braided-stream environments, and there may have been periodic contributions of glacial outwash from ice advancing out of the Gold Lakes and Lakes Basin area southwest of Mohawk Valley. The only lacustrine sediments that were recognized in the valley-fill sequence were proglacial sediments of limited extent that are associated with the glacial outwash facies. Rather than rapid exhumation of the section following the emptying of the hypothesized lake by stream capture, recent tectonic activity along the Quaternary faults that bound Mohawk Valley may have altered the Feather River grade and encouraged downcutting.

The eastern arm of Mohawk Valley, extending from Clio to Portola, contains as much as 100 m of trough cross-bedded cobble to pebble gravel and planar and trough cross-bedded coarse and medium sand, which is interpreted as braided-stream deposits. Sections exposed in the western arm of Mohawk Valley consist, in their lower parts, of pebble gravel, which is dominated by pre-Tertiary clasts and is interpreted as glacial outwash that is overlain by massive organic-rich silt and clay interbedded with blocky to fissile peat beds as much as 1 m thick. Diatom assemblages are dominated by benthic species of *Cocconeis, Cymbella, Epithemia, Eunotia, Fragilaria, Gomphonema, Navicula, Pinnularia, and Surirella*, indicating fresh marsh environments that had very shallow water depths of 1 m or less. The peaty intervals contain the Rockland Ash (400,000 yr B.P.) in their lower parts and Summer Lake LL (180,000 yr B.P.) and Tule Lake 2079, 264, or 2036 (155,000–160,000 yr B.P.) tephra layers.

The fine-grained sections are overlain by glacial outwash that was associated with ice descending into the valley from the Gold Lakes and Lakes Basin region to the southwest. Proglacial lacustrine deposits of limited lateral extent are present within the outwash complexes as indicated by varved fine-sand and silt couplets, poorly sorted quartz-rich silt beds containing dropstones, and contorted beds of diamict grading laterally into slump blocks surrounded by wood-bearing silt and silty sand. A log, buried in diamict and slumped into a proglacial lake and lying approximately 3 km downstream from the Tioga Stage ice termini in Jamison and Gray Eagle Creeks, yields an age of 18,715 ± 235 yr B.P.

Comparison with the eastern Sierra Nevada glacial chronology indicates that the late Wisconsin deposits are well correlated. Correlation also may exist between the oldest outwash in Mohawk Valley and the Sherwin Till, although tills with Sherwin properties in the Gold Lakes and Lakes Basin area have not been identified.

The peat-rich section in western Mohawk Valley, spanning the age range of approximately 500,000 to 155,000 yr B.P., has no apparent stratigraphic breaks and also presents no evidence of outwash correlatable to the older Tahoe (Tahoe II), Mono Basin, or pre-Mono Basin tills of the eastern Sierra Nevada.

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