

Climatic Oscillations 10,000–155,000 yr B.P. at Owens Lake, California Reflected in Glacial Rock Flour Abundance and Lake Salinity in Core OL-92

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Received March 25, 1997

Chemical analyses of the acid-soluble and clay-size fractions of sediment samples (1500-yr resolution) reveal oscillations of lake salinity and of glacial advances in core OL-92 back to 155,000 yr B.P. Relatively saline conditions are indicated by the abundance of carbonate and smectite (both pedogenic and authigenic), reflected by Ca, Sr, and Mg in the acid-soluble suite, and by Cs₂O, excess MgO, and LOI (loss on ignition) in the clay-size fraction. Rock flour produced during glacial advances is represented by the abundance of detrital plagioclase and biotite in the clay-size fraction, the ratio of which remains essentially constant over the entire time span. These phases are quantitatively represented by Na₂O, TiO₂, Ba, and Mn in the clay fraction. The rock-flour record indicates two major ice-advances during the penultimate glacial cycle corresponding to marine isotope stage (MIS) 6, no major advances during the last interglaciation (entire MIS 5), and three major advances during the last glacial cycle (MIS 2, 3, and 4). The ages of the latter three correspond rather well to ³⁶Cl dates reported for Sierra Nevada moraines. The onset of the last interglaciation is shown by abrupt increases in authigenic CaCO₃ and an abrupt decrease in rock flour, at about 118,000 yr B.P. according to our time scale. In contrast, the boundary appears to be gradual in the $\delta^{18}\text{O}$ record in which the change from light to heavy values begins at about 140,000 yrs B.P. The exact position of the termination, therefore, may be proxy-dependent. Conditions of high carbonate and low rock flour prevailed during the entire period from 118,000 yr B.P. until the glacial advance at 53,000 yr B.P. signaled the end of this long interglaciation. © 1997 University of Washington.

INTRODUCTION

Owens Lake, located at the eastern base of the central Sierra Nevada (Fig. 1), was the terminus of the Owens River prior to the lake's complete desiccation shortly after 1913 due to river diversion by the City of Los Angeles. During

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earlier wetter cycles, the lake overflowed to fill a series of downstream basins including China Lake Basin, Searles Valley, Panamint Valley, and ultimately, Death Valley (Smith and Street-Perrott, 1983). In 1992 the U.S. Geological Survey drilled a 323-m-deep core (OL-92) into Owens Lake sediments near the depocenter of the basin to obtain a continuous record of silty-clay sediment spanning the last 800,000 yr. A multiparameter reconnaissance study of the entire core (ca. 7000-yr resolution), reported in a 13-chapter summary volume (Smith and Bischoff, 1997), revealed cycles characterized by closed and overflowing lake conditions.

We report here the results of a second part of an intermediate-resolution study aimed at resolving shorter-term events over just the past 155,000 yr, using sediment samples of ca. 1500-yr resolution taken over the uppermost 90 m of OL-92. The companion study (Menking *et al.*, 1997) reports on grain size, TIC (total inorganic carbon) content, and stable isotopic measurements, proxies which tend to reflect variations in water flow and which are most representative of variations during closed-lake conditions. The present study was concerned primarily with seeking a proxy for glacial advances. We found that the mineralogy and chemistry of the clay-fraction provide indices of glacial rock flour. We also report on the composition of the acid-soluble carbonate fraction which gives further detail on closed-lake conditions.

PREVIOUS WORK

The reconnaissance study of Smith and Bischoff (1997) provided an approximate time–depth model for the entire core, based on radiocarbon dates from the top 31 m, the Bishop Ash (759,000 yr, Sarna and Pringle, 1992) at 304 m, ten within-Brunhes paleomagnetic excursions, and a compaction-corrected mass-accumulation rate of 51.4 g/cm/1000 yr (Bischoff *et al.*, 1997b). Application of this model to observed sediment parameters indicates that Owens Lake was saline, alkaline, and biologically productive at times of decreased water-flow, and was generally hydrologically flushed and relatively unproductive during times of increased

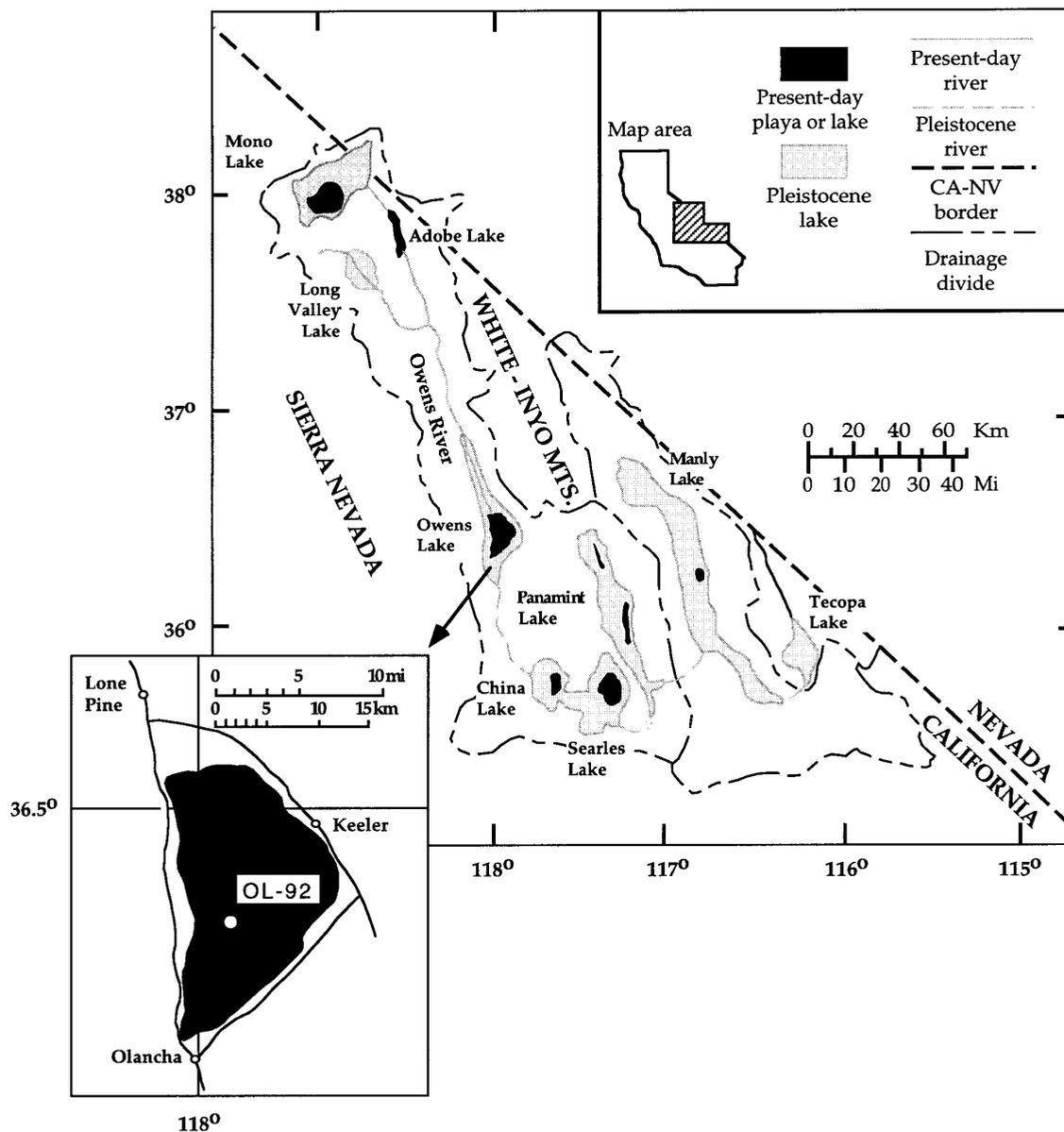


FIG. 1. Location of Owens Lake and core OL-92. Extent of modern lakes and playas indicated by dark gray pattern, maximum pluvial lakes by light gray outline.

water-flow. Grain size, abundance of CaCO_3 , organic carbon, clay mineralogy, cation-exchange capacity of the clay fraction, fossil pollen, fish, ostracodes, and diatoms (see summary by Smith *et al.*, 1997) all show cyclic variation down the core. CaCO_3 abundance, in particular, strongly reflects an approximately 100,000-yr dominant cycle, characteristic of global ice-volume indicated by the MIS $\delta^{18}\text{O}$ record. Four of the last five marine isotope terminations are clearly shown in the OL-92 record.

At the other end of the resolution spectrum, Newton (1991) intensively studied three piston cores (4 to 11.5 m in length)

taken at Owens Lake near OL-92 for lithology, mineralogy, magnetics, and diatom stratigraphy to define the late-Pleistocene–Holocene stratigraphy. Following on this work, Benson *et al.* (1996) carried out a similar study on a pair of longer piston cores (33-m total length), the sediments of which span the maximum glacial conditions of the past 50,000 yr. They report on organic and inorganic carbonate, $\delta^{18}\text{O}$, and magnetic susceptibility. These workers interpreted the fine-scale susceptibility peaks as reflecting glacial advances, the ages of several of which correspond to ^{36}Cl age estimates of Sierra Nevada moraines reported by Phillips *et al.* (1996).

In our companion study (Menking *et al.*, in press), proxies of grain size, carbonate content, and $\delta^{18}\text{O}$ all revealed many high-frequency oscillations and showed that dry conditions of the last interglacial appear to span the entire period from 50,000 to 120,000 yr B.P. This dry period was punctuated by numerous short periods of overflow. We also noted that the timing of the Termination II (MIS 5/6 boundary) is proxy dependent, with the carbonate and grain-size records showing an abrupt change at ca. 120,000 yr B.P., while the $\delta^{18}\text{O}$ shows a gradual increase after ca. 140,000 yr B.P.

PHILOSOPHY OF APPROACH

With a few notable exceptions, the carbonate fraction of the sediment seems to be a faithful monitor of water residence time in Owens Lake, and its abundance and its chemical and isotopic composition record periods of relative aridity with a high degree of resolution. In closed-lake conditions, CaCO_3 is trapped within the basin (Bischoff *et al.*, 1997a), whereas during overflow, it is transported downstream to precipitate in the lower basins. During times of increased water-flow the lake was flushed and biologically unproductive, so CaCO_3 does not serve as a proxy for high-resolution events such as glacial advances.

For these overflow periods we sought a proxy for glacial advances in rock-flour abundance in the sediments. The primary runoff to the Owens River is from the Eastern High Sierra, where bedrock is dominated by granitic plutons (Bateman, 1961) containing plagioclase and biotite. The bulk chemistry of the Owens Lake sediments, on a carbonate-free basis, was found to be close to that of a granodiorite, indicating that most of the sediments come from the Sierra Nevada (Bischoff *et al.*, 1997a). We reasoned that a rock-flour component would be most recognizable by the abundance of biotite and plutonic feldspar in the clay-size fraction, presuming that only glacial abrasion can reduce these minerals to clay size. In contrast, we reasoned that in the absence of glaciation, the clay-size fraction should be dominated by weathering processes. While X-ray diffraction (XRD) provides a semiquantitative measure of biotite, we found that Na and Ca contents provide a quantitative measure of plagioclase. Except for ion-exchange sites, Na and Ca do not occur in any of the common pedogenic clay minerals, and, therefore, measured solid-phase Na and Ca should be entirely from detrital plagioclase. Thus, we saturated our clay-fraction splits with Cs^+ ion to flush all exchangeable Na and Ca from the samples (Beetem *et al.*, 1962). The amount of Cs taken up by the sample, therefore, is a measure of its cation-exchange capacity (CEC), which, in turn, should be a measure of the relative abundance of weathering-zone clay minerals such as smectite and a measure of warm weathering conditions.

METHODS

Channel samples, each 70-cm long (total of 108 samples, each representing ~ 1500 yr), were collected from the core by passing a U-shaped spatula down the split face of the core. Care was taken to avoid sampling slumped or otherwise questionable segments of the core, more common in the upper 20 m (30,000 yr) where the sediment was uncompacted and less competent. In addition, samples of modern Owens River mud were collected at seven locations (see Menking *et al.*, 1997) ranging from the headwaters in Long Valley down to the lower-most reaches of the river, to represent an approximation of the Holocene river input to Owens Lake. Samples were dried, lightly ground in an agate mortar, and then repeatedly washed with hot deionized-water to remove interstitial salts. Aliquots were separated for the acid-soluble fraction and for the clay fraction of the present study. At the same time we separated aliquots for carbonate content, stable isotope, and grain-size analyses for the companion study. The acid-soluble split was leached in 3N HCl overnight and analyzed for Ca, Mg, and Sr by standard atomic-absorption spectroscopy. Organic matter and carbonate were removed from the clay fraction splits by soaking 48 hr in a combination sodium acetate-acetic acid solution buffered at pH 5 with 5% hydrogen peroxide, following procedures of Jackson (1969). The <2 micrometer fraction was isolated by centrifugation following procedures of Galehouse (1971). A portion was then filtered, rinsed in distilled water, and then suspended and stirred in 0.09 molal CsCl solution for 24 hr, after which it was collected on filter paper, further rinsed with distilled water, and dried. This portion was then dissolved and analyzed for Cs and the major rock-forming oxides, SiO_2 , Al_2O_3 , Fe_2O_3 , MgO, CaO, Na_2O , K_2O , TiO_2 , P_2O_5 , and 22 minor elements by inductively-coupled plasma (ICP) emission spectrometry. Another split was taken for determining the loss on ignition (LOI) at 900°C and another for XRD determination of the clay minerals, following methods outlined in Moore and Reynolds (1989).

RESULTS

Acid Soluble Fraction

Results (Table 1) show that CaCO_3 makes up an average of 12 wt% of the sediments, ranging between 0.8 and 54%. XRD analyses of high carbonate samples in the reconnaissance study (Bischoff *et al.*, 1997a) indicated that calcite is the dominant carbonate mineral, with minor dolomite and aragonite found in a few samples. Minerals characteristic of higher salinity and/or playa conditions, such as gypsum and gaylussite, were sought but not detected.

As shown in the earlier studies (Bischoff *et al.*, 1997a; Menking *et al.*, 1997), abundant CaCO_3 generally signals long residence time of lake water and/or closed lake condi-

TABLE 1

Contents of Carbonate and Acid-Leachable Ca, Mg, and Sr and Components Derived Therefrom For Sediments from OL-92

	Total sediment (<i>n</i> = 108)	Spilling-lake sediments (<i>n</i> = 28)	Closed-lake sediments (<i>n</i> = 28)
mmoles/100 g sediment			
CO ₃	132.0	20.2	255.0
Ca	119.0	21.0	203.4
Mg	73.3	45.2	137.4
Sr	0.32	0.06	0.57
wt% in sediment			
CaCO ₃	11.9	2.1	20.3
MgCO ₃	1.1	<0.1	4.3
soluble Mg-silicate	8.0	6.1	11.5
SrCO ₃	0.03	0.009	0.05
mol % of carbonate			
CaCO ₃	89.0	99.7	79.5
MgCO ₃	9.8	<0.1	20.2
SrCO ₃	0.3	0.3	0.3

Note. CO₃ data are from total inorganic carbon analyses (TIC) reported in Menking *et al.* (1997). Ca, Mg, and Sr are from 3*N* HCl leach of sediment.

tions while times of low CaCO₃ abundance represent high water-flow and spilling conditions. Chemical mass balance calculations based on the lake area at the spill elevation and the prediversion Ca flux from the Owens River suggested that sediments should contain an average of about 11 dry wt% CaCO₃ (or 7% as CO₃) at the limit of a steady-state full, but not spilling, lake (Bischoff *et al.*, 1997a).

Thus, for the sake of comparison and contrast of the present data set, we selected a subgroup of 28 samples with high CO₃ abundance, arbitrarily taken as $\geq 11\%$ CO₃ (average 15%), to represent closed-lake conditions, and another 28 samples of low CaCO₃ abundance ($\leq 2.5\%$ CO₃, average 1.2%) to represent strongly spilling conditions, (Fig. 2). Although the division of the two groups is arbitrarily based on CaCO₃ content, the differences are paralleled by other chemical and mineralogical proxies outlined below.

The plot of Ca versus CO₃ (Fig. 2) shows that for Ca contents up to 6%, all the CO₃ is essentially pure CaCO₃. At higher concentrations of Ca, however, there is an increasing excess of CO₃, noted also in the earlier study (Bischoff *et al.*, 1997a) which is attributed to a MgCO₃ component. In calcite-dominated systems, progressive salinity increase produces a sequence from low-Mg to high-Mg calcite (Müller *et al.*, 1972). The acid-leachable Mg, however, exceeds this excess CO₃, increasingly so in high-CaCO₃ samples, and this "excess" Mg is attributed to a Mg-silicate phase. We postulate that the bulk of this phase is authigenic Mg-hydroxysilicate, varying crystalline and amorphous (Jones, 1986). Such acid-soluble authigenic phases form in saline lakes by reaction of dissolved Mg and silica in alkaline

solution, both by homogeneous precipitation and/or surface precipitation on preexisting clastic phyllosilicates that were either suspended in the water column or at the sediment-water interface.

The data allow a semiquantitative estimate of the phase makeup of the acid soluble component (Table 1) as follows. Although the Mg:Si ratio of the authigenic silicate may be somewhat variable, for semiquantitative estimates we use the stoichiometry of dehydrated sepiolite (i.e., Mg₄Si₆O₁₅(OH)₂). Sepiolite tends to form in preference to other Mg-hydroxysilicates in most high-silica natural solutions at surface conditions (Garrels and Mackenzie, 1967). Using this methodology we see that the total sediment has about 12% CaCO₃, 1% MgCO₃, and 8% soluble Mg-silicate. Spilling-lake sediments contain 2% CaCO₃, no MgCO₃, and about 6% Mg-silicate, while closed-lake sediments contain 20% CaCO₃, 4% MgCO₃, and about 12% Mg-silicate. On a mole-% basis CaCO₃ makes up 99.7% of the carbonate fraction in the spilling-lake samples, and CaCO₃ makes up about 80% and MgCO₃ about 20% of the closed-lake samples. SrCO₃ makes up 0.3 mole% of the carbonate component in both the spilling and closed suites (Table 1). Such constancy of the SrCO₃ content suggests that the detrital carbonates are an insignificant fraction of the total carbonate, even in the spilling-lake sediments where carbonate abundance is low. It is unlikely that the bedrock carbonates in the Owens drainage basin would have the same Sr/Ca ratio as the authigenic carbonates formed in Owens Lake. Analysis of the seven modern Owens

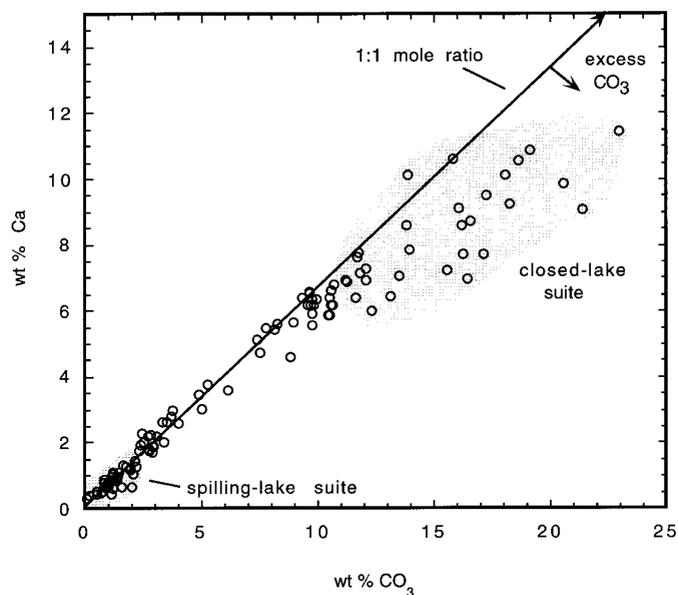


FIG. 2. Plot of acid-soluble Ca against total bulk CO₃ (TIC from Menking *et al.*, 1997) for samples of present study. For Ca $\leq 6\%$, all the CO₃ is as CaCO₃. For Ca $> 6\%$ excess CO₃ is due to a MgCO₃ component. Shaded areas represent subsets of samples taken to represent open and closed-lake conditions.

TABLE 2

Semiquantitative Estimate of Average Composition of Clay Size Fraction of Spilling and Closed-Lake Suites of Samples from OL-92 and of Modern Owens River Sediments, Compared with Modal Mineralogical Composition of Eastern Sierra Nevada Plutons

Mineral	Composite sediment	Spilling-lake sediments	Closed-lake sediments	Modern Owens River	Plutonic composite
Smectite	29	13	52	37	—
Biotite	25	25	19	31	6
Vermiculite	4	4	6	<1	—
Kaolinite	3	3	1	<1	—
Plagioclase	17 (An ₃₆)	25 (An ₃₅)	11 (An ₄₂)	15 (An ₃₃)	40 (An ₃₅)
Orthoclase	11	15	6	8	25
Quartz	11	14	6	8	25

Note. Mineral abundances given as weight percent of air-dried sample. For plutonic composite, area-weighted modal analysis of 18 plutons taken from Bateman (1961); for sediments, plagioclase content and An composition calculated from Na₂O and CaO contents given in Tables 3. Orthoclase and quartz ratios to plagioclase were assumed the same as in the plutonic composite given above. The remainder from 100% is assigned to smectite, biotite, and vermiculite + kaolinite, the relative percentages of which were determined by peak area calculations using multiplicative factors given by Hallberg *et al.* (1978) as modified by Hay *et al.* (1991).

River sediments revealed CO₃ contents generally below 0.5%. Also, the stable isotope composition of the glacial-stage carbonates indicates they are authigenic (Menking *et al.*, in press).

Clay-Size Fraction

X-ray diffraction. XRD results show that the <2-micrometer fraction of Owens Lake and modern Owens River sediments contain major amounts of smectite, biotite, quartz, and feldspar, and minor amounts of vermiculite and kaolinite. The reconnaissance study noted that the dominant clay minerals were smectite and illite (Menking, 1997), with smectite predominating in the closed-lake intervals and illite in the spilling intervals. Illite and biotite have almost identical XRD characteristics, and the chemical analyses in the present study indicate that what had been called illite on reconnaissance was, in fact, very fine-grained detrital biotite. Semiquantitative relative proportions of smectite, biotite, vermiculite and kaolinite (Table 2) were estimated from the XRD results using peak areas and multiplicative factors given by Hallberg *et al.* (1978). The proportions of plagioclase, orthoclase, and quartz were estimated from the chemical analyses as discussed below. As noted in the reconnaissance study, smectite predominates in the closed-lake sediments and biotite in the spilling lake sediments. The spilling suite contains an average of 52% smectite and 19% biotite, while the closed suite contains 13% smectite and 25% biotite. In contrast the modern Owens River sediments contain

37% smectite and 31% biotite. The reconnaissance study (Bischoff *et al.*, 1997a) had inferred that lake smectite, particularly from the closed-lake sequences, contained both a pedogenic and authigenic component, the relative proportions of which could not be quantified. The clay composition presented here for the modern river sediment represents the modern pedogenic input, and, to a first approximation, is perhaps typical of closed-lake periods. The relative amount of smectite in the river sediments compared to that in the closed-lake lake sediment, by a simple mass-balance calculation, indicates that the closed-lake lake sediments were enriched by an authigenic component equivalent to about 31 wt% of the <2-micrometer fraction, presumably of a Mg-silicate composition as inferred from the acid-soluble fraction. This suggests that the typical closed-lake smectite is about half pedogenic and half authigenic.

Chemical composition. Averaged chemical composition for the carbonate-free clay size fraction of the composite sediment, spilling and closed-lake suites, and modern Owens River sediment are shown along with a plutonic composite for comparison (Table 3). Bateman (1961) presented geologic maps and chemical and modal analyses of 18 plutonic masses which dominate the bedrock of the east-central Sierra Nevada and which characterize the glaciated drainage sources of the Owens River. We calculated an area-weighted average of the chemical and modal analyses of these plutons to represent the bulk detrital load (rock flour) that might have been supplied to the Owens River during glacial times (Table 3). The plutonic composite has the composition of a granodiorite with a modal composition of 40% plagioclase (An₃₄), 25% orthoclase, 25% quartz, and 6% biotite (Table 2). The clay fraction of both spilling and closed-lake sediments from Owens Lake differs from the plutonic composite by having generally elevated Fe, Mg, and ignition loss, and depleted Na, K, and Ca, indicative of the varying proportions of pedogenic and authigenic clay minerals and the mechanical enrichment of plutonic biotite in the sediments.

A triangular diagram of MgO-Fe₂O₃-TiO₂ (Fig. 3) shows that the spilling-lake clays plot close to the field of Sierran plutonic rocks which in turn is within the field of biotite compositions from the literature. This relationship shows that the bulk of Mg, Fe, and Ti content in both Owens Lake spilling-lake clays and in Sierran plutonic rocks is from biotite. River clays define a small field displaced slightly toward saponitic smectites from the biotite field showing the addition of pedogenic smectites. The closed-lake clays, in contrast, define a long linear trend away from both the biotite and river clay fields toward the MgO apex. Thus, the closed-lake sediment (Table 3) has a significant Mg enrichment (MgO/Al₂O₃ = 0.64 compared to 0.29 for river clays) and water content (12.97% LOI compared to 10.3% for river clays) indicative of a hydrous Mg-rich authigenic phase. We tested this idea by mathematically adding a sepiolite

TABLE 3
Averaged Chemical Composition of Clay-Size Fraction of Sediments from OL-92, Clay-Size Fraction of Modern Owens River Sediments and Composite Eastern Sierra Plutonic Rocks

Percent	Composite sediment	Spilling-lake sediment	Closed-lake sediment	Modern Owens River	Plutonic composite
SiO ₂	56.58	54.25	55.31	57.80	67.76
Al ₂ O ₃	12.97	16.6	10.69	13.45	15.07
Fe ₂ O ₃ (t)	7.62	9.07	6.82	9.53	3.91
MgO	5.22	4.11	7.82	3.86	1.41
CaO	1.24	1.78	0.9	1.01	3.28
Na ₂ O	1.27	1.97	0.72	1.19	3.27
K ₂ O	2.7	3.75	2.08	2.61	4.1
TiO ₂	0.64	0.88	0.49	0.54	0.46
P ₂ O ₅	0.33	0.29	0.31	0.30	0.16
Na ₂ O _{ce}	1.3	1.04	1.78	1.16	—
LOI	10.0	6.25	12.97	10.3	0.29
Sum	99.90	99.99	99.89	101.35	99.71
ppm					
Mn	855	1172	687		
As	57	31	61		
Ba	638	901	434	924	
Be	2	2	2		
Ce	46	47	46		
Co	21	26	18		
Cr	47	46	50		
Cu	91	79	99		
Ga	20	27	15		
La	29	32	28		
Li	361	141	682		
Nb	16	20	14		
Nd	19	19	19		
Ni	35	29	51		
Pb	31	32	31		
Sc	11	13	9	12	
Sn	136	75	209		
Sr	166	202	137	134	
Th	27	33	22		
V	161	168	166		
Y	11	12	11	23	
Zn	165	218	138		

Note. LOI, loss on ignition at 900°C; Na₂O_{ce}, Exchangeable cation capacity calculated as Na₂O weight-equivalent of analyzed Cs₂O. For plutonic composite, analyses of 18 plutons given by Bateman (1961) were averaged after weighing each by relative outcrop area.

component to the river clay composite and found that the composition of the closed-lake clay composite can be reproduced. Taking the composition Mg₄Si₆O₁₅(OH)₂·6H₂O, we calculate the amount of sepiolite needed to add to the river sediment composite to adjust the MgO/Al₂O₃ ratio (0.29) to equal that of the closed-lake sediment (0.64). We then add the stoichiometric-equivalent amounts of SiO₂ and H₂O and then normalize the mixture to 100% to calculate the new composition (Table 4). The calculation shows that the addition of 23.6 g of sepiolite to 100 g of average river clay makes the mixture have a MgO/Al₂O₃ ratio equal to that of the closed-lake sediment. The resulting normalized bulk-composition is remarkably similar to the composition of the closed-lake clay composite for every one of the major oxides

and LOI (Table 4). Thus, by this analysis the closed-lake clay has about 21% of a sepiolite-like authigenic phase. This compares with the estimate of 31% based on the mass balance calculation from the semiquantitative XRD results. These estimates are minima, because an unknown portion of the authigenic component might have dissolved during the treatment to remove carbonate.

A plot of Na₂O versus CaO (Fig. 4) of the river and lake clays shows a single linear trend which coincides remarkably well with the dilution trend of the Sierran plutonic composite. The spilling-lake clays have the highest concentrations of Na₂O and CaO while the closed-lake clays have the lowest. River clays plot with the closed-lake clays. The trend and slope coincidence show unequivocally that the bulk of

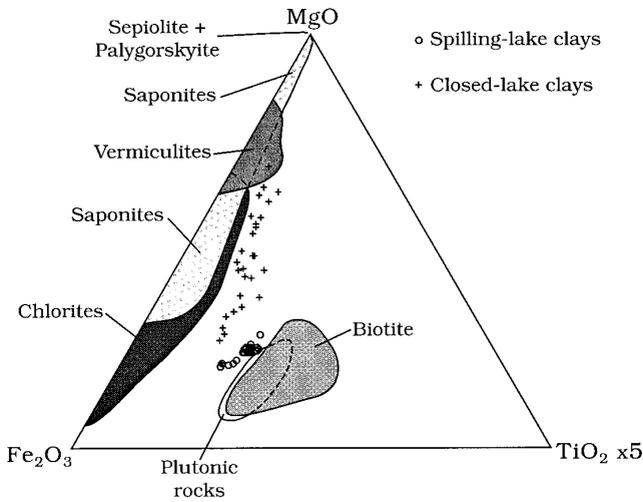


FIG. 3. Diagram showing MgO-Fe₂O₃-TiO₂ relations for clay-fraction of the present study compared with various clay minerals (Weaver and Pollard, 1975), biotite (Deer, Howie, and Zussman, 1962), and Sierran plutonic rocks (Bateman, 1961). Spilling-lake clays plot close to the field of Sierran plutonic rocks which in turn plots within the field of biotite, showing that the bulk of Mg, Fe, and Ti in both Owens Lake spilling-lake clays and in Sierran plutonic rocks is from biotite. River clays define a small field displaced slightly toward saponitic smectites from the biotite field showing the addition of pedogenic smectites. The closed-lake clays trend toward the MgO apex indicating an authigenic Mg enrichment.

both the nonexchangeable CaO and Na₂O of the river and lake clay-size fractions are due to detrital plutonic plagioclase of andesine composition (Ab₃₅). The grouping of spilling and closed-lake clays on this diagram shows the supply of this plagioclase is greater during spilling-lake conditions as defined by low carbonate abundance. Therefore, we conclude that the plagioclase represented by the Na₂O and CaO in the <2-micrometer fraction is a rock-flour component whose abundance reflects the variation in glacial activity, and that glacial activity generally coincides with times of spilling-lake conditions.

DISCUSSION

Distribution of Components

The chemical and XRD data indicate that the lake and river clays are mixtures primarily of biotite, authigenic, and pedogenic smectite and detrital rock-flour whose amounts can be estimated. Based on Na₂O and CaO we calculate that plagioclase accounts for 25% of the spilling-lake composite, 11% of the closed-lake composite and 15% of the river clays (Table 2). This contrasts with 40% plagioclase for the plutonic composite. The plutonic composite also has 25% quartz and 25% K-feldspar, but only about 6% modal biotite, showing the lake and river sediments are greatly enriched in biotite over plagioclase compared to the plutonic source.

TABLE 4
Results of Theoretical Addition of a Sepiolite Component (Mg₃Si₆O₁₅(OH)₂·6H₂O) to Composite of Modern Owens River Clay to Adjust MgO/Al₂O₃ to Equal That of the Closed-lake Composite

Percent	Average modern Owens River clay	Grams sepiolite per 100 g river clay	Theoretical mixture (normalized)	Closed-lake clay composite
SiO ₂	57.80	13.10	57.11	55.31
Al ₂ O ₃	13.45		10.84	10.69
Fe ₂ O ₃ (t)	9.53		7.66	6.82
MgO	3.86	5.96	7.92	7.82
CaO	1.01		0.80	0.9
Na ₂ O	1.19		0.96	0.72
K ₂ O	2.61		2.10	2.08
TiO ₂	0.54		0.43	0.49
P ₂ O ₅	0.30		0.24	0.31
LOI	10.3	4.52	11.90	12.97
Sum	100.19	23.58	100.00	98.11

Note. Resulting theoretical mixture was normalized to 100% to allow comparison with complete composition of the closed-lake composite. Results show remarkable agreement between theoretical mixture and closed lake composite for all components.

Figure 3 suggests that TiO₂ should be a proxy for biotite. A plot of Na₂O versus TiO₂ (Fig. 5) for lake clays shows a linear relationship significantly enriched in TiO₂ compared

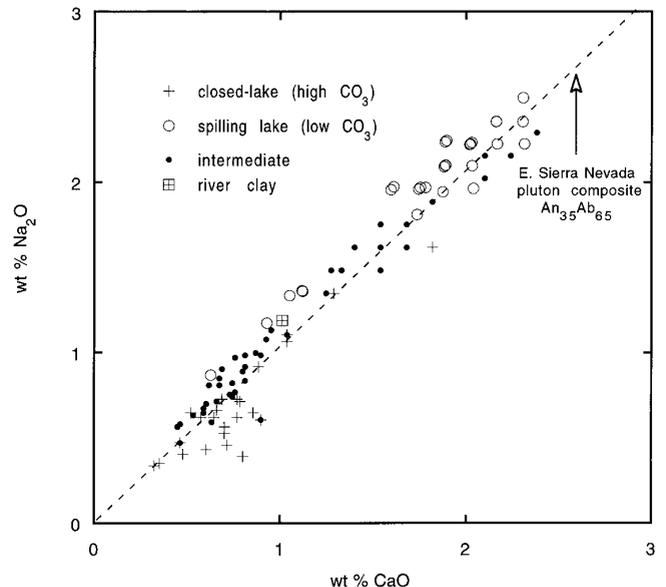


FIG. 4. Plot of Na₂O versus CaO for the clay fraction of present study. The linear trend coincides with the Sierran plutonic dilution trend, showing that the bulk of both CaO and Na₂O is detrital plutonic-plagioclase of andesine composition (Ab₃₅).

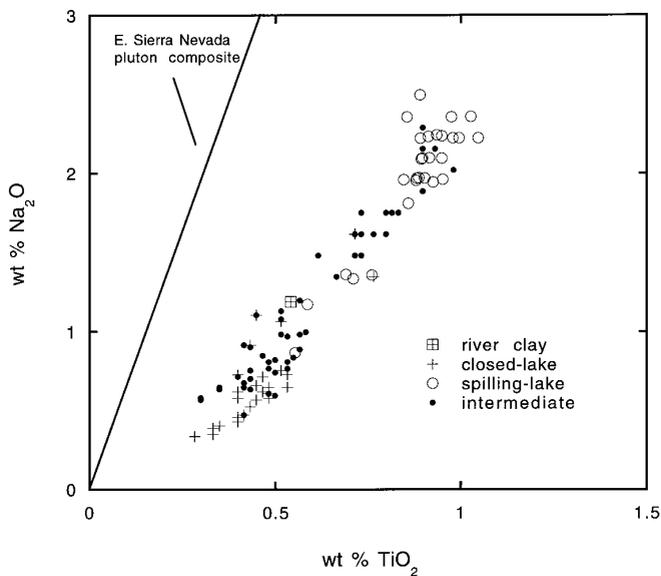


FIG. 5. Plot of Na_2O versus TiO_2 for the clay fraction of present study, showing a linear relationship. The relative enrichment of TiO_2 , the biotite proxy, over Na_2O , the plagioclase proxy, compared to the plutonic composite is probably a consequence of biotite's slower settling velocity and its enhanced transport to the deeper parts of the lake.

to the plutonic composite, indicating that biotite is enriched over plagioclase in the rock flour. This enrichment would be expected because the core site is located near the depositor of the lake, and finely ground biotite is more buoyant and is transported out to the deeper parts of the lake preferentially to plagioclase. By this reasoning, the quartz and K-feldspar of the plutonic rock-flour would presumably have transport characteristics similar to plagioclase. We assume that the river and lake clays contain ratios of quartz and K-feldspar to plagioclase essentially equal to that in the plutonic composite, and we calculate their abundances accordingly (Table 2). Plagioclase, K-feldspar, and quartz, therefore, account for about 54% of the spilling-lake clay, 23% of the closed-lake clay, and 31% of the river clay. Assuming the remainder is made up of the clay suite of primarily smectite and biotite, with minor contributions of vermiculite and kaolinite, we calculate the actual abundances from the relative proportions of these minerals which were determined from the XRD results (Table 2).

Proxies for Climate

Interglacial conditions. Closed-lake conditions result from low effective-precipitation in the drainage basin and are indicated by the abundance of carbonate and smectite (both pedogenic and authigenic). These components are quantitatively represented by the content of Ca, Sr, and Mg in the acid-soluble suite, and by Cs_2O , LOI, and smectite abundance in the clay-size fraction. Along with Li and excess MgO, these

components are seen to covary rather faithfully on time plots (Figs. 6 and 7) and show particular detail in the time interval from 55,000 to 120,000 yr, apparently corresponding to the warm MIS 5. Sr covaries exactly with CaCO_3 confirming that Sr resides in the CaCO_3 phase (Fig. 6). Cs_2O is a measure of ion-exchange capacity and, therefore, is related to the abundance of smectite, as is LOI. Excess MgO (Fig. 7) is defined as $\% \text{MgO} - 0.29 (\% \text{Al}_2\text{O}_3)$ where 0.29 is the $\text{MgO}/\text{Al}_2\text{O}_3$ ratio for river clay. Thus, excess MgO of the clay fraction is a measure of authigenic smectite, as is the soluble Mg-silicate component of the acid soluble fraction (Fig. 6). That Li covaries exactly with excess MgO (Fig. 7) indicates that Li is fixed within the authigenic smectite phase.

Glacial conditions. We take the abundance of rock flour as a proxy for Sierran glacial advances and infer that rock flour is proportional to the abundance of plagioclase and biotite in the carbonate-free clay-fraction. Because the abundance of these minerals is estimated from the carbonate-free clay fraction, it is not affected by CaCO_3 dilution. These phases are quantitatively represented in the carbonate-free

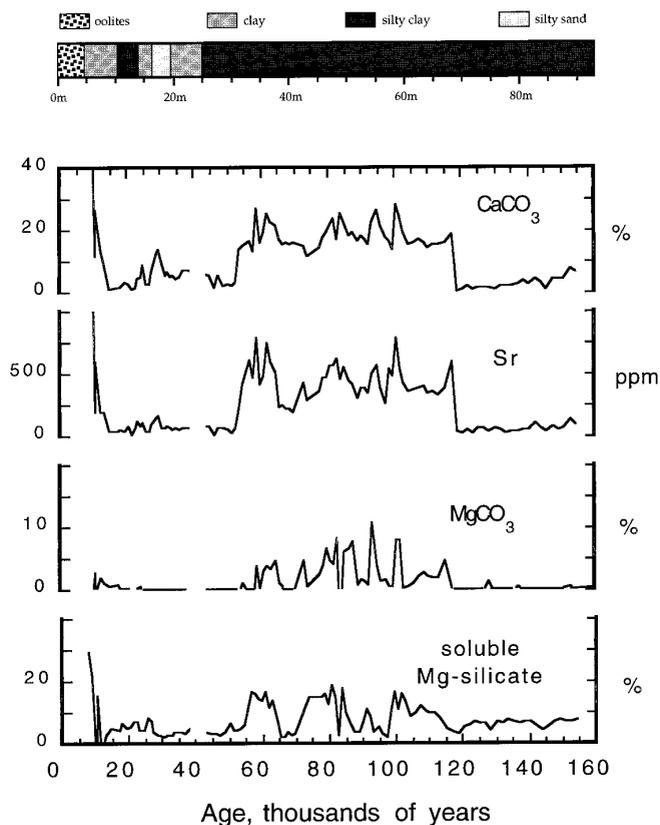


FIG. 6. Time plot of acid-soluble components. Data gap from 39,000 to 44,000 yr B.P. is due to lack of core recovery in this interval. CaCO_3 , MgCO_3 , and soluble Mg-silicate covary and show particular abundance and fine-scale detail from 55,000 to 120,000 yr. Sr covaries exactly with CaCO_3 showing that Sr resides in the CaCO_3 phase.

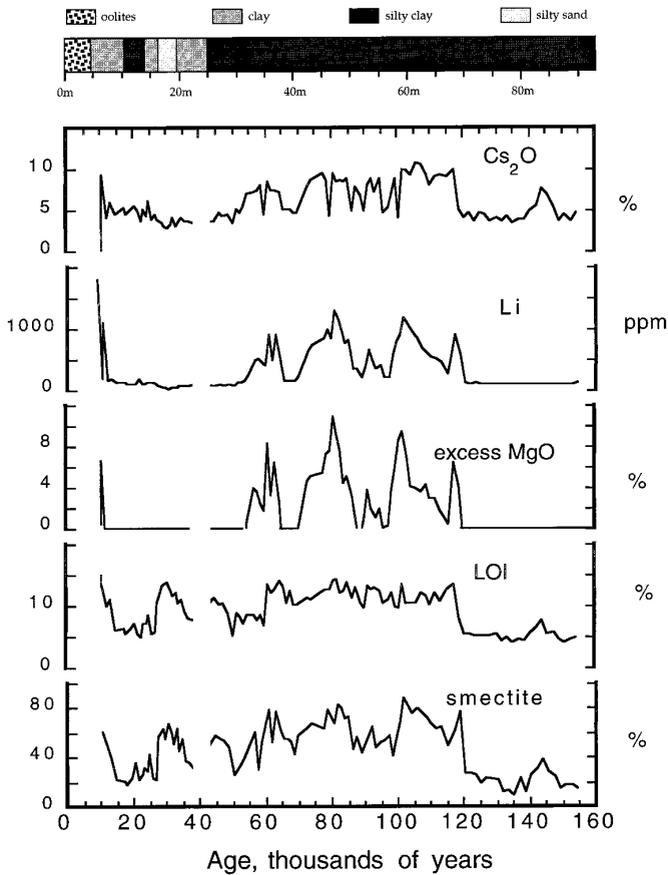


FIG. 7. Time plot of clay-fraction components sensitive to closed-lake or interglacial conditions. Cs_2O , Li, excess MgO, LOI, and smectite covary and show particular abundance and fine-scale detail from 55,000 to 120,000 yr. Cs_2O is a measure of ion-exchange capacity and, therefore, is related to the abundance of smectite, as is LOI. Excess MgO is a measure of authigenic smectite. These components covary in harmony with the acid-soluble components of Fig. 6.

clay fraction by Na_2O , TiO_2 , and biotite abundance by XRD. These components, along with Mn and Ba, are seen to covary rather closely on time plots (Fig. 8) showing particular detail from 10,000–50,000 yr corresponding to cold MIS 2, 3, and 4, and from 120,000–155,000 yr corresponding to cold MIS 6. Ba is a common minor constituent in plagioclase substituting for Ca, and divalent Mn is a common minor constituent in biotite (Deer *et al.*, 1962), substituting for Mg and divalent Fe. The close similarity of the Mn and Ba time plots (Fig. 8) shows that each faithfully represents the abundance of their mineral host, and that the relative proportion of plagioclase and biotite remains approximately constant through the various climatic cycles.

Climatic Events and Comparison to Other Records

We compile the results of the present study with our previous study (Menking *et al.*, 1997) and compare them with

other paleoclimate records (Fig. 9), namely the high-resolution study of nearby short cores (Benson *et al.*, 1996), the marine isotope record, and the Greenland ice core record (GRIP) and the nearby Devils Hole record. The record of Sierran glacial advances is summarized in the Na_2O content, and the record of the cycling between open- and closed-lake conditions by the content of CaCO_3 , authigenic Mg-silicate, and the $\delta^{18}\text{O}$ content of the carbonate.

The time-depth scale for OL-92 (Bischoff *et al.*, 1997b) is based on radiocarbon control for the uppermost 24 m (to 30,000 yr), the Bishop ash bed at 304 m (at 759,000 yr) and the inference of a constant mass-accumulation rate of 51.4 $\text{g}/\text{cm}^2/1000$ yr. The average mass-accumulation rate calculated for the radiocarbon-dated segment, using bulk densities derived from pore-water measurements, was found to be essentially the same as for the entire core down to the Bishop Ash, suggesting that an approximately constant accumulation rate prevailed throughout the past 759,000 yr. The resulting time-depth curve is remarkably coincident with a similar plot independently derived from ten within-Brunhes

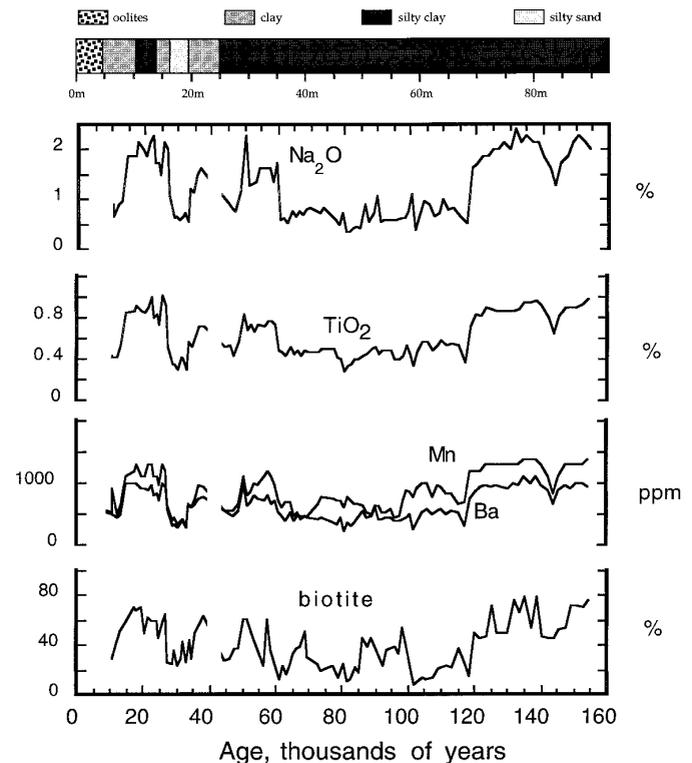


FIG. 8. Time plot of clay-size fraction components sensitive to glacial advances. Na_2O , TiO_2 , and biotite covary in harmony, along with Mn and Ba, showing particular detail from 10,000 to 50,000 yr and from 120,000 to 160,000 yr. Ba is a minor constituent of plagioclase and Mn is a minor constituent in biotite. The close similarity of the Mn and Ba patterns shows that each follows the abundance of their mineral host, and that the relative proportion of plagioclase and biotite remains approximately constant through the various climatic cycles.

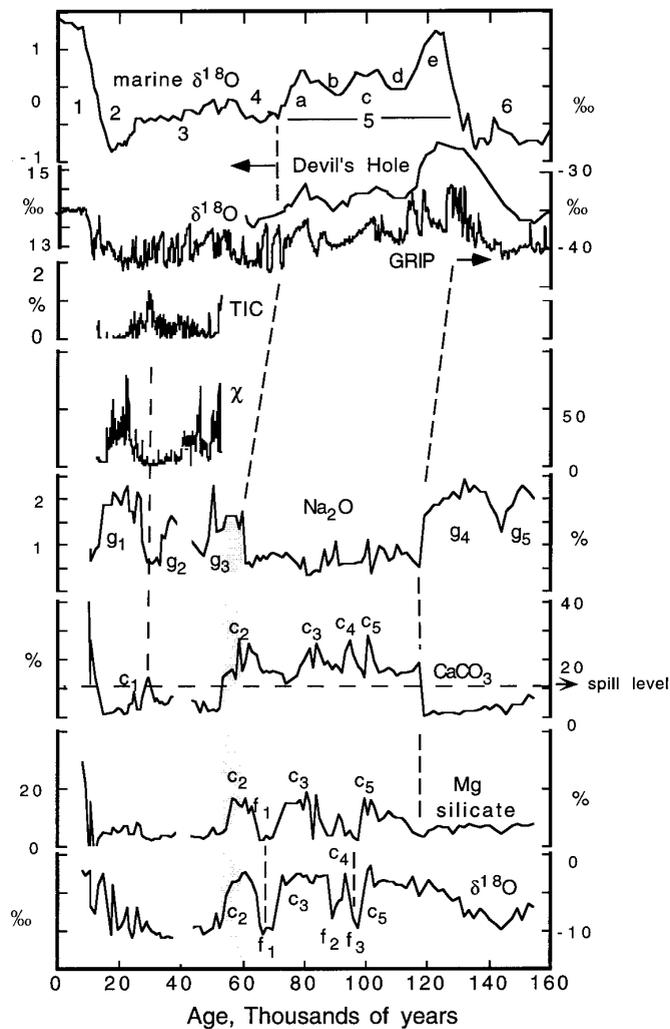


FIG. 9. Time plots of Na_2O , CaCO_3 , and soluble Mg-silicate from present study, and $\delta^{18}\text{O}$ of the companion study (Menking *et al.*, 1997), compared to high-resolution study of nearby Owens Lake short cores (TIC and magnetic susceptibility, χ , of Benson *et al.*, 1996), the marine-isotope record of Martinson *et al.* (1987), the isotope record of the Greenland GRIP ice core (Dansgaard *et al.*, 1993), and the isotope record of Devils Hole, Nevada (Winograd *et al.*, 1992).

paleomagnetic events identified by Glen and Coe (1997). For the purposes of the present study the model is reasonably robust back to about 50,000 yr, but becomes less certain beyond. The position of Termination II (MIS 5/6 boundary) based on the abrupt change in TIC of the reconnaissance study, for example, appears at about 120,000 yr (Bischoff *et al.*, 1997a) instead of about 128,000 yr predicted from orbital tuning (Martinson *et al.*, 1987), and about 140,000 yr B.P. as seen in the Devils Hole record (Winograd *et al.*, 1992). Thus, the time–depth model may be as much as 8000 yr in error at around 100,000 yr, and is subject to modification as future progress is made on absolute dating of the core.

The data indicate that full interglacial conditions in the Owens Basin had been attained by 118,000 yr on our time scale, as shown by an abrupt increase in CaCO_3 and an abrupt decrease in Na_2O (Fig. 9). Similarly, we interpret the end of this interglaciation to occur at about 53,000 yr in OL-92 as shown by abrupt decreases in CaCO_3 and Mg-silicate and the onset of high contents of Na_2O . Na_2O content remains low throughout the last interglacial and CaCO_3 remains high. The boundary at 118,000 yr is represented by a clear lithologic contact at which laminated reddish-brown muds abruptly and conformably overlie massive, burrowed gray muds. The boundary is also manifest in abrupt changes in several other parameters measured in the reconnaissance study, including smectite abundance (Menking, 1997); pollen, in which *Pinus* replaces cedars and junipers (Litwin *et al.*, 1997); diatoms, in which saline forms replace freshwater forms (Bradbury, 1997); and ostracodes, in which the saline forms replace freshwater forms (Carter, 1997). In contrast, the boundary appears to be gradual in the $\delta^{18}\text{O}$ record reported in our companion study (Menking *et al.*, 1997) in which the change from light to heavy values begins at about 140,000 yr and gradually increases to a maximum by 118,000 yr (Fig. 9). The $\delta^{18}\text{O}$ record, therefore, is more consistent with the Devils Hole chronology, both in timing and in the gradual increase of the proxy. The exact position of the termination, therefore, may be proxy-dependent.

In the marine record, the MIS 5 warm period extends from 130,000 to 73,000 yr B.P. (57,000 yr duration) where it is bounded by an abrupt cooling event comprising the 4/5 stage boundary (Martinson *et al.*, 1987). As reviewed by Muhs (1992), there is debate about the meaning of the “last interglacial” in various marine and continental settings, whether it was short (10,000 yr) and corresponds to stage 5e, or whether it was long (60,000 yr) and corresponds to all of stage 5. Interglacial conditions at Owens Lake appear to span the entire time from 118,000 to 53,000 yr (65,000 yr). Moreover, the Owens record seems to lack the prominent peak of stage 5e characteristic of the marine record (Fig. 9) and seems to suggest that the “last interglacial” was long in the Western Great Basin. The offset between this span and MIS 5 reflects either an error in our age–depth model, or alternatively, a real time-lag between changes in Northern Hemisphere ice-volumes and the manifestation of local climate change in lake geochemistry and sedimentology.

The CaCO_3 record shows five maxima (c_1 – c_5) reflecting periods of peak effective dryness (Fig. 9). The first, c_1 , a doublet peak at 25,000 to 33,000 yr (reaching 12 wt% CaCO_3), occurs within MIS 3, an interstadial within the last glacial cycle. c_1 is mirrored by a distinct minimum in the Na_2O record and shows clearly on the LOI and smectite records (Fig. 7) but has no counterpart in the Mg-silicate or the $\delta^{18}\text{O}$ record (Fig. 9). It also is seen in the same time interval in the TIC record (Benson *et al.*, 1996) which shows

a peak value of 1.38 wt% C. Further, the absolute concentration of carbonate in the corresponding peak in the two records is equivalent, i.e., 1.38 wt% C is equivalent to 11.5 wt% CaCO₃, which compares to 12 wt% observed in the present study. These results demonstrate that the chronology and the carbonate proxies of the cores used in the two studies (Benson *et al.*, 1996; and this study) correspond rather well during this time period.

The four other maxima in the CaCO₃ record occur within MIS 5. Three peaks, c₂ (53,000–65,000 yr B.P.), c₃ (77,000–86,000 yr B.P.), and c₅ (100,000–103,000 yr B.P.) have corresponding peaks in the Mg silicate and δ¹⁸O records (Fig. 9) whereas c₄ (92,000–96,000 yr) has a corresponding peak in the δ¹⁸O record, but not in the Mg-silicate record. In the reconnaissance study (Bischoff *et al.*, 1997a) the limiting sediment-content of CaCO₃ that separates closed and overflowing was calculated to be about 11 wt%, based on a mass balance between the Ca flux from the modern Owens River, the mass accumulation rate of the lake sediments, and the area of the Owens Lake basin at spill level. The CaCO₃ content of all samples from the last interglacial in the present study exceeds this limit, suggesting closed conditions throughout that time. Continuous closure over 65,000 yr is unlikely, however, because it would create extreme salinity and a suite of evaporite minerals, none of which, nor their pseudomorphs, have been observed (Bischoff *et al.*, 1997a). Also, data from downstream Searles Lake indicates that it was receiving spill water, at least periodically, during this same interval (Smith, 1983, 1991; Bischoff *et al.*, 1985). In addition, the diatom record suggests there were several short-lived spilling events between 80,000 and 90,000 yr (Bradbury, 1997). Thus, short spilling events occurred which are not recorded in the resolution (1500 yr/sample) of the present study. These seem to be shown, however, in both the δ¹⁸O record and Mg-silicate records (Fig. 9) which appear remarkably correlated. Three spilling events (f₁, f₂, and f₃) appear to be reflected in both records, f₁ at 65,000–71,000 yr B.P., f₂ at 85,000–90,000 yr B.P. and f₃ at 92,000–97,000 yr B.P., consistent with the diatom and ostracode records. f₁ perhaps corresponds to MIS 5b and f₂ and/or f₃ to MIS 5d, both of which are clearly seen in the Devils Hole record and seem to be reflected in the GRIP record as well (Fig. 9).

We infer that glacial advances are indicated by peaks in the Na₂O record (Fig. 9). Two major advances (g₄ and g₅) are seen during the penultimate glacial cycle with broad peaks from 120,000–140,000 yr B.P. (g₄) and from 146,000 to 155,000 yr B.P. (g₅). Na₂O values are uniformly low during the last interglacial, but three major peaks are seen during the following last glacial cycle (g₁, g₂, and g₃). Among these, the glacial advance at 48,000–60,000 yr B.P. (g₃) signals the end of the long interglacial and presumably correlates with the MIS 4 boundary. This advance is unique in that its early part seems to have a 6000-yr overlap with

closed lake conditions of c₂ (53,000–65,000 yr B.P.) shown in the CaCO₃, Mg-silicate, and δ¹⁸O records (see shaded zone in Fig. 9). The remaining advances are centered around 31,000–37,000 yr B.P. (g₂) and 14,000–27,000 yr B.P. (g₁), and the gap between them, which coincides with the interstadial dry period c₁, is marked by low, interglacial levels of Na₂O. The age spans of g₁, g₂, and g₃ correspond rather well to ³⁶Cl dates reported for Sierra Nevada moraines by Phillips *et al.* (1996). They obtained a range of 40,000–50,000 yr B.P. for younger Tahoe moraines (corresponding to our g₃); 30,000 to 32,000 yr B.P. for Tioga 1 moraines (corresponding to our g₂), and 10,000–25,000 yr B.P. for Tioga 2 and 3 (corresponding to our g₁).

The magnetic-susceptibility record of Benson *et al.* (1966) shows peaks which correspond to the younger Tahoe (our g₃) and to Tioga 1 and 2 (our g₁), but show no peak from 30,000 to 40,000 yr B.P. corresponding to Tioga 1 (our g₂), even though their δ¹⁸O values for this span indicate cold conditions. Although core OL-92 has some disturbed zones within the uppermost 35 m (last 50,000 yr, caused by the slumping of younger material during withdrawal of the drill-string), we noted these zones and were careful to avoid them in our sampling. The section corresponding to g₃ lacked any disturbed zones. Benson *et al.* (1966) concluded that the Tioga 1 advance must have been relatively minor and, therefore, that it did not show up in the lake record. Our results indicate the Tioga 1 advance is, indeed, recorded in the Owens Lake record by our rock-flour proxies. The magnetic susceptibility record apparently reflects primarily the abundance of diagenetic greigite (Fe₃S₄). Benson *et al.* (1996) inferred that greigite was controlled by the abundance of detrital iron-bearing minerals, and thus to glacial erosion. Greigite is only one mineral component of the common authigenic black iron-sulfides which form in organic-rich marine and lacustrine sediments. The greigite component is not necessarily constant, and besides being limited by the detrital source of iron, its abundance is limited by the availability of dissolved sulfate and the concentration of metabolizable organic matter (Berner, 1970). Thus, the number of variables which might affect the abundance of greigite make it less than a unique proxy for rock flour.

SUMMARY AND CONCLUSIONS

The foregoing allows the following conclusions:

1. Closed-lake conditions which result from times of low effective-precipitation and river discharge are reflected by the abundance of carbonate and smectite (both pedogenic and authigenic). These components are quantitatively represented by the content of Ca, Sr, and Mg in the acid-soluble suite, and by Cs₂O, LOI, and smectite abundance in the clay-size fraction. Closed-lake conditions generally correspond to interglacial conditions in the Owens Basin.

2. Typical closed-lake sediment contains 20% CaCO₃, 4% MgCO₃, and about 12% Mg-silicates, and the corresponding clay fraction contains 52% pedogenic and authigenic smectite and 19% detrital biotite.

3. Typical open-lake sediment contains 2% CaCO₃, no MgCO₃, and about 6% Mg-silicate, and the corresponding clay fraction contains 13% pedogenic smectite and 25% detrital biotite, but no authigenic smectite.

4. Modern Owens River sediments, presumed to represent typical interglacial detrital input, generally contain less than 0.5% carbonate, and the clay fraction consists of 37% pedogenic smectite and 31% detrital biotite.

5. Rock flour is a proxy for glacial advances, and is proportional to the abundance of plagioclase and biotite in the clay-size fraction, the ratio of which remains essentially constant over the past 155,000 yr. These phases are quantitatively represented by Na₂O, TiO₂, Ba, and Mn in the clay fraction.

6. Rock flour reaches two major peaks, signaling two glacial advances, during the penultimate glacial cycle (MIS 6), no major advances during the last global-interglacial (entire MIS 5), and three major advances during last global-glacial cycle (MIS 2, 3, and 4). The ages of the latter three correspond rather well to dates reported for Sierra Nevada moraines.

7. The onset of the closed-lake conditions signaling the last interglacial is shown by abrupt increases in CaCO₃ and Mg-silicate and an abrupt decrease in rock flour at 118,000 yr B.P. with the present time scale. The δ¹⁸O record, in contrast, shows a gradual increase beginning at about 140,000 yr B.P., in agreement with the Devils Hole record. Conditions of high carbonate and low rock flour prevailed during the entire period until the glacial advance at 60,000 yr signaled the end of this long interglacial. This advance is unique in that its early part occurred during closed-lake conditions.

ACKNOWLEDGMENTS

The manuscript was greatly improved for readability and scientific consistency from careful reviews by Platt Bradbury, Steve Colman, G. I. Smith, Scott Stine, and an anonymous reviewer. To them our sincere thanks are given.

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