Responses of sediment geochemistry to climate change in Owens Lake sediment: An 800-k.y. record of saline/fresh cycles in core OL-92

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ABSTRACT

Geochemical parameters of sediments from drill hole OL-92 indicate that Owens Lake was saline, alkaline, and highly productive during interglacial periods, and was hydrologically open and relatively unproductive during glacial periods. Abundance of CaCO₃, organic carbon, and cation-exchange capacity of the clay fraction show cyclic variation down the core. Six minima in these components during the past 500 k.y. are interpreted as caused by intensive overflow that occurred during Sierran glacial advances. Maxima in these components indicate closed-lake conditions, reflecting warmer and more arid interglacial climates. The pattern of CaCO₃ abundance suggests that closed lake conditions predominated over the past 500 k.y. The absence of gaylussite and gypsum in the sediments, however, indicates lake salinity never exceeded about 15 wt %, a limit which requires flushing of accumulated salts every 10 k.y.

Oscillations of CaCO₃ generally indicate a 100-k.y. dominant cycle, a characteristic of the marine δ¹⁸O record. Four of the last five marine isotope terminations are clearly shown in the Owens Lake record. The last interglacial at Owens Lake appears to have occurred between 120 ka and 50 ka. The roughly 10-k.y. offset between this interval and marine oxygen-isotope stage 5 reflects either error in the age-depth model, or alternatively, a time lag between changes in Northern Hemisphere ice volumes and the manifestation of local climate change in lake geochemistry and sedimentology.

INTRODUCTION

Prior to its desiccation in the decade following 1913, a result of source-water diversion by the City of Los Angeles, Owens Lake was historically below its spill point and was the terminus of the Owens River. As a consequence, the lake was saline and alkaline (Gale, 1914). Saline conditions likely characterized the second half of the arid Holocene and, presumably, the earlier interglacials as well. During earlier wet periods that are thought to be coeval with the periodic Sierran glacial advances, however, down-stream evidence at Searles Lake (Bischoff et al., 1985; Smith et al., 1983; Smith, 1991) shows that Owens Lake had been intensely overflowing, and therefore, must have been flushed with fresh water.

Owens Lake core OL-92 was located near the depocenter of the Owens Lake basin in order to obtain a complete record of the alternating closed and overflowing cycles in a stratigraphic and chronological framework. The postulate of the present study is that the sediment composition should reflect the cycling between relatively wet and dry periods.

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Sediments from Owens Lake drill hole OL-92 were analyzed for major oxides and minor elements, acid-leachable cations, total organic carbon (TOC), carbonate (CO₃), and cation-exchange capacity (CEC). Results show cyclic variation of sediment composition down the core, reflecting alternating global glacial and interglacial conditions that correspond, in general, to the marine isotope cycles. Details of sampling, analytical methods, along with complete data in depth format are given in the data depository of the Owens Lake Drilling Project (Bischoff et al., 1993a). Sediment data are given chronological format in this presentation using the time-depth relations for core OL-92 presented elsewhere in this volume (Bischoff et al., 1993b).

**SAMPLING**

Both channel and point samples were taken from the drill core. Channel samples, which are composite strip samples, were taken in a continuous series to represent the entire recovered sedimentary column without gaps (91 samples). This methodology avoids the bias of single point samples which may not represent the entire sedimentary unit from which they were taken (for point-sample studies of core OL-92 see Bischoff et al., 1993b, and Menking et al., 1993b). The advantage of the channel samples is that each represents a smoothed or running mean of conditions represented by the time span of the sample. The advantages are that no important events are missed (except for gaps in core recovery) and that geochemical budget calculations can be carried out. The time resolution of such samples, however, is inverse to the thickness of the section sampled. In the present study, we took strip samples approximately 3 m long, each deemed to represent about 7,000 years of deposition. Samples were taken from the working half of the core with a U-shaped spatula, resulting in a continuous semi-cylindrical strip about 1.5 cm wide, 1 cm deep, and about 3 m long. The point samples were taken during the drilling operations at every 2 or 3 m (120 samples). These samples were specially preserved and used for determination of water content and pore water chemistry as reported in the accompanying report (Bischoff et al., 1993c; Friedman et al., this volume). These samples were also analyzed for organic carbon and for carbonate content (but not the other components). These results are reported here to supplement similar analyses of the channel samples.

**LABORATORY PROCEDURES**

**Initial processing and splits**

Samples were rinsed in distilled water, dried at 60 °C and mechanically homogenized. One aliquot was split for cesium (Cs) treatment and major oxide analysis, and the remaining sample was used for determination of total organic carbon (TOC) and carbonate (CO₃), x-ray diffraction (XRD) determination of carbonate minerals, and acid-leachable Mg and Ca.

**Cesium-ion treatment and major elements**

An aqueous CsCl solution was used to displace all exchangeable cations in the sample with Cs ions (following Beetem et al., 1962). Thus, the amount of Cs taken up by the sample is a measure of its cation-exchange capacity (CEC). The CEC, 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. A sample split was suspended and stirred in 0.09 molal CsCl solution for 24 hours, after which it was collected on filter paper and rinsed with distilled water. The sample was then dried at 60 °C and analyzed for Cs and the major rock-forming oxides SiO₂, Al₂O₃, Fe₂O₃, MgO, CaO, Na₂O, K₂O, TiO₂, P₂O₅, and MnO by X-ray fluorescence (XRF) spectrometry. A split of each channel sample was also analyzed for minor elements by semiquantitative optical emission spectroscopy.

**Other properties**

Organic carbon and carbonate were analyzed from the bulk sample by standard coulometry which successively measures the carbonate as CO₂ released by strong acid attack and then the total carbon in the sample (Engleman et al., 1985). Splits of bulk samples were leached in 3 molar HCl overnight for analysis of acid-leachable Ca and Mg by standard atomic-absorption spectroscopy. Standard XRD scans of powder mounts were performed on a selection of 36 carbonate-rich samples.

**RESULTS**

**Bulk composition**

Analytical results for individual samples are given in Bischoff et al. (1993a). Contents of CO₃ (7.5% average) and TOC (0.92% average) vary widely as summarized in Table 1. The composition of the acid-insoluble residue, however, is remarkably homogeneous. The average bulk composition, normalized after removing carbonate, organic carbon, and acid-soluble Ca and Mg (Table 2), is similar in all major oxide components to granodiorite, the predominant rock of the Sierra Nevada batholith (Bateman et al., 1963). The minor compo-

**TABLE 1. AVERAGED CHARACTERISTICS OF SEDIMENTS FROM OWENS LAKE OL-92 DRILL HOLE**

<table>
<thead>
<tr>
<th>Component</th>
<th>Average ± 1σ</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carbonate (CO₃)</td>
<td>7.5 ± 6.3%</td>
</tr>
<tr>
<td>Organic carbon (TOC)</td>
<td>0.92 ± 0.87%</td>
</tr>
<tr>
<td>CEC (meq/100g)</td>
<td>32.7 ± 6.8 meq/100g</td>
</tr>
<tr>
<td>Grain density</td>
<td>2.63 ± 0.05 g/cc</td>
</tr>
</tbody>
</table>

*Values are reported as mean ± 1σ. Data averaged from tabulated results presented in Bischoff et al., 1993a.
†Cation exchange capacity for clay-size fraction on a carbonate-free basis.
shows an elongate pattern with respect to the Fe$_2$O$_3$ apex with the Lamarck Granodiorite at the center of the trend. Samples relatively enriched in Fe$_2$O$_3$, tending toward average shale, are clay-rich samples, whereas those in the opposite direction are rich in arkosic sand. The Fe$_2$O$_3$-SiO$_2$-Al$_2$O$_3$ (truncated) triangular plot (Fig. 1C) shows an elongate trend with respect to the SiO$_2$ apex with granodiorite at the midpoint. The SiO$_2$-depleted side trends toward average shale and represents more clay-rich samples, whereas the SiO$_2$-enriched side represent sandy units. The bifurcation of the trend on the SiO$_2$-enriched side distinguishes between quartz sands and arkosic sands.

**Variation of carbonate and organic carbon fractions**

The CO$_3$ and TOC co-vary down core (Fig. 2). It is remarkable that both point samples (representing 40 years of sediment accumulation) and channel samples (representing 7,000 years of sediment accumulation) show almost coincident patterns, indicating that even on the small scale of the point samples the sediments are representative of larger thicknesses of sediment. This observation points to slowly changing (millennial scale) homogeneous depositional conditions. The CO$_3$ and TOC display sharp minima, very close to zero values during the most recent glacial maximum that occurred between 25 ka and 17 ka. These results suggest that the residence time of water in the lake was short, preventing the buildup of dissolved salts sufficient for CaCO$_3$ precipitation and high biologic productivity. Five similar minima in these parameters occur between 25 ka and 500 ka are interpreted as successively older glacial maxima. Conversely, the recurring and broader maxima in CO$_3$ and TOC are interpreted to represent full interglacial conditions during which the lake was the terminus and was saline and biologically highly productive. Before 500 ka (below 230 m), there is a striking change in depositional conditions from silty clays to thick sandy units as described in the accompanying reports (Smith, this volume; Smith et al., this volume; Menking, this volume), which may signal irregular fluctuations from lacustrine to nonlacustrine conditions for this lower part of the core. For the section representing the most recent 500 k.y., however, lacustrine conditions apparently prevailed.

**Composition of the acid-soluble fraction**

Results of XRD indicate that calcite is the dominant carbonate mineral, with detectable dolomite occurring in about a third of the samples, and aragonite in only two (Bischoff et al., 1993a). Minerals characteristic of higher salinity and/or playa conditions, such as gypsum and gaylussite, were sought but not detected. Gaylussite, in particular, would be the first mineral to form (after calcite) at increasing salinity. Even if later leached by low-salinity waters, gaylussite would leave distinctive pseudomorphs composed of CaCO$_3$ (see below, and Bischoff et al., 1991). The lack of gaylussite or its pseudomorphs suggest the lake had not attained the required salinity throughout the past 800 k.y. The rel-

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**Table 2. Averaged Composition of Sediment from Owens Lake OL-92 Drill Hole, on an Acid-Insoluble Basis (Carbonate-Free), Compared to Granodiorites and Average Shale**

<table>
<thead>
<tr>
<th></th>
<th>Owens Lake Sediment</th>
<th>GSP-1</th>
<th>Lamarck Granodiorite</th>
<th>Average Shale</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oxides (wt %)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SiO$_2$</td>
<td>68.60</td>
<td>67.38</td>
<td>66.92</td>
<td>58.1</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>14.77</td>
<td>15.25</td>
<td>15.19</td>
<td>15.4</td>
</tr>
<tr>
<td>Fe$_2$O$_3$</td>
<td>5.05</td>
<td>4.32</td>
<td>5.05</td>
<td>7.52</td>
</tr>
<tr>
<td>CaO</td>
<td>2.92</td>
<td>2.02</td>
<td>3.79</td>
<td>3.11</td>
</tr>
<tr>
<td>MgO</td>
<td>2.41</td>
<td>0.96</td>
<td>1.74</td>
<td>2.44</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>2.67</td>
<td>2.80</td>
<td>3.16</td>
<td>1.3</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>3.53</td>
<td>5.53</td>
<td>3.82</td>
<td>3.24</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>0.62</td>
<td>0.66</td>
<td>0.47</td>
<td>0.65</td>
</tr>
<tr>
<td>P$_2$O$_5$</td>
<td>0.26</td>
<td>0.28</td>
<td>0.18</td>
<td>0.17</td>
</tr>
<tr>
<td>MnO</td>
<td>0.13</td>
<td>0.04</td>
<td>0.08</td>
<td>0.06</td>
</tr>
<tr>
<td>Total</td>
<td>99.96</td>
<td>99.24</td>
<td>100.40</td>
<td>100.00</td>
</tr>
</tbody>
</table>

| Oxides (ppm)    |                 |       |                      |               |
|-----------------|-----------------|-------|                      |               |
| B                | 61              | <3    | 310                  |               |
| Ba               | 980             | 1,300 | 460                  |               |
| Be               | 1.3             | 1.5   | <4                   |               |
| Co               | 14.4            | 6     | 8                    |               |
| Cr               | 31.1            | 13    | 500                  |               |
| Cu               | 26.2            | 33    | 192                  |               |
| Ga               | 31.2            | 22    | 50                   |               |
| Mo               | 4               | 1     |                      |               |
| Ni               | 21.5            | 13    | 24                   |               |
| Pb               | 22.2            | 51    | 20                   |               |
| Sc               | 7               | 7     | 7                    |               |
| V                | 75.5            | 53    | 120                  |               |
| Y                | 14.0            | 30    | 28                   |               |
| Zr               | 56              | 500   | 120                  |               |

*Data averaged from tabulated results reported by Bischoff et al., 1993a. Ignition loss and acid-leachable CaO and MgO were subtracted from bulk sediment composition and the difference normalized to 100%.

GSP-1 is a granodiorite collected near Silver Plume, Colorado (Flanagan, 1976).

Lamarck Granodiorite was collected from east central Sierra Nevada, California (Bateman et al., 1963).

Oxides from Clark, 1924, minor elements from Rankama and Sahama, 1950.
J. L. Bischoff and Others

Figure 1. Triangular diagrams showing major oxide composition of sediments from Owens Lake core OL-92 compared to average shale and granodiorite (GSP-1 and Lamarck Granodiorite, see Table 2). (A) Al₂O₃-Na₂O-K₂O. (B) Fe₂O₃-Na₂O-K₂O. (C) Truncated triangle (lower left corner) of Fe₂O₃-SiO₂-Al₂O₃. Diagrams show the general similarity of Owens Lake sediment to granodiorite, but with small-scale variability attributable to sediment-size fractionation.

Relative amounts of Ca and Mg carbonates, calculated by balancing the analyzed acid-leachable Ca and Mg against analyzed CO₃ (Bischoff et al., 1993a), indicate an average of 95 mole % CaCO₃ and 5 mole % MgCO₃, for the total carbonate fraction. This suggests that 86% of the total acid-leachable Mg is actually noncarbonate (Fig. 3), which we postulate to be authigenic Mg hydroxysilicates, varying crystalline and amorphous, and including such phases as sepiolite, kerolite, and stevensite (Jones, 1986). Such acid-soluble authigenic phases form in saline lakes by reaction of dissolved Mg and silica in alkaline solution, both reacting directly with each other and/or reacting with preexisting clastic phyllosilicates that were either suspended in the water column or at the sediment-water interface. In general, it is not possible to distinguish detrital and authigenic phyllosilicates in mixtures by XRD. Thus, calculating averages from data given by Bischoff et al. (1993a) on a weight basis indicates that 9% of the total bulk Mg is carbonate, 57% is acid-soluble authigenic silicate, and 34% is in the nonleachable clastic component. Figure 3:
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Figure 2. Variation of carbonate (CO$_3$) and total organic carbon (TOC) with depth and age of sediments from Owens Lake drill hole OL-92. Solid lines are from channel samples, dotted lines from point samples. Lithology generalized from core log presented in Smith (this volume). Lacustrine conditions prevailed back to 500 ka (~220 m) below which fluvial and lacustrine conditions may have alternated. Both point and channel samples show almost coincident patterns. Values of CO$_3$ and TOC are very close to zero during the most recent glacial maximum at 17–25 ka suggesting the lake was intensely overflowing during glacial advances.

Figure 3. Variation in relative amounts of CaCO$_3$, MgCO$_3$, and acid-soluble Mg-silicate component with depth and age of sediments from Owens Lake drill hole OL-92. Lithology generalized from core log presented in Smith (this volume). Abundance of both carbonate Mg and authigenic Mg silicate follows that for total carbonate. High contents of these phases are likely indicators of elevated salinity caused by terminal lake conditions. Based on Ca budget calculations (see text) sediments with less than about 0.11 mol/100 g CaCO$_3$ represent overflowing conditions, while those with more CaCO$_3$ represent closed lake conditions.

Figure 4. Co-variation of cation exchange capacity (CEC in meq/100 g) of clay fraction (measured by bulk Cs uptake divided by wt % clay size) and smectite abundance in clay fraction with age and depth in sediments from Owens Lake drill hole OL-92. Sediment-size data and smectite abundance taken from Menking et al. (1993b) and Menking et al. (1993a). Lithology generalized from core log presented in Smith (this volume). The CEC co-varies with smectite abundance, and both reflect interglacial conditions.

Cation-exchange capacity (CEC)

The Cs content of a Cs-exchanged sample is a function of the amount of clay in the sample, and of the CEC of the clay. The CEC of the clay fraction was obtained by normalizing the Cs content (carbonate-free basis) to the weight fraction of the <2μ component of the sample (reported in Menking et al., 1993a). The clay-normalized CEC is seen to co-vary with the relative abundance of smectite in the clay fraction (Fig. 4), an independent confirmation of its validity. The CEC is in phase with the carbonate cycles (Fig. 2) back to 500 ka. As does CO$_3$, CEC indicates a minimum coinciding with the last glacial maximum (25–17 ka), and other minima occur successively deeper in the core coinciding with carbonate minima. This correlation suggests that during glacial maxima the clay-size material has a low exchange capacity, perhaps representing a glacial rock-flour component. Beyond 500 ka (~230 m), the normalized CEC pattern is decoupled from the carbonate pattern. If CEC is a reflection of drainage-basin conditions rather than basin-deposition conditions, then CEC cycles might reflect climatic cycles even though the depositional basin is alternating between lacustrine and nonlacustrine conditions. The average CEC of clay material in the core is 32.7 meq/100 g (Table 1), which compares to a shows that abundance of both carbonate Mg and authigenic Mg silicate follows that for total carbonate, and that both Mg phases are likely indicators of saline and alkaline conditions.
range of 80–150 meq/100 g for pure smectites and to about 10–40 meq/100 g for pure illite (Grim, 1968). Menking et al. (this volume) report that illite and smectite are the dominant clay minerals in the Owens sediment. During the interglacials, CEC reaches values within the pure smectite range, whereas during the glacial times CEC is within the range of pure illite.

**DISCUSSION**

**Evolution of closed lake salinity**

Closed lake conditions give rise to sediments rich in Ca-Mg carbonates, TOC, authigenic Mg silicates, and high CEC smectites, in contrast to overflowing lake conditions where these components approach zero values. Because minerals characteristic of high salinity and/or playa conditions, such as gypsum and gysplussite, were not detected suggests the lake had not attained this salinity at any time during the past 800 ka. If closed conditions prevailed for sufficient time, even with the lake level immediately below the sill, such minerals would eventually form. It is, therefore, possible to establish the maximum time between spilling events using a simple evaporation model for the Owens River to determine the order of formation of minerals more soluble than CaCO$_3$. We consider salinity evolution for two water levels of Owens Lake: at its spill level and at its historic, pre-diversion level. In both cases, we begin with the lake composed of fresh Owens River water, and we assume that the flux of dissolved salts from the Owens River into Owens Lake has been about the same as that of provided by the modern Owens River. Records of river discharge and the dissolved load composition are essentially continuous from the present back to 1906 and include unpublished records of the Los Angeles Department of Water and Power (LADWP) from 1934–1992 and published records of Gale (1914) for 1906–1912, of Wilcox (1946) for 1929–1944, and of Hollett et al. (1991) for 1974–1985. We take the composition of Owens River water as that averaged by Hollett et al. (1991) corrected by us for water imports from the Mono Basin. Records of river discharge and the dissolved load composition are essentially continuous from the present back to 1906 and include unpublished records of the Los Angeles Department of Water and Power (LADWP) from 1934–1992 and published records of Gale (1914) for 1906–1912, of Wilcox (1946) for 1929–1944, and of Hollett et al. (1991) for 1974–1985. We take the composition of Owens River water as that averaged by Hollett et al. (1991) corrected by us for water imports from the Mono Basin as shown in Table 3. The salinity (total dissolved salts, TDS) is 241 mg/L and is characterized by Na and Ca as major cations with HCO$_3$ dominating the anions followed by SO$_4$ and Cl. **Average discharge of the Owens River needed to maintain overflow compares well with measurements made during 1939–1940 and 1969–1970, which averaged 1.27 m/yr decrease in depth (Smith and Street-Perrott, 1983). At this evaporation rate, the increased discharge of the Owens River needed to maintain the lake at its spill point is simply the ratio of surface areas between spill point and historic level, or 2.13 times (see Benson and Pailllet, 1989). During cooler glacial periods, the evaporation rate was probably less, and the river discharge required to maintain overflow, therefore, was somewhat less than 2.13 times historic.**

The initial salt buildup in the lake is simply the flux of salts divided by the lake volume which calculates to be 4.25 mg/L/yr. Thus, we allow an annual increase of 4.25 mg/L/yr dissolved salt of the composition given in Table 3 and calculate the progression of the water composition (Fig. 5) using the computer program PHRPITZ (Plummer et al., 1988). Because of inhibition of nucleation, most CaCO$_3$-producing alkaline lakes seem to be at about ten-fold supersaturation with respect to calcite.
equilibrate with the atmosphere and calcite is allowed to precipitate at ten-fold supersaturation. Thus, from this point the lake water becomes increasingly saline and alkaline (pH 9–10), characterized by linear increases in Na–CO$_3$–Cl–SO$_4$, and with Ca and Mg becoming vanishingly small. Salinity increases continuously for the next 55 k.y. (or 8 k.y. for the 1872 lake level) until it reaches 150,000 mg/L at which point gaylussite (Na$_2$Ca(CO$_3$)$_3$·5H$_2$O) begins to precipitate. No other new minerals become saturated during the next two-fold concentration of the brine to 300,000 mg/L, at which point the simulation was terminated (Fig. 5).

Mono Lake, 187 km to the northwest of Owens, a lake with a similar water chemistry and a similar history to Owens (Phillips, 1877), is presently at saturation with gaylussite, which is forming abundantly in the lake (Bischoff et al., 1991). Gaylussite began forming after the salinity exceeded about 100,000 mg/L in the 1970s as a consequence of the diversion of source waters by LADWP begun in 1941. Thus, gaylussite would be expected to form in Owens Lake not far beyond its point of saturation, and its (or its pseudomorph’s) absence in the Owens sediments, in particular, or other soluble minerals in general, implies that Owens Lake could not have been at its historic level for any time during the past 800 k.y. for longer than about 8 k.y., or closed at its spill level for longer than 55 k.a.

The salinity of the pre-diversion Owens Lake was about 6–8 wt % (Gale, 1914), only about halfway to gaylussite saturation. Table 3 compares the reconstructed historic-lake composition (from an 1876 analysis of the anhydrous salts given in Gale, 1914) to that calculated from the evaporation model above for the same Cl concentration (13,300 mg/L). The model predicts pH, CO$_3$, and HCO$_3$ of the natural lake rather well, but it overestimates Na by a factor of 1.5 and SO$_4$ by a factor of 2.5. The abundance of reduced sulfur in the sediments (Tuttle, 1993) corroborated by SO$_4$ deficiencies in the pore waters (Bischoff et al., 1993b) suggests that significant SO$_4$ was removed by reduction in the bottom waters of the lake. The Na deficiency is explained, in part, by ion-exchange reactions in which river-suspended Ca smectites convert to Na smectites upon flocculation into the Na-dominated lake water (Jones and Van Denburgh, 1966). Clays suspended in Owens River water will have 99% of their exchange sites occupied by Ca, and conversely, those suspended in lake water of 1876 will be 99% occupied by Na, according to criteria of the U.S. Salinity Laboratory Staff (1954). The liberated Ca ions react to precipitate additional CaCO$_3$.

\[ \text{Ca}^{2+} + 2\text{HCO}_3^- = \text{CaCO}_3 + \text{H}_2\text{O} + \text{CO}_2 \]  

CaCO$_3$ budget and criteria of overflow

The time periods subtended by carbonate minima are short relative to the time periods of the carbonate maxima down the core, suggesting that closed lake conditions were periodically interrupted by brief periods of overflow (Fig. 2). The CaCO$_3$ content in the sediment can provide some information about when the lake was overflowing. The maxima of sediment CaCO$_3$ content in the core signal times of minimal overflow or closed lake conditions. It is also possible to approximate the limiting
content of CaCO₃ that marks the transition from overflow to closed conditions, given knowledge of lake area at spill, Ca flux from the Owens River, and sediment mass-accumulation rate. Owens Lake precipitates CaCO₃ within a minimum of 38 years after termination of intense spilling. Because the lake water is carbonate-dominated and Ca-limited, the rate of production of CaCO₃ in the closed lake is equal to the Ca flux provided by the Owens River. The flux of Ca is given by the product of the average river Ca concentration and average discharge. We calculate average Ca concentration for the lower Owens River for the years 1934–1992 to be 0.6 mmolal from unpublished LADWP data for 1934–1992. This concentration is almost identical to an average of 0.6 mmolal given by Hollett et al. (1991) from U.S. Geological Survey bimonthly analyses for 1974–1985. An average of 0.44 mmolal is calculated from data given in Gale (1914) for the years 1906–1912, and 0.72 mmolal from the data given by Wilcox (1946) for the years 1929–1946. Thus, average Ca concentration is between 0.44 and 0.70 mmolal. An additional source of Ca derives from Na-Ca exchange on suspended clays as they enter Owens Lake. Based on the ratio of CEC to acid-leachable Ca observed for the sediment, on the order of 5% of the river-supplied Ca could have been supplied by ion-exchange, or an additional 0.035 mmolal of Ca. Thus, including this ion-exchange of Ca for Na in the upper limit, the modern Ca flux is between 2 and 3.4 x 10⁸ mol/yr.

Because the Owens Lake basin is relatively flat, we consider the sediment area to be approximately equal to the surface area of the lake, or 6.22 x 10¹² cm² for the lake just at spill level. Therefore, the mass-accumulation rate (MAR) of CaCO₃ per unit area will be 0.03–0.055 mol/cm²/k.y. For the past 800 k.y. the bulk-sediment MAR has been approximately constant at 51.4 g/cm²/k.y. (Bischoff et al., this volume chapter 8). Therefore, assuming that CaCO₃ is evenly distributed over the floor of the lake, the limiting CaCO₃ content was between 0.06–0.11 mol/100 g sediment, or 6–11 wt % CaCO₃. Thus, sediments with less than about 0.11 mol/100 g represent overflowing conditions, whereas those with more represent closed lake conditions (Fig. 3). The pattern shown in Figure 3 for the past 500 k.y. indicates that the lake was overflowing about 34% of the time, and closed for about 66% of the time. The lake appeared to be continuously closed for periods of as much as 50- to 70-k.y. duration punctuated by briefer, 20- to 30-k.y. periods of intense overflow. In particular, the entire period from 120–50 ka seems to represent closed conditions. Data from downstream in Searles Lake, however, indicates that it was receiving spill water, at least periodically, during this same interval (Smith et al., 1983; Smith, 1991; Bischoff et al., 1985). Also, the patterns observed in the freshwater diatom remains suggest that there were several short-lived spilling events between 80 ka and 90 ka (Bradbury, this volume). Thus, it appears that these events were too short to be discernible within the 8-ka resolution of the channel samples.

In addition, the limiting CaCO₃ content is probably somewhat underestimated because of the assumptions that the Owens River is the only source of dissolved Ca, that the present Ca flux has been approximately constant through time, and that CaCO₃ is evenly distributed over the floor of the lake. For example, we do not consider eolian input of pedogenic CaCO₃. In addition, much of the dissolved load of the Owens River is supplied by the geothermal system of Long Valley (Smith, 1976). Thus, Ca flux of the Owens River may vary through time with the fluctuations in magmatic activity in the Long Valley caldera.

**Paleoclimate indicators and correlation to SPECMAP**

Cycling between closed and open lake conditions, best shown by CaCO₃ and CEC contents, likely reflects the cycles between glacial and interglacial, and these correlate reasonably well with the pattern of global glaciations. The oscillations of the oxygen-isotope record of marine foraminifers through time are now widely accepted as recording the mass of water stored in polar ice. In order to understand how the Owens Lake record reflects the global cycles, we compare the CO₃ and CEC patterns of OL-92 with the marine δ¹⁸O standard of SPECMAP (Imbrie et al., 1984) back to the major break in Owens sedimentary conditions at 500 ka (Fig. 6). The marine record back to about 700 ka is characterized by distinct 100-k.y. major cycles which terminate abruptly the progressively developing glacials (Termination), and with minor cycles of 41, 23, and 19 k.y. The minor cycles apparently reflect variations in insolation caused by astronomical rhythms, whereas the cause of the 100-k.y.

![Figure 6. Carbonate and CEC (in meq/100 g) variations in sediments from OL-92 compared to SPECMAP (Imbrie et al., 1984) for the past 500 k.y. of lacustrine conditions. Cycles in carbonate and CEC seem to reflect the dominant 100-k.y. cycles of SPECMAP. Terminations I, II, IV, and V, seem to be represented by abrupt increases in CEC and carbonate in the Owens record. Moreover, both carbonate and CEC indicate a marked glacial-interglacial transition at about 290 ka, which seems to have no counterpart in SPECMAP. The overflow model based on the CaCO₃ budget suggests that Owens Lake was over flowing about 34% of the time during the past 500 k.y.](image-url)
cycle remains in dispute (see, for example, Ruddiman and Wright, 1987; Winograd et al., 1992; Imbrie et al., 1993). The 100-k.y. cycles are reflected in the Owens Lake record (Fig. 6) where abrupt increases in CO$_3$ and CEC seem to correlate with at least four of the last five Terminations of SPECMAP. The minor cycles, however, are not revealed in spectral analysis of the Owens data, but the results are inconclusive because of the coarse resolution of the channel samples (8 k.y.) and the less-than-complete core recovery. Moreover, because the marine record reflects polar ice volume, and the Owens Lake record presumably reflects precipitation in the eastern Sierra Nevada, exact correlation should not be expected. Terminations I, II, IV, and V seem to be strongly reflected in the Owens record (Fig. 6), whereas Termination III is less obvious. Plots of CO$_3$, TOC, and CEC (Figs. 2 and 6) show a marked glacial/interglacial transition at about 290 ka in the Owens record, which seems to have no counterpart in SPECMAP. Chronological uncertainties in both SPECMAP, which assumes astronomical forcing for the 100-k.y. cycles, and the Owens core, which is based on inferences of constant mass accumulation rate between widely spaced controls (Bischoff et al., this volume, chapter 8) may explain the approximate ±5% differences in the positions of the Terminations. Changes in Owens sediment signaling the end of the last glacial are coincident with SPECMAP Termination I at about 13 ka. Changes that appear to relate to Termination II (oxygen-isotope stages 5/6 boundary) appear at 120 ka at Owens compared to 128 ka in SPECMAP. This Termination is shown very markedly in OL-92 for several other parameters as well, including smectite abundance (Menking, this volume), pollen, in which Pinus replaces cedars and junipers (Litwin et al., this volume), diatoms, in which saline forms replace freshwater forms (Bradbury, this volume), and ostracodes, in which the saline forms replace freshwater forms (Carter et al., this volume). In contrast, the fresh-water $\delta$18O record from nearby Devils Hole, Nevada, dated by high-precision U/Th (Winograd et al., 1992; Ludwig et al., 1992) show changes corresponding to Termination II at 140 ± 3(2s) ka. The 120-ka timing of this event at Owens is within 67° and 7° of SPECMAP and Devils Hole timing, a discrepancy perhaps within the uncertainty of the Owens chronology.

The warm period of oxygen-isotope stage 5 in SPECMAP extends from 128–70 ka (58 k.y. duration) where it is bounded by an abrupt cooling event, the stages 4/5 boundary. 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 k.y.) and corresponds to stage 5e, or whether it was long (60 k.y.) and corresponds to all of stage 5. The corresponding conditions at Owens Lake appear to span the entire time from 120 ka to about 50 ka, a similar time span to all of stage 5. Moreover, the Owens record seems to lack the prominent 5e of SPECMAP (Fig. 6). Thus, the Owens record seems to suggest that the “last interglacial” was long in the Western Great Basin. The approximately 10-k.y. offset between this span and marine oxygen-isotope stage 5 reflects either the error in the age-depth model (i.e., 10% error), 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.

Discrepancies notwithstanding, the general pattern of the global glacial cycles as recorded in the marine record appear to be reflected in the sediments of Owens Lake.

SUMMARY

The chemical composition of sediments from OL-92 and their variation through time allow the following conclusions:

1. Owens Lake sediments are composed predominantly of silts and clays of bulk composition close to granodiorites of the eastern Sierra Nevada. The sediment also contains significant but variable amounts of authigenic Ca and Mg carbonates and Mg silicates.

2. Records of CO$_3$ and TOC show cyclic co-variation back to about 500 ka. Maximum glacial conditions are clearly seen at 25–17 ka, where CO$_3$ and TOC display conspicuous and sharp minima, very close to zero values. These results suggest that at glacial maxima the lake was overflowing with cold fresh water and was relatively nonproductive. Five similar minima in CO$_3$ and TOC, going back to 500 ka are interpreted as successively older glacial maxima. Conversely, the recurring and broader CO$_3$ and TOC maxima are interpreted to represent full interglacial conditions during which the lake was closed, saline, and biologically productive. Closed lake conditions give rise to sediments rich in CO$_3$ and TOC, authigenic Mg-silicates, and clays of high CEC, in contrast to overflowing lake conditions where these components decrease to close to zero.

3. The CEC of the clay fraction reflects smectite abundance and shows a minimum coinciding with the glacial maximum at 25–17 ka. Other CEC minima occur in the core at the same points of CO$_3$ and TOC minima. This correlation suggests that during glacial advances the clay-size material has a low exchange capacity, perhaps representing a glacial rock-flour component.

4. Minerals characteristic of extreme salinity and/or playa conditions, such as gypsum and gaylussite, were not detected, suggesting that Owens Lake had not attained the required salinity at any time during the past 800 k.y. Calculations of progressive salinization of Owens Lake indicates that the first mineral to form beyond CaCO$_3$ is gaylussite, but not until the salinity reaches 15 wt %. For the lake to acquire this salinity at its spill level with present Owens River flux of dissolved salts 55 k.y. are required, but only 8 k.y. are required to reach the historic (1872) level.

5. The limiting sediment-content of CaCO$_3$ that separates closed and overflowing is about 11 wt %, calculated from 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. Applying this limit to the pattern of CaCO$_3$ abundance suggests that the lake was
closed about 66% of the time over the past 500 k.y., periodically interrupted by briefer periods of overflow.

6. The CaCO$_3$ oscillations generally track and correlate with those of the marine $\delta^{18}O$ cycles back to about 500 ka. Four of the last five marine isotope terminations are clearly shown in the Owens core. Terminations I, II, IV, and V are clearly interrupted by briefer periods of overflow.

in the Owens core. Terminations I, II, IV, and V are clearly shown at the appropriate times by abrupt increases in the carbonate content of the sediments, reflecting change from overflowing to closed lake conditions. The last interglacial at Owens Lake appears to span from 120 ka to about 50 ka. The roughly 10-k.y. offset between this span and marine oxygen-isotope stage 5 reflects either the error in the age-depth model (i.e., 10% error), 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.

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