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Unusual Holocene and late Pleistocene carbonate sedimentation in Bear Lake, Utah and Idaho, USA

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Abstract

Bear Lake (Utah–Idaho, USA) has been producing large quantities of carbonate minerals of varying mineralogy for the past 17,000 years. The history of sedimentation in Bear Lake is documented through the study of isotopic ratios of oxygen, carbon, and strontium, percent organic carbon, percent CaCO₃, X-ray diffraction mineralogy, HCl-leach inorganic geochemistry, and magnetic properties on samples from three piston cores. Historically, the Bear River, the main source of water for Great Salt Lake, did not enter Bear Lake until it was artificially diverted into the lake at the beginning of the 20th century. However, during the last glacial interval, the Bear River did enter Bear Lake depositing red, calcareous, silty clay. About 18,000 years ago, the Bear River became disconnected from Bear Lake. A combination of warmer water, increased evaporation, and increased organic productivity triggered the precipitation of calcium carbonate, first as calcite. As the salinity of the lake increased due to evaporation, aragonite began to precipitate about 11,000 years ago. Aragonite is the dominant mineral that accumulated in bottom sediments of the lake during the Holocene, comprising an average of about 70 wt.% of the sediments. Aragonite formation in a large, cold, oligotrophic, high latitude lake is highly unusual. Lacustrine aragonite usually is found in small, saline lakes in which the salinity varies considerably over time. However, Bear Lake contains endemic ostracodes and fish, which indicate that the chemistry of the lake has remained fairly constant for a long time. Stable isotope data from Holocene aragonite show that the salinity of Bear Lake increased throughout the Holocene, but never reached highly evolved values of $\delta^{18}\text{O}$ in spite of an evaporation-dominated water balance. Bear Lake hydrology combined with evaporation created an unusual situation that produced large amounts of aragonite, but no evaporite minerals.

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Keywords: Calcite; Aragonite; Organic carbon; Oxygen, carbon, and strontium isotopes; Leach chemistry

1. Introduction

Bear Lake is an oligotrophic, mesosaline lake in northeastern Utah and adjacent Idaho (Fig. 1) that presently occupies the southern end of the Bear Lake Valley, which is formed by a half graben (Colman, 2005). The lake is 32 km long and 6–13 km wide

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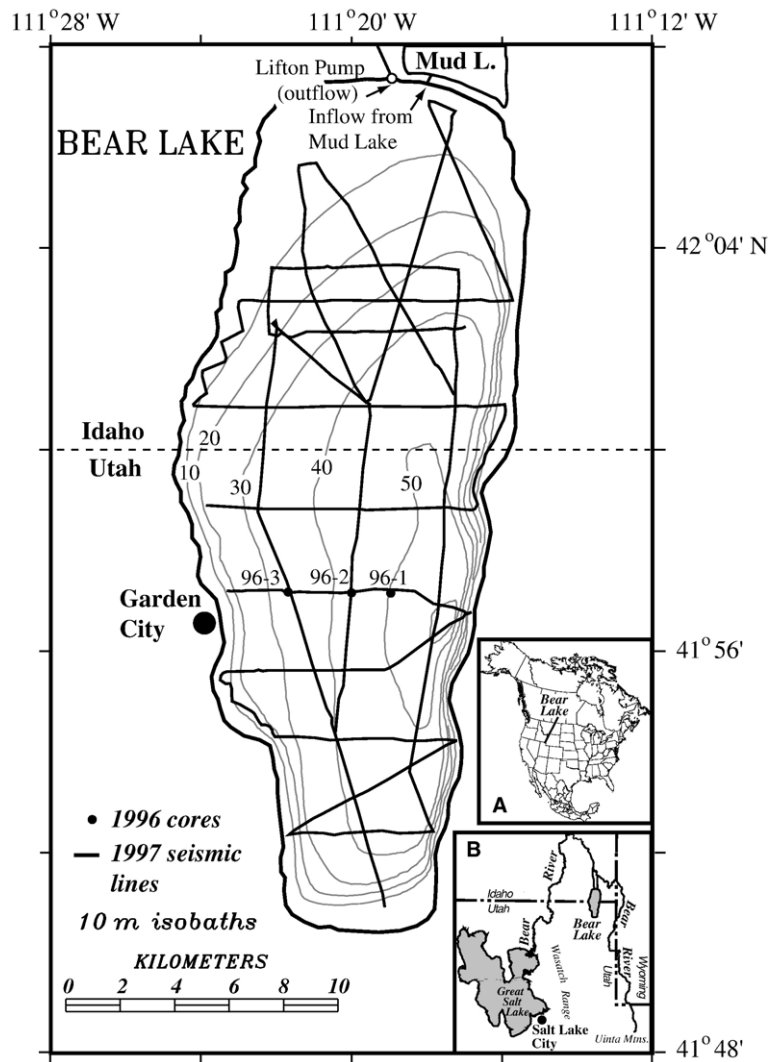


Fig. 1. Bathymetric map of Bear Lake Utah–Idaho, showing the locations of 1996 piston cores and 1997 acoustic-reflection lines. Inset A shows the location of Bear Lake within North America. Inset B shows the location of Bear Lake relative to Bear River and Great Salt Lake.

with an area of 280 km² at full capacity. Maximum depth is 63 m, with a mean depth of 28 m (Birdsey, 1989). The present elevation of the lake is 1805 m above sea level (mbsl) when full, but this level may have varied considerably through time with highstands of perhaps 11 m above the present level (Laabs and Kaufman, 2003). Historic low lake levels occurred during the 1930s and in 2004. Post-depositional uplift in this tectonically active valley hampers an accurate reconstruction of water levels in the valley. The natural watershed of the lake is relatively small, with a basin-area:lake-area ratio of only 4.8:1 (Wurtsbaugh and Luecke, 1997). Historically, the Bear River has not flowed into Bear Lake, but a part of its flow was diverted into Bear Lake through a series of canals,

beginning in 1909 with completion in 1918 (Birdsey, 1989), making Bear Lake a reservoir to supply irrigation water and hydroelectric power downstream. The diversion inflow increased the basin-area:lake-area ratio considerably to 29.5:1.

Acoustic-reflection (3.5 kHz) profiles obtained in 1997 and chirp (4–24 kHz) acoustic profiles and side-scan sonar data obtained in 2002 (Colman, 2005) indicate that the principal structure of the basin is a half graben, with a steep normal-fault margin on the east and a ramp margin on the west (Fig. 2). Acoustic reflectors diverge toward the master fault, forming eastward thickening sediment wedges, so that sedimentary units pinch out to the west. Acoustic profiles indicate that Holocene sediments are thin or missing

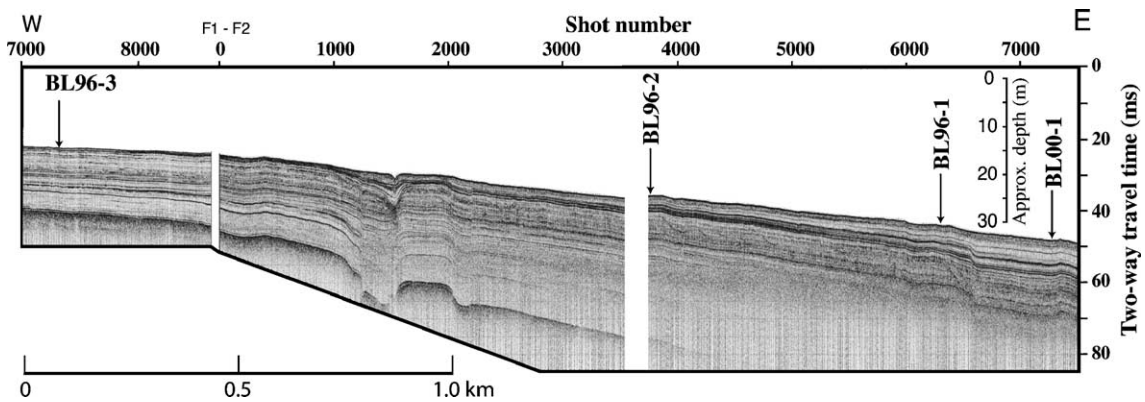


Fig. 2. Interpreted 2002 chirp acoustic-reflection profile across Bear Lake showing the locations of 1996 piston cores (Fig. 1; Colman, 2005). The location of the chirp line is approximately the same as the 1997 acoustic-reflection line west of Garden City, Utah in Fig. 1 showing the locations of the three cores.

in water depths shallower than about 30 m, but the deepest part of the lake contains at least several hundred meters of sediment, and the basin could be as much as six million years old (Colman, 2005). Acoustic profiles and other observations further indicate that there is reworking and redistribution of sediment from shallow sites to deeper sites (focusing) in Bear Lake (Colman, 2005). The presence of shallow-water diatoms in some horizons indicates some sediment transport from the lake margins to the center (Smoot, 2005).

The purpose of this paper is to examine the glacial to Holocene evolution of sedimentation in Bear Lake. In particular, why is aragonite the dominant carbonate mineral that formed during the Holocene in this large, cold, deep, oligotrophic, high latitude, montane lake?

2. Materials and methods

Piston cores were collected in 1996 from three localities in Bear Lake (Fig. 1) using the University of Minnesota, Limnological Research Center's Kullenberg coring system. Core BL96-1 is 5 m long from a water depth of 50 m. Core BL96-2 is 4 m long from a water depth of 40 m. Core BL96-3 is 4 m long from a water depth of 30 m. The three cores provide a composite section that records geochemical proxies of climate and environmental change over more than 30,000 years. The Holocene sediments are carbonate-rich (predominantly aragonite), and the glacial-age sediments consist of calcareous silty clay.

Representative detrital material from throughout the catchment was obtained by sampling alluvial sediment from sites along streams in the local Bear Lake catchment and along Bear River above Bear Lake (Fig. 3). Samples were subdivided by sieving into four grain-

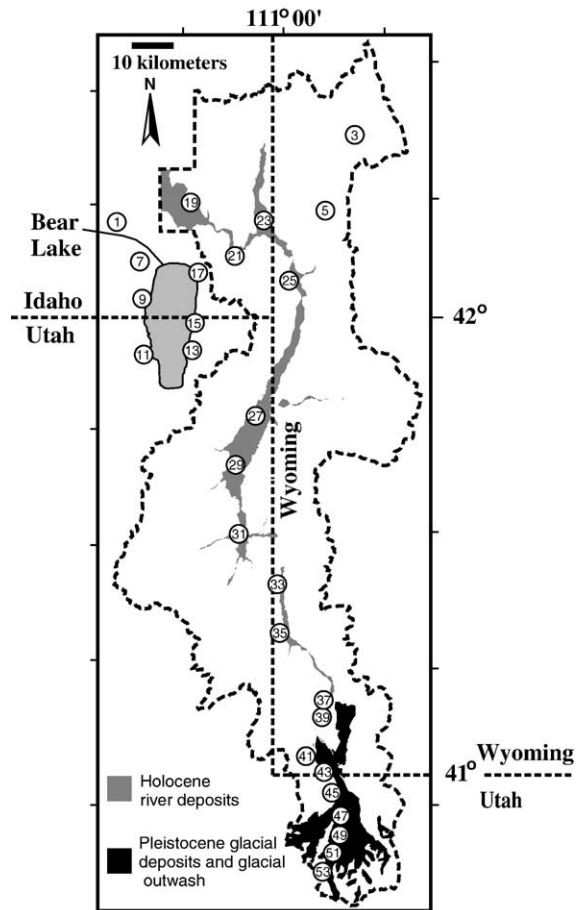


Fig. 3. Locations of stream-sediment samples. Odd and even numbered samples were collected at each site, but only the odd numbers are shown. Odd numbered samples were taken from the stream bottoms; even numbered samples were from stream-bank or overbank deposits.

size fractions: pebbles, coarse sand and granules, fine and medium sand, and silt plus clay (<63 μm).

Each of the three cores from the lake was cut into 1-m-long sections, and each section was sectioned longitudinally into a working half and an archive half. The working half of each section was cut into 1-cm-thick slabs for analyses. Most of the analyses described below were performed on every fourth 1-cm sample (e.g., 0–1 cm, 4–5 cm, 8–9 cm, etc.). Exceptions to this analytical scheme are noted.

Concentrations of total carbon (TC) and inorganic carbon (IC) in samples from the three piston cores were determined by coulometric titration of CO_2 following extraction from the sediment by combustion at 950 $^\circ\text{C}$ and acid volatilization, respectively (Engleman et al., 1985), in USGS laboratories, Denver, CO. Weight percent IC was converted to weight percent CaCO_3 by dividing by 0.12, the fraction of carbon in CaCO_3 . Organic carbon (OC) was determined by difference between TC and IC.

Semi-quantitative estimates of mineral contents in splits of the carbon samples and in the silt plus clay fractions of stream-sediment samples were determined by standard X-ray diffraction (XRD) techniques (e.g., Moore and Reynolds, 1989) in USGS laboratories, Denver, CO. Each sample was packed into an aluminum holder and scanned from 15 $^\circ$ to 50 $^\circ$ 2 θ at 2 $^\circ$ 2 θ /min using Ni-filtered, Cu-K α radiation at 45 kV, 30 ma, and peak intensities recorded as counts per second (cps). Results are reported as the peak intensity of the main XRD peak (I ; in cps) for each mineral, and as a percentage of the sum of the main XRD peak intensities of all minerals. These percentage calculations should be used with caution because they do not reflect the different X-ray mass absorption characteristics of different minerals. More quantitative estimates of aragonite and low-Mg calcite, the dominant minerals deposited over the last 18,000 years, were calculated by partitioning the percentage of total CaCO_3 , determined by coulometry, using the intensity ratio of the main XRD peak of aragonite to that of low-Mg calcite ($I-A/I-C$), and curves of percent aragonite (of total CaCO_3) versus $I-A/I-C$ as determined by Chave (1954) and Lowenstam (1954). The curve of Lowenstam (1954) is closely represented by the equation

$$\begin{aligned} \% \text{ aragonite} = & -4.4992 + [105.3188 \\ & \times (I - A/I - C)]/[0.2362 \\ & + (I - A/I - C)]. \end{aligned} \quad (1)$$

Measurements of ratios of stable isotopes of carbon and oxygen were made in the isotope laboratory in the

Department of Geology and Geophysics, University of Minnesota. Samples were reacted with 100% anhydrous phosphoric acid at 75 ± 1 $^\circ\text{C}$. Evolved gases were cryogenically purified to remove water and non-condensable gases. The purified CO_2 was introduced into an isotope-ratio mass spectrometer through a capillary, and isotopic ratios of carbon and oxygen were measured against a reference standard of known isotopic composition. Carbon- and oxygen-isotope measurements on the silt plus clay fractions of seven stream-sediment samples were made in the isotope laboratory at the University of Arizona. Results of all analyses are reported in the standard per mil (‰) δ -notation relative to the Vienna Pee Dee Belemnite (VPDB) marine-carbonate standard for carbon and oxygen:

$$\delta\text{‰} = [(R_{\text{sample}}/R_{\text{VPDB}}) - 1] \times 10^3 \quad (2)$$

where R is the ratio ($^{13}\text{C} : ^{12}\text{C}$) or ($^{18}\text{O} : ^{16}\text{O}$).

Measurements of Sr isotope ratios were made on samples from core 96-2, and silt plus clay fractions of stream-sediment samples, in USGS laboratories in Denver, CO. Samples were leached in 5 M acetic acid, and the leachate was centrifuged and purified with conventional ion-exchange methods. Samples were loaded on a single tantalum filament with phosphoric acid. Isotope ratios were measured with an automated VG54 sector multi-collector, thermal ionization mass spectrometer in dynamic mode. Mass dependent fractionation was corrected assuming a $^{86}\text{Sr}/^{87}\text{Sr}$ ratio of 0.1194. Strontium isotope ratios are reported relative to SRM-987 standard value of 0.71025.

The silt plus clay fractions of stream-sediment samples were analyzed for 40 major, minor, and trace elements by inductively coupled, argon-plasma, atomic-emission spectrometry (ICP-AES) by XRAL Laboratories, Toronto, Canada. Rock standards (USGS) were included with the sediment samples, and 10% of the samples were duplicated. The precision, determined by analyzing rock standards and duplicate sediment samples, is better than 10%, and usually is better than 5%, at a concentration of 10 times the limit of detection.

Analyses of acid-soluble elements were conducted in USGS laboratories in Menlo Park, CA. Samples were leached in 3 N HCl overnight, and the supernatant was analyzed by inductively coupled, argon-plasma, atomic-emission spectrometry (ICP-AES) for major components (percent) Ca, Fe, and Mg, and minor components (parts per million, ppm) Mn, Ba, and Sr (Bischoff et al., 2005).

Wet sediment from every other 1-cm sample from core 96-2, and the silt plus clay fraction of stream-sediment samples, was placed in a nonmagnetic 3.2 cm³ plastic box for measurements of magnetic properties. Hard isothermal remanent magnetization (HIRM; see King and Channel, 1991) was determined on samples from core 96-2 and on fluvial samples from throughout the catchment (Fig. 3). Isothermal remnant magnetization was first imparted in a field of 1.2 T (IRM_{1.2T}), and measured with a high-speed spinner magnetometer. A second IRM was then imparted in the opposite direction in a field of 0.3 T (IRM_{0.3T}). HIRM, calculated as (IRM_{1.2T} – IRM_{0.3T})/2, largely reflects the content of high-coercivity minerals such as hematite and greigite.

Accelerator mass spectrometer (AMS) ¹⁴C ages were obtained from samples of various materials including pollen concentrates, ostracodes, and bulk-sediment OC (Colman et al., 2005; Table 1). The “pollen” samples were prepared using standard palynological methods and contain some charcoal and other refractory organic matter. Reservoir corrections were made for AMS dates on bulk OC by subtracting 480 years from the raw dates, and by subtracting 370 years from dates on ostracodes (Colman et al., 2005). Reservoir-corrected and calibrated radiocarbon ages are expressed in kiloyear BP (e.g., 26.6 cal ka). Raw and calibrated radiocarbon ages for each of the three cores are given in Table 1.

3. Results

3.1. The cores

Each of the Kullenberg piston cores collected in 1996 penetrated a maximum of 5 m. The collected sediments were deposited over the last 31 cal ka, well beyond the last glacial maximum (LGM; ca. 21 cal ka). As seen in Fig. 2, the sedimentary sequence in Bear Lake consists of a series of eastward dipping strata the shallowest of which pinch out to the west. Therefore, the three cores (96-1, 96-2, and 96-3, Fig. 1) contain overlapping sequences. This can be seen in the acoustic profiles in Fig. 2, and in the profiles of calcite and aragonite versus depth shown in Fig. 4. The upper aragonite-dominated sequence, which was deposited over the last 7 ky, is present throughout the 500 cm of section in core 96-1, and in the upper 170 cm in core 96-2. Based on ¹⁴C ages, the bottom of core 96-1 just missed the top of the calcite-dominated sequence that occurs at 170–200 cm in core 96-2. This calcite layer in core 96-2 overlies a lower ara-

Table 1

Radiocarbon ages on materials separated from sediment samples from cores BL-96-1, BL-96-2, and BL96-3

Core ID	Depth (cmblf)	Material	¹⁴ C age (years)	Error (years)	Calib. age (cal year BP)
<i>BL96-1</i>					
1-A-52	48	Pollen	1070	40	970
1-C-55	252	Pollen	3320	40	3545
1-E-3	401	Pollen	5260	40	6055
1-E-95	493	Bulk (TOC)	6740	50	7555
<i>BL96-2</i>					
2-A-13	3	Bulk (TOC)	1620	50	1480
2-B-9	99	Pollen	3435	60	3690
2-B-31	121	Ostracodes	4620	60	5320
2-B-31	121	Pollen	4230	40	4830
2-B-61	151	Ostracodes	5460	50	6280
2-B-61	151	Pollen	5260	40	5990
2-B-73	163	Pollen	5810	50	6625
2-B-85	175	Pollen	6420	50	7300
2-B-85	175	Bulk (TOC)	6970	50	7775
2-C-10	201	Pollen	8265	70	9170
2-C-21	212	Ostracodes	9070	60	10,220
2-C-21	212	Pollen	8580	40	9540
2-C-55	246	Pollen	10,300	60	12,135
2-D-7	299	Pollen	12,710	50	14,960
2-D-7	299	Bulk (TOC)	13,110	60	15,605
2-D-8	300	Pollen	12,545	90	14,670
2-D-17	307	Rotifer	12,400	80	14,515
2-D-21	313	Pollen	14,000	100	16,790
2-D-41	333	Pollen	16,200	75	19,320
2-D-61	353	Pollen	18,550	140	22,025
2-D-81	373	Pollen	21,000	110	24,750
2-D-93	385	Pollen	21,300	150	25,095
2-D-101	393	Pollen	22,600	60	26,600
<i>BL-96-3</i>					
3-A-18	5	Pollen	4440	40	5010
3-A-33	20	Pollen	10,940	75	12,980
3-A-46	33	Pollen	12,800	100	15,440
3-A-90	77	Pollen	19,980	60	23,670
3-C-13	201	Pollen	21,850	230	25,735
3-C-89	277	Pollen	23,400	130	27,520
3-D-89	377	Pollen	26,700	170	31,280

Data from Colman et al. (2005).

cmblf=centimeters below lake floor.

TOC=total organic carbon.

gonite-dominated sequence. Core 96-3 contains only a small amount of Holocene sediment. Aragonite that occurs at the top of core 96-3 may correspond to part of the lower aragonite sequence, or it may contain aragonite from both the upper and lower aragonite sequences.

The 400-cm section recovered in core 96-2 contains all of the lithologic units deposited over the last 26,000 cal. year, and appears to be nearly continuous and intact. Therefore, we will focus our attention on that core. A second-order polynomial fit through the cali-

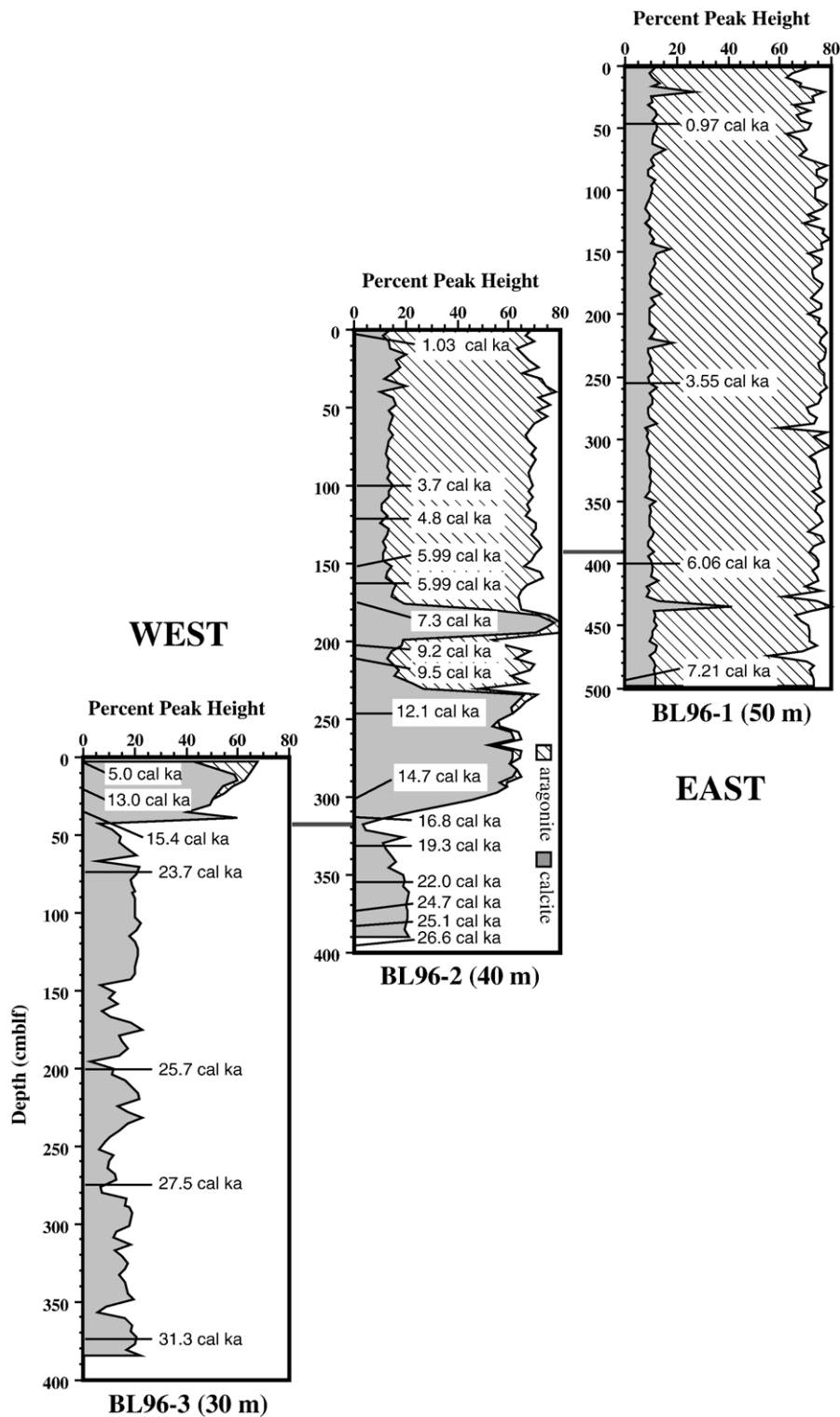


Fig. 4. Profiles of values of % calcite and % aragonite (based on XRD peak intensities) in cores 96–1, 96–2, and 96–3 showing correlations between cores based on carbonate mineralogy and radiocarbon ages. Locations of most radiocarbon ages are shown, and all radiocarbon ages are given in Table 1. Core 96–2 appears to contain the entire sequence deposited over the last 26,600 years.

brated ages of Colman et al. (2005) was used to calculate the ages of all samples. According to the age model, about 900 years of sediment record are missing from the top of core 96-2. This missing time may indicate an unconformity and (or) loss of sediment during coring. Because resuspension and focusing of sediment from shallower to deeper parts of the lake are known to be active processes in Bear Lake (Colman, 2005) hiatuses may exist in the condensed Holocene section in core 96-2 (relative to that in core 96-1). The relatively smooth fit of radiocarbon data versus depth (Colman et al., 2005) suggests that there are no major gaps in the record.

3.2. Mineralogy

The hydrologic and chemical evolution of Bear Lake is reflected in both the allogenic and endogenic minerals found in the sediments. Sediments deposited between 26 and 18 cal ka consist predominantly of quartz with minor amounts of low-Mg calcite and dolomite (Unit 1, Fig. 5). The quartz (and minor feldspar, not shown) is detrital, brought in by the Bear River during the last glacial interval (LGI; 24–15 cal ka). Some or all

of the low-Mg calcite and dolomite deposited during this time also may be detrital. Most of core 96-3, like the bottom of core 96-2, consists predominantly of quartz with minor low-Mg calcite, dolomite, and feldspar. Sediments of late glacial age (18–16 cal ka) in core 96-2 are characterized by decreasing amounts of quartz, suggesting that the Bear River was in the process of abandoning Bear Lake, and they also contain increasing amounts of calcite, indicating increased precipitation of endogenic CaCO_3 (Unit 2, Fig. 5). Sediments deposited during the transition between the LGI and Holocene (16–11 cal ka) consist of about 40% CaCO_3 as low-Mg calcite, about 30% quartz, and minor dolomite and aragonite (Unit 3, Fig. 5). The percentage of low-Mg calcite in Unit 3 calculated from XRD peak height is about 60% (Fig. 4). The peak-height method overestimates % calcite and underestimates % quartz and % aragonite. The true quartz content in Unit 3 probably is more like 50–60%.

Precipitation of CaCO_3 abruptly changed from predominantly calcite to predominantly aragonite at about 11 cal ka (Unit 3–4 boundary, Fig. 5). Aragonite precipitation dominated deposition for about 2 ky (Unit 4, Fig. 5), then, just as abruptly, returned to

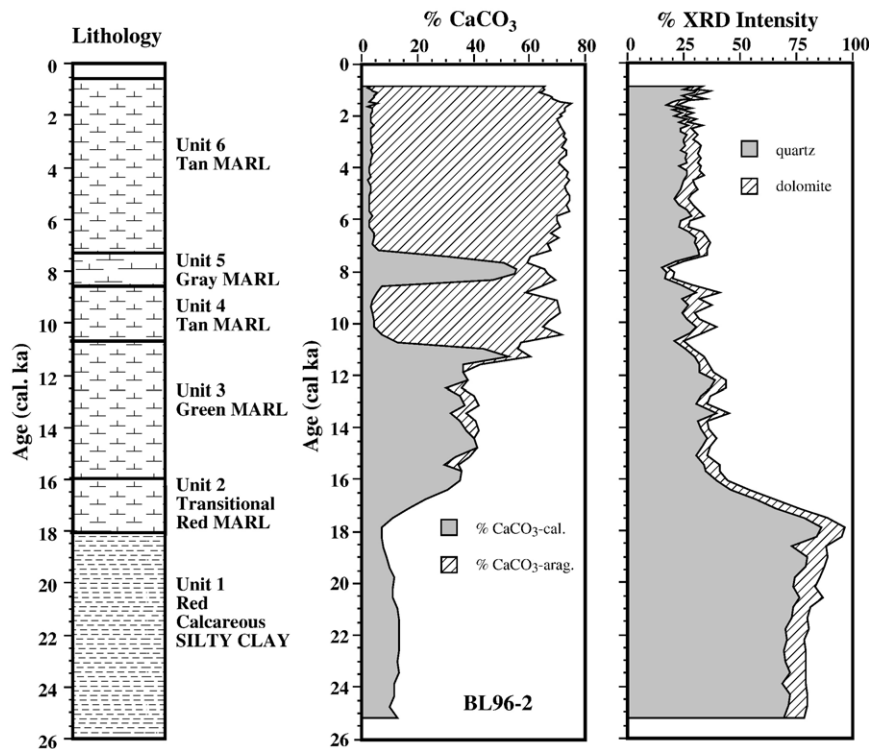


Fig. 5. Lithology and profiles of percent CaCO_3 as aragonite and calcite (see text for method of calculation), and percentages of dolomite and quartz (calculated from XRD peak intensities), all versus age (in cal ka), for core 96-2.

calcite precipitation for another 1 ky (Unit 5, Fig. 5). By 7 cal ka, aragonite again dominated the sediments, with minor quartz, low-Mg calcite, and dolomite (Unit 6, Fig. 5). This mode of carbonate sedimentation continued for the rest of the Holocene. Core 96-1, like Unit 6 in core 96-2, consists predominantly of aragonite with minor low-Mg calcite (Fig. 4), dolomite, and quartz. Note that the percent aragonite in core 96-2 calculated from XRD peak intensities (Fig. 4) is greatly underestimated, and percent calcite greatly overestimated. This is because of the large difference in the X-ray mass absorption characteristics of the two minerals. The partitioning of CaCO_3 between aragonite and calcite based on peak intensity ratios (Eq. (1)) gives a more accurate estimate of the proportion of aragonite to calcite (Fig. 5).

3.3. Magnetic properties

Values of MS and HIRM can be used to document differences in magnetic mineralogy within the Bear Lake and Bear River catchments (Fig. 6). Based on these magnetic properties, the sampled region (Fig. 3) can be divided into three areas. First, sediments in the local catchment of Bear Lake are characterized by MS values that are about three times higher than those along the Bear River. Second, the Bear River above Bear Lake and below the Utah–Wyoming border is characterized by low values of both MS and HIRM. Third, the Bear River south of Wyoming is characterized by low MS but high HIRM, probably arising from hematite-cemented Neoproterozoic rocks in the Uinta Mountains (Ashby et al., 2001). This hematite cement may be the main source of high iron (Fe) contents of Bear River sediments (Fig. 6), although red iron oxide coatings also supply some of the Fe. The Fe content of Bear River sediments decreases more gradually than HIRM downstream from the Uinta Mountains.

Variations in magnetic properties in core 96-2 in part reflect differences in magnetic properties of the source areas of detrital clastic material. In comparison to the entire core, the lower part of Unit 1 is characterized by high HIRM and moderate MS (Fig. 7). Values of HIRM and MS are consistent with sediment older than 24 cal ka being comprised mostly of hematite-rich sediment derived from the headwaters of Bear River in the Uinta Mountains. No other source of sediment can explain the high values of HIRM without calling upon some mechanism to concentrate the high-coercivity magnetic phases in the lake sediments. The hematite-rich material is interpreted to be glacial flour because (1) a large fraction of the sediment

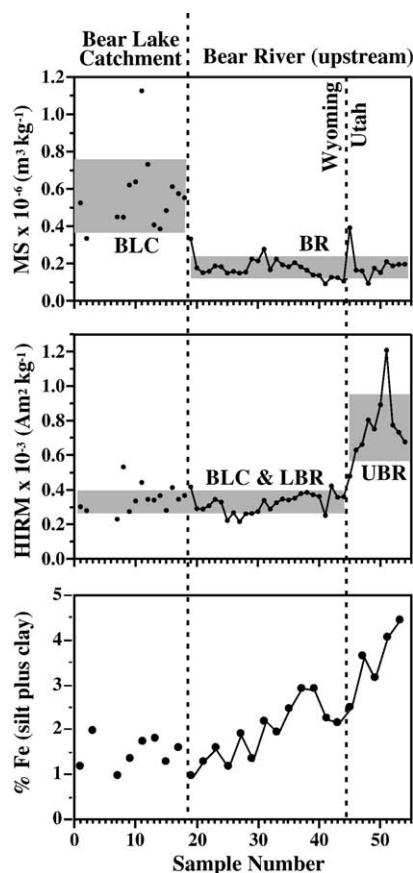


Fig. 6. Magnetic susceptibility (MS), hard isothermal remnant magnetization (HIRM), and percent Fe of the silt plus clay fractions of catchment samples. Sample locations are shown in Fig. 3. The shaded boxes are centered on mean MS and HIRM values for Bear River (BR), upper Bear River (UBR), Bear Lake catchment (BLC), and BLC plus lower Bear River (LBR). The boxes extend to one standard deviation on either side of the mean.

apparently was derived from a small fraction of the Bear River catchment that was glaciated during the LGI, and (2) the age of this part of the section falls within the LGI. Attribution of these sediments to a source in the Uinta Mountains requires that the Bear River was connected to Bear Lake at this time.

After 25 cal ka, HIRM decreased to a local minimum at about 23 cal ka, increased for a brief interval, and then began a long decrease to about 11 cal ka. In the lower part of this interval (upper part of Unit 1; Fig. 7) MS varies inversely with HIRM until reaching a maximum at about 18 cal ka. The inverse relation between HIRM and MS suggests that the content of hematite-rich glacial flour from the Uinta Mountains decreased and was replaced by sediment derived from other parts of the catchment including material with relatively high MS values from the local Bear Lake catchment. Values of HIRM at 18 cal ka are consistent

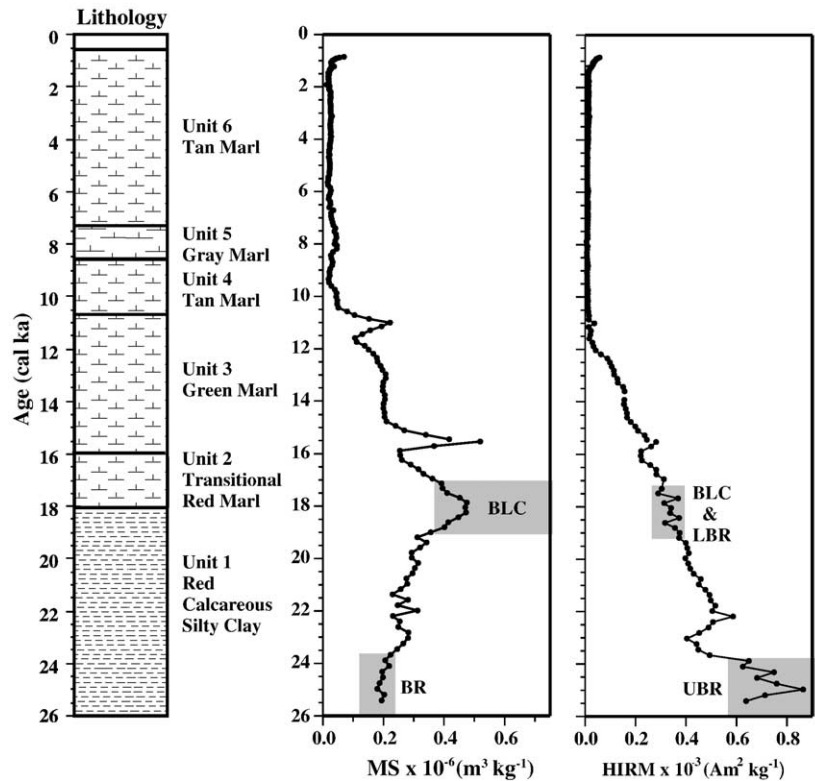


Fig. 7. Lithology, magnetic susceptibility (MS), and hard isothermal remnant magnetization (HIRM) versus age in core 96-2. The shaded boxes extend to one standard deviation on either side of the mean for catchment samples from the areas indicated in Fig. 6.

with sediment composed of detrital material containing little, if any, material from the Uinta Mountains. The MS values in sediment at the top of Unit 1 indicate that it contains about 75% material from the local Bear Lake catchment and 25% from Bear River.

Above Unit 1, both MS and HIRM decrease upwards. Two well-defined peaks in MS interrupt this trend to lower values. The origins of these peaks are not firmly established, but magnetite was observed in a few glass shards in a magnetite separate from the upper of these two peaks (R. Reynolds, written communication, 2004), suggesting that this peak is due to disseminated volcanic ash. The two peaks have similar wavelengths and shapes suggesting similar origins. The decreases in HIRM and MS across Unit 2 (by factors of 1.5 and 2, respectively) can be largely explained by dilution of detrital material by endogenic carbonate (Fig. 5). The further upward decrease in HIRM and MS throughout Unit 3 and into the base of Unit 4 cannot be explained by further dilution because the change in content of detrital material (e.g., quartz; Fig. 5) is relatively small, whereas both HIRM and MS decline by more than an order of magnitude. In Holocene sediments, ferrimagnetic minerals have been

so thoroughly destroyed by post-depositional alteration that paramagnetic minerals probably account for a significant fraction of MS.

3.4. Carbon content

Although the carbonate minerals in Holocene sediments in core 96-2 are a mixture of calcite and aragonite (with minor dolomite), the overall bulk CaCO_3 content is relatively constant at about 70% (Fig. 8). This suggests that there was a mass balance of total CaCO_3 that was maintained through changes in mineralogy, which presumably reflect changes in environmental conditions (discussed below). The increase in CaCO_3 in Holocene sediments, relative to LGI sediments, could represent an increase in the rate of CaCO_3 accumulation in the sediments, or it could represent lack of dilution by detrital clastic material (the red clay) that dominated the LGI sediments. We used sedimentation rates, calculated from the age model, and measured dry bulk densities to calculate bulk-sediment mass accumulation rates (MAR), and MARs of CaCO_3 and OC (Fig. 8). The MAR of CaCO_3 does show considerable variation throughout the Holocene section,

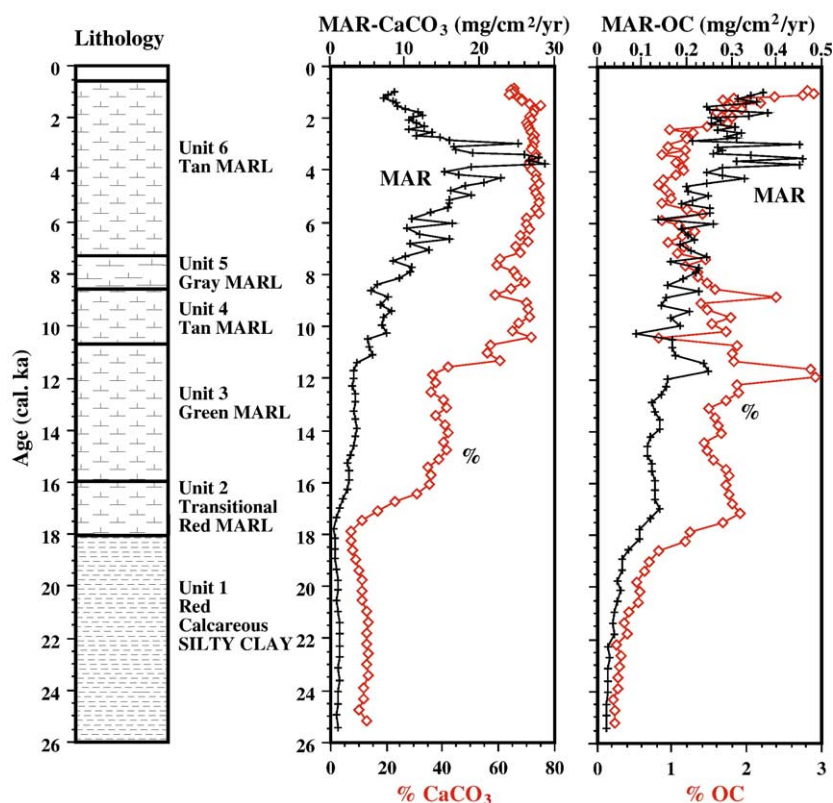


Fig. 8. Lithology, and profiles of percentages and mass accumulation rates (MAR) of CaCO_3 , and organic carbon (OC) versus age in core 96-2.

peaking in sediments deposited between 3 and 4 cal ka. In the late glacial-age sediments that contain only calcite (16 to 11 cal ka), the overall CaCO_3 content is about 40% (Fig. 5). Smoot (2005) described the fine-grained sediments deposited during the glacial-Holocene transition and all of the Holocene as various colored “clays”. The CaCO_3 content of all of these sediments is high enough ($>30\%$) to classify them as marls (Dean, 1981; Fig. 8). The red clayey glacial sediments contain enough CaCO_3 (about 5%) to be classified as calcareous clay. These red, carbonate-poor LGI sediments are in marked contrast to the light tan, aragonitic Holocene marls (Fig. 5).

The OC content of glacial-age sediments is $<0.5\%$, but increases gradually to about 2.0% in sediments deposited about 17 cal ka (Fig. 8). The OC content decreases in Holocene sediments (from about 2.0% to 1.0%), but increases abruptly in sediments deposited between 3 and 1 cal ka to almost 3%. The OC MAR, like % OC, increases upward in glacial sediments, fluctuates considerably in Holocene sediments, and, like CaCO_3 MAR, peaks in sediments deposited between 3 and 4 cal ka (Fig. 8). The gradual increases in percentage and MAR of OC throughout the LGI suggest that lake productivity was increasing through the LGI.

3.5. Carbon and oxygen isotopes

The range of values of $\delta^{18}\text{O}$ (8‰) and $\delta^{13}\text{C}$ (6‰) in carbonate is considerable (Fig. 9). Values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in Holocene marls of a typical hard-water, through flowing, hydrologically open, north-temperate lake usually vary by 2–4‰ (e.g., Talbot, 1990). In general, values of both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ increase from glacial to Holocene sediments, but with considerable, and important, fluctuations, particularly in values of $\delta^{18}\text{O}$ (Fig. 9). A high positive covariance exists between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in marls deposited after 16 cal ka (Fig. 10A). This correlation is independent of carbonate mineralogy (aragonite or calcite), but, as can be seen in Fig. 9, values of $\delta^{18}\text{O}$ tend to be lower (more negative) in intervals dominated by calcite than in those dominated by aragonite. Values of $\delta^{13}\text{C}$ also are lower in the calcite deposited between 9 and 7 cal ka (Fig. 9). Values of $\delta^{13}\text{C}$ in calcite that is present in low amounts in glacial sediments deposited before 18 ka are depleted in ^{13}C by about 2‰ relative to the more abundant calcite deposited between 16 and 12 cal ka (Figs. 9 and 10A). The glacial-age sediments are even more depleted than the carbonate rocks (limestones and dolomites) in the Bear River Range to the west of the lake as indicated by an average value of $\delta^{13}\text{C}$ of 0.13‰

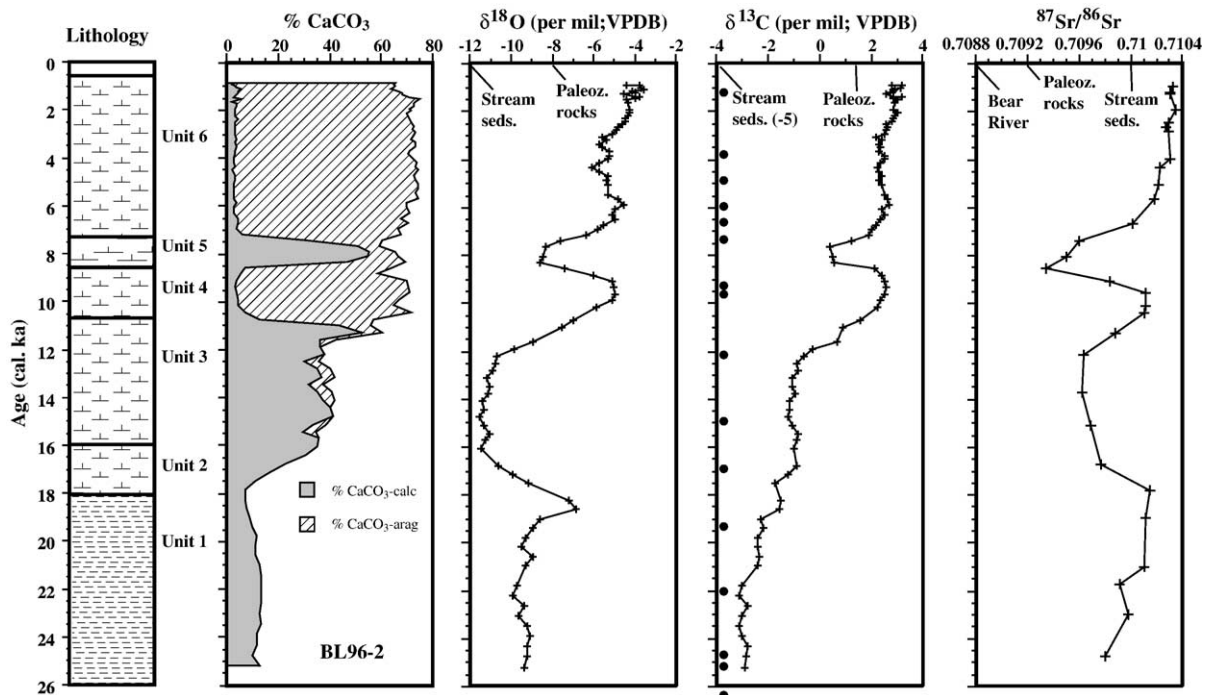


Fig. 9. Lithology, and profiles of percent CaCO_3 as aragonite and calcite (see text for method of calculation), and of $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, and $^{87}\text{Sr}/^{86}\text{Sr}$ in bulk carbonate versus age in core 96-2. The black dots on the left margin of the profile for $\delta^{13}\text{C}$ are the locations of radiocarbon dates.

(range, -2 to 2.8‰) in 10 samples of these rocks (Bright et al., in press). Values of $\delta^{18}\text{O}$ in glacial-age sediments are similar to those in the same 10 samples of Paleozoic carbonate rocks (average, -7.9‰ ; range, -13 to -5‰).

A positive covariance between values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in lacustrine carbonates often occurs in water bodies with relatively long residence times, and if the correlation is high ($r > 0.7$), this implies that the lake was hydrologically closed (Talbot, 1990). Covariant trends may have remarkably long-term persistence through major environmental changes (Talbot, 1990), and this certainly applies to Bear Lake.

Values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in the aragonite in upper Holocene marls recovered in cores 96-1 (Fig. 10B) and 96-2 (Fig. 10A) are highly enriched in ^{18}O and ^{13}C relative to the calcite-rich marls deposited during the glacial-Holocene transition (Figs. 9 and 10A). Similarly, low-Mg calcite in glacial-age calcareous clay in core 96-3 has values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ (Fig. 10B), which are comparable to those in glacial-age calcareous clay in core 96-2 (Fig. 10A).

3.6. Strontium isotopes

Overall the profile of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in carbonate samples from core 96-2 is similar to those of $\delta^{13}\text{C}$, and ^{18}O , although there are fewer analyses (Fig. 9). The

very low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in Bear River water (0.7088; Fig. 9) should be an excellent tracer (Bouchard et al., 1998) of when Bear River entered Bear Lake. Based on this criterion, it appears that some Bear River water was entering the lake prior to 12 cal ka, and between 9 and 7 cal ka, an interpretation that is consistent with carbon and oxygen isotope data also indicating reentry of Bear River. Ratios of $^{87}\text{Sr}/^{86}\text{Sr}$ of about ~ 0.71 in carbonate deposited before 18 cal ka probably represent a mixture of endogenic carbonate with values < 0.7095 and detrital carbonate with values > 0.71 . Carbonate in the silt and clay fraction in Bear River sediments north of the Uinta Mountains have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios about 0.7107 (Table 2), and the silt and clay fraction of one sample from the Uinta Mountains had a $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7191 (Table 2). Ratios of $^{87}\text{Sr}/^{86}\text{Sr} > 0.71$ in sediments deposited after 7 cal ka reflects the influence of waters and carbonate sediments in west-side creeks and springs. West-side waters must have dominated the pre-diversion hydrologic budget of the lake by a percentage estimated to have been 95–99% (Dean et al., 2005).

The sediments in west-side creeks contain abundant low-Mg calcite and dolomite (Table 2), presumed to be derived from Paleozoic carbonate rocks in the Bear River Range to the west of the lake. However, the average $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in 20 samples of these carbonate rocks is 0.7092 (range 0.7071–0.7161; Bright et al., in

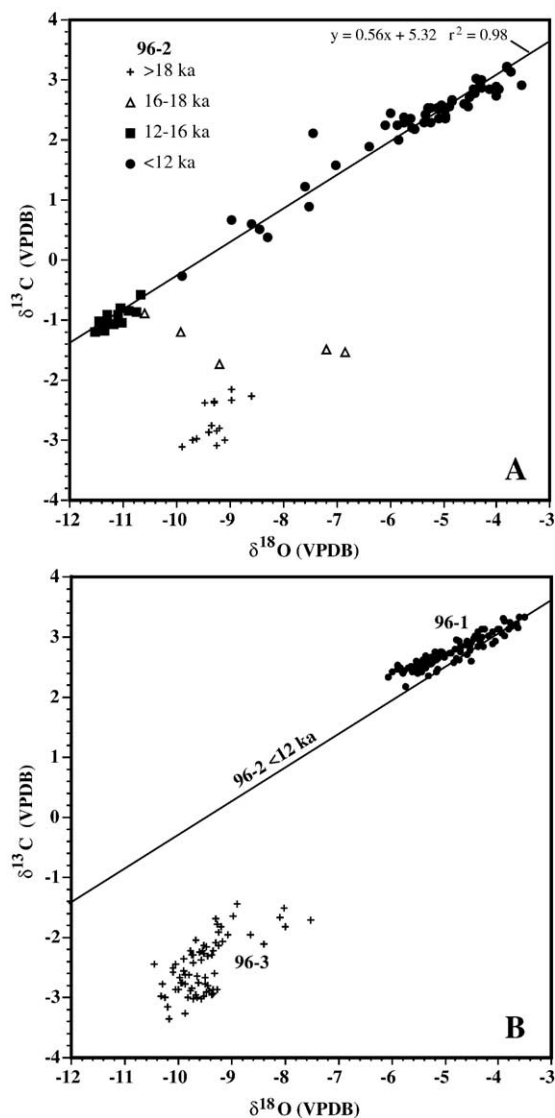


Fig. 10. Cross plots of values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in samples from core 96-2 (A), and cores 96-1 and 96-3 (B). The regression line and equation in panel A is for samples from 96-2 younger than 12 ka; the regression line is repeated in panel B.

press), which is depleted relative to bulk carbonate in stream sediments (Table 2). Endogenic tufa picked from the sediments of several west-side streams have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios that are similar to those in bulk carbonate in stream sediments (0.710+; Table 2), and similar to ratios in bulk carbonate deposited in the lake over the last 4000 years (Fig. 9).

3.7. Leach chemistry

Concentrations of HCl-extractable elements in bulk carbonate generally reflect the mineralogy of the sedi-

ments. Highest concentrations of Mg and Mg/Ca ratios generally are associated with low-Mg calcite and the red clay, which contains dolomite, whereas highest concentrations of Sr and Ba, and highest Sr/Ca ratios, are associated with aragonite (Fig. 11). Slightly elevated Ba concentrations in the calcareous red clay probably are due to substitution of Ba for Ca in detrital carbonate. The average bulk-sediment Ba concentration in the silt and clay fraction of carbonate-bearing stream sediments in the Bear River south of Bear Lake is 400 ppm ($n=14$; samples 17–53, Fig. 3) with maximum values of 1000 (samples 51 and 53). The peak in Mg/Ca in sediments deposited about 18 cal ka (Fig. 11) may be due to an influx of dolomite from the west-side creeks because the high MS values in those sediments (Fig. 7) suggest a predominance of sediment from the local Bear Lake catchment. However, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in sediments deposited in the lake over the last 4000 years are more indicative of those in endogenic carbonate than in Paleozoic carbonate rocks (Fig. 9). Paleozoic carbonate rocks have much lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (ca. 0.7092), but there is no indication of sediments with these lower ratios entering the lake.

The calcareous red clay deposited during the LGI when Bear River was connected to Bear Lake contains relatively high concentrations of Mn and Fe (Fig. 11), some of which was extracted by the HCL leach. Concentrations of Fe are expressed on a carbonate-free (cf) basis to remove the dilution effect of CaCO_3 , thereby revealing actual changes in the character of Fe in the noncarbonate fraction. Concentrations of Fe in the silt plus clay fractions of bulk stream-sediment samples from the Uinta Mountains are up to 4.5% (Fig. 6). Much if not most of the Fe probably is present as hematite based on the high HIRM values (Fig. 6), but some is present as iron oxide coatings on the red silty clay. Some of the Mn probably is present as oxyhydroxides particularly in the red silty clay, but the relatively high Mn contents of the calcite formed at about 8 cal ka and prior to 11 cal ka (Fig. 11) indicate that Mn also is substituting for Ca in calcite.

4. Discussion: the glacial–Holocene evolution of the Bear Lake chemical system

4.1. Evolution of other Lakes in the Great Basin

The presence of red calcareous clay deposited in Bear Lake during the LGI coincides in time with the growth of large lakes in the northern Great Basin, including Bonneville, Lahontan, and Russell (Thompson, 1990; Thompson et al., 1986, 1993; Benson and

Table 2

X-ray diffraction (XRD) mineralogy and oxygen, carbon, and strontium isotopic composition of the silt+clay (<63 μm) fraction of stream sediments in the Bear Lake catchment

Bear River spl. # stream	% quartz	% dolomite	% calcite	% feldspar	% CaCO_3	del ^{18}O -carb	del ^{13}C -carb	87Sr/86Sr
<i>West-Side Creeks</i>								
1 Paris Creek	41.61	55.62	0.00	2.76	0.0	−14.3	−6.06	
1 Paris Creek						−8.4	−0.20	
3 Hobble Creek	68.57	2.88	13.64	14.91	15.7	−12.4	1.20	
7 St. Charles Creek	51.99	36.05	5.58	6.38		−13.0	−5.74	0.71004
9 Fish Haven Creek	57.84	25.60	11.48	5.07				0.71030
11 Hodges Canyon	81.42	2.88	4.81	10.90	0.0	−14.9	−8.20	
Swan Creek						−13.4	−5.20	
Bloomington Creek						−15.8	−7.20	
<i>East-Side Creeks</i>								
13 South Eden Cyn.	63.44	4.35	28.09	4.12	25.6	−13.4	−7.25	
13 South Eden Cyn.						−11.4	−5.00	
15 North Eden Cyn.	84.60	3.28	8.20	3.93	10.8	−13.5	−4.76	
15 North Eden Cyn.						−12.4	−5.00	
17 Indian Creek	69.24	2.74	21.15	6.86		−11.6	−4.49	0.70768
<i>B.R. N. of Uintas</i>								
Bear River						−12.6	−4.80	
Bear River						−12.8	−6.54	
19 Bear River	84.67	2.94	8.95	3.44	9.9	−12.7	−5.00	
21 Bear River	75.25	8.45	11.26	5.04				0.70907
23 Bear River	78.14	4.49	11.37	5.99	12.1	−11.5	−4.00	
25 Bear River	81.31	4.98	6.54	7.17	11.3	−11.9	−5.40	
27 Bear River	74.97	5.57	16.03	3.43				0.71067
29 Bear River	84.98	3.07	9.56	2.39	13.2	−12.8	−5.70	
31 Bear River	76.94	5.53	16.60	4.15				0.71078
33 Bear River	85.29	5.19	8.76	2.98	4.5	−12.2	−5.00	
35 Bear River	80.46	6.11	12.34	8.11	11.6	−11.4	−5.90	
37 Bear River	82.73	5.96	9.95	9.08				
39 Bear River	84.72	5.79	8.48	4.39	8.0	−10.0	−5.00	
41 Bear River	76.95	7.39	14.52	5.92	12.9	−10.1	−5.30	
43 Bear River	70.16	8.89	19.74	10.20	22.2	−9.3	−3.50	
<i>B.R. in Uintas</i>								
45 Bear River	71.62	5.34	21.60	11.58				
51 Bear River	83.14	0.00	0.00	16.86				
53 Bear River	87.76	0.00	0.00	12.24				0.71910

See Fig. 3 for sample locations.

XRD percentages were calculated from peak intensities.

Isotope data are on the carbonate fraction.

Thompson, 1987; Currey, 1990). At that time, Bristlecone Pine occupied the present Piñon/Juniper life zone implying that the climate was somewhat wetter, and summer temperatures were as much as 10° below modern. The increased moisture was due to increased winter precipitation produced by enhanced cyclonic flow (i.e., increased domination of the Aleutian lows) (Kutzbach et al., 1993). This change in circulation apparently was caused by a splitting of the polar jet stream, and southerly displacement of the southern branch of the jet (Kutzbach, 1987; COHMAP Members, 1988). Precipitation is greatest under the axis of the polar jet, decreases

abruptly south of the axis, and decreases less abruptly north of the axis. Consequently, the Great Basin was wetter during the LGI and the Pacific Northwest was drier (Barnowsky et al., 1987; Thompson et al., 1993). This precipitation effect would have been enhanced by the temperature contrast between the Laurentide ice sheet to the north and the relatively warm sea-surface temperatures along the Pacific coast that likely intensified the jet. Sediments deposited during the LGI in Owens Lake in southeastern California at a latitude of 36.5° N contain an ostracode assemblage dominated by *Cytherissa lacustris*, which indicates very cold, stable

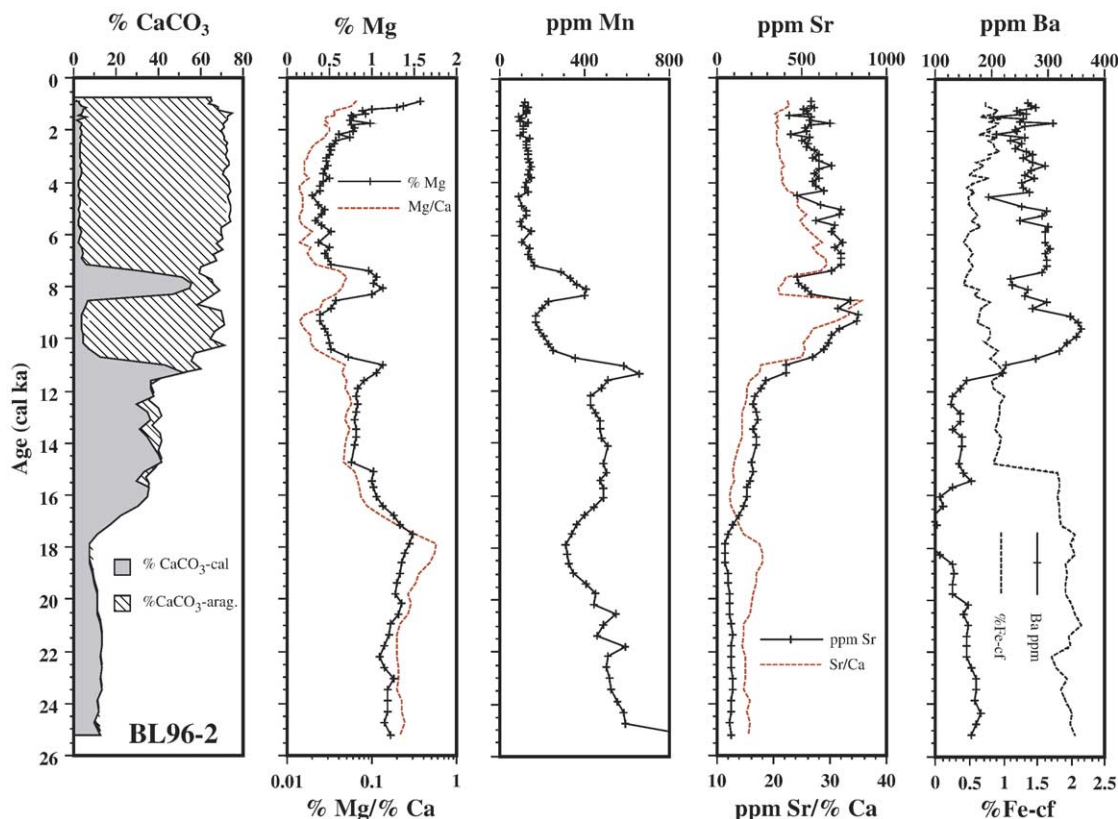


Fig. 11. Profiles of percent CaCO_3 as aragonite and calcite, and HCl-soluble % Mg, Mg/Ca ratio, ppm Mn, ppm Sr, Sr/Ca ratio, ppm Ba, and % Fe on a carbonate-free (cf) basis in bulk carbonate versus age in core 96-2.

limnoclimatic conditions (Bradbury and Forester, 2002). Based on the modern distribution of *C. lacustris*, its presence in large numbers implies that polar air masses were present year-round at the latitude of Owens Lake. It is difficult to determine how much wetter the LGI was, but modeling studies (Hostetler et al., 1994) indicate that the presence of the lakes had a considerable influence on the precipitation over these large lake basins. This study estimates that at 20 cal ka, precipitation in the Bonneville basin with the lake was 3% greater in January and 38% greater in July than it was without the lake due to lake-atmosphere feedbacks.

Lakes in the southern Great Basin reached their maximum levels during the LGI. Pluvial Lakes Manix and Mojave were overflowing into pluvial Lake Manly (Death Valley) as early as 24 cal ka until about 17 cal ka (Enzel et al., 2003; Wells et al., 2003). During the LGI, Lake Estancia in central New Mexico reached its highest and most persistent stand (Allen and Anderson, 1993, 2000; Anderson et al., 2002), and a lake occupied the San Agustin Plains of west-central New Mexico (Markgraf et al., 1984). Also during the LGI, major wetlands occupied the now dry basins in southern

Nevada (Quade et al., 2003 and references therein). Because precipitation decreases rapidly south of the polar jet, the presence of persistent lakes and wetlands in the southern Great Basin suggests that the storm tracks that delivered this moisture must have been at latitudes between 32° and 35° N (e.g., Enzel et al., 2003).

Although the levels of the large lakes in the northern Great Basin continued to increase during the LGI, they did not reach their maximum levels until after 18 cal ka, followed by precipitous decreases (Benson and Thompson, 1987; Thompson et al., 1986, 1993; Currey, 1990; Oviatt et al., 1992). This suggests that the moisture-delivering storm tracks migrated northward at about 17 cal ka, when the pluvial lakes of the southern Great Basin began to dry up, and Lakes Bonneville and Lahontan in the northern Great Basin approached maximum levels.

4.2. Evolution of Bear Lake

Even at its highest elevation (1620 masl), Lake Bonneville was still more than 100 m below the level

of Bear Lake (ca. 1805 masl today). However, the increased precipitation that caused the expansion of Bonneville, together with the lake-effect increase in precipitation due to Bonneville itself (Hostetler et al., 1994), undoubtedly increased the flow of the Bear River so that the river and lake were connected. The most likely place for the two to be connected is at the north end where the present threshold of Bear Lake is only about 2 m above modern lake level (Laabs and Kaufman, 2003). This connection lasted until about 18 cal ka when the quartz content began to decrease from 80% to 30% over about a 2000-year period, and the CaCO_3 content, as calcite, increased from about 10% to 40% (Unit 2, Figs. 5 and 8).

The sediments in Bear Lake deposited prior to about 18 cal ka were dominated by detrital siliciclastics with minor amounts of low-Mg calcite and dolomite. Some of the low-Mg calcite might be endogenic but we have no way of knowing for sure. These sediments are hematite-rich (high values of HIRM; Fig. 7) indicating that they were derived mainly from glacial flour in the headwaters of Bear River in the Uinta Mountains (Fig. 6; Rosenbaum, 2005). We suggested that $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in sediments deposited before 18 cal ka, which are intermediate between those of stream sediments and west-side creeks (>0.71 ; Table 2) and that of Bear River (0.7088; Fig. 9), might be due to a mixture of detrital and endogenic carbonate. Such a mixture also is indicated by the values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in sediments deposited before 18 cal ka. Values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in low-Mg calcite in the red, quartz-rich LGI sediments (-9‰ and -3‰) are much lower than those of Holocene aragonite (-4‰ and $+3\text{‰}$; Figs. 9 and 10). Values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in the silt plus clay fraction of west-side stream sediments average -14.2 and -6.1 ; those in the fine fraction of east-side stream sediments average -13.8 and -5.8 ; and those in the fine fraction of Bear River sediments average -12.7 and -5.7 (Table 2). Some of these stream sediments contain substantial amounts of low-Mg calcite and (or) dolomite (Table 2). In particular, the west-side stream sediments contain up to 55% dolomite (Table 2) due to the abundance of Paleozoic carbonate rocks, mostly dolomites, in the Bear River Range to the west of the lake. All of the values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in detrital carbonate in stream sediments (Table 2) are much lower than those of Unit 1 sediments (Fig. 9), suggesting that the carbonate in Unit 1 sediments is a mixture of detrital and ^{18}O - and ^{13}C -enriched endogenic carbonate.

Values of $\delta^{13}\text{C}$ increased steadily after 22 cal ka and into the Holocene (Fig. 9), likely due to increased lake productivity and burial of ^{13}C -depleted OC (e.g.,

McKenzie, 1985), as indicated by parallel increases in OC concentration and MAR (Fig. 8). A combination of increasing water temperature, increased evaporation, and higher organic productivity likely produced an increase in the precipitation of CaCO_3 as calcite between 18 and 16 cal ka (Unit 2, Figs. 5 and 8).

As the concentration of CaCO_3 in the sediments increased to 40% between 18 and 16 cal ka, values of $\delta^{18}\text{O}$ in bulk carbonate declined to -11‰ , the lowest values in the entire glacial–Holocene sequence (Fig. 9). These lower values might be due to greater influx of Bear River water, but the influx of detrital clastic material was decreasing at that time (decrease in quartz; Fig. 5), and there are no corresponding decreases in values of $\delta^{13}\text{C}$ or $^{87}\text{Sr}/^{86}\text{Sr}$ (Fig. 9). In fact, the interval between 18 and 16 cal ka is the only interval in the core where there is no covariation between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ (Fig. 10A). Also, this is the time interval when the levels of Lakes Bonneville and Lahontan were rapidly decreasing, indicating that the Pacific moisture source had been cut off. The initial decline of Lake Bonneville to the Provo level at 17.5 cal ka was due to the catastrophic Bonneville flood down the Snake River, but the decline from the Provo level after about 16 cal ka was due to the warmer, drier climate (Oviatt, 1997). The low values of $\delta^{18}\text{O}$ in calcite deposited between 16 and 12 cal ka are most likely due to continued abundance of cold-season precipitation (winter snow), particularly in the Uinta Mountains, the headwaters of Bear River. There is a positive covariation between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in this interval, but values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ each vary by only about 1‰ (Fig. 10A). We have little specific information on the deglacial history of the Uinta Mountains. Based on recessional moraine evidence, Munroe (2003) concluded that the glacier that occupied the drainage of the Henrys Fork of the Green River in the northern Uinta Mountains was present below 3300 m until about 10 cal ka. Woodrat midden pollen studies at Dutch John Mountain on the north-eastern end of the Uinta Mountains by Jackson et al. (in press) suggest that late-glacial conditions were wetter than during the full glacial period. In general, alpine glaciers in the ranges of the Basin and Range were at their maximum extent at 20 cal ka (Thompson et al., 1993), and persisted until about 14 cal ka (Thompson, 1990). Therefore, it is reasonable to assume that the Bear River received the bulk of its water from snow melt during the LGI, at least until about 12 cal ka, and that some of that water continued to enter Bear Lake even though the sediment supply, indicated by quartz (Fig. 5), was greatly reduced after 16 cal ka.

By 12 cal ka, values of both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$, but especially $\delta^{18}\text{O}$, began to increase rapidly (Fig. 9), most likely due to increased evaporation. By about 11 cal ka, a salinity threshold was reached and aragonite began to precipitate (Unit 3–4 boundary, Fig. 5). In the marine environment, aragonite formation relative to calcite is favored by warmer temperatures and higher salinity (e.g., Burton and Walter, 1987; Morse and Mackenzie, 1990), which is why aragonite is only found in warm, shallow marine environments. Aragonite formation in a large, cold, oligotrophic, high latitude, montane lake is highly unusual. Lacustrine aragonite usually is found in small saline lakes in which the salinity varies considerably over time. However, Bear Lake contains endemic ostracodes and four or five endemic species of fish (Wurtsbaugh and Luecke, 1997), which indicate that the chemistry of the lake has remained fairly constant for a long time. The production of large quantities of aragonite under an evaporative climate indicates that snow-pack-sourced ground-water discharge remained high during warm evaporative summers. Bear Lake hydrology combined with evaporation created an unusual situation that produced large amounts of aragonite, but no evaporite minerals (e.g., gypsum and halite). The thick sedimentary package in the lake also argues for relative stability. Acoustic reflection profiles show that the deepest part of the lake contains at least several hundred meters of sediment that might be as much as six million years old (Colman, 2005). A 120-m-long core was collected from the deepest part of the lake with the GLAD800 system in 2000 (BL00-1, Fig. 2; Dean et al., 2002a). The dominant carbonate mineral in this core is calcite, but aragonite occurs in sediments deposited during the last two interglacial intervals (marine oxygen isotope stages 5 and 7; Dean, 2005; Bright et al., in press).

The key variable for aragonite formation appears to be a high Mg:Ca ratio (Müller et al., 1972), and an increase in Mg:Ca ratio usually is accompanied by an increase in salinity (Kelts and Hsü, 1978). According to the model of Harvie and Weare (1980) the predicted sequence of carbonate minerals formed with increasing Mg:Ca ratio should be low-Mg calcite (<5 mol% Mg), high-Mg calcite (>10 mol% Mg), and then aragonite. Müller et al. (1972) suggested that low-Mg calcite forms in lakes with a Mg:Ca ratio <2, high-Mg calcite forms in lakes with a Mg:Ca ratio of 2–12, and aragonite only precipitates in lakes with a Mg:Ca ratio >12. The presence of dissolved Mg strongly inhibits the precipitation of calcite and favors the precipitation of aragonite (e.g., Bischoff and Fyfe, 1968; Berner, 1975).

The diversion of Bear River into Bear Lake at the beginning of the 20th century changed the chemistry of Bear Lake and created an experiment in carbonate chemistry that is reflected in carbonate minerals in the sediments (Dean et al., 2005). Today, the total dissolved solids (TDS) content and the Mg:Ca ratio (in mg/L) are about 500 and 1.7:1 (Dean et al., 2005), and the lake is precipitating high-Mg calcite. The TDS content and Mg:Ca ratio in a sample of Bear Lake water collected about the time of diversion in 1912 were 1100 and 37:1 (Kemmerer et al., 1923). Thus, at about the time of Bear River diversion, the salinity of Bear Lake was about twice that of today, but the Mg:Ca ratio was more than 20 times higher. Therefore, prior to diversion aragonite was the preferred polymorph of CaCO_3 .

Because the more open crystal lattice of aragonite can more readily accommodate substitution of larger Sr^{2+} and Ba^{2+} ions for the smaller Ca^{2+} ions than the less open lattice of calcite (Morse and Mackenzie, 1990), Ba and Sr concentrations and the Sr:Ca ratio in the leachate also increased with aragonite precipitation (Fig. 11). The Mg^{2+} ion is smaller than the Ca^{2+} ion and can easily substitute for Ca^{2+} in the calcite lattice. Calcite in Bear Lake sediments usually contains more Mg than aragonite (Fig. 11), but that also depends on the Mg:Ca ratio of the water. The three-fold increase in Mg in aragonite over the last 4000 years, to values that are as high or higher than those in calcite deposited between 17 and 11 cal ka (Fig. 11), probably is due to a three-fold increase in the Mg:Ca ratio in the late Holocene lake (Fig. 11). The Mn^{2+} ion also is smaller than the Ca^{2+} ion and can easily explain the higher Mn content in calcite than in aragonite (Fig. 11). Manganese carbonates (in fact all transition-element carbonates) are isostructural with calcite, whereas strontium carbonate is isostructural with aragonite (Morse and Mackenzie, 1990).

By 10 cal ka, values of $\delta^{18}\text{O}$ had increased by 6‰ from -11‰ to -5‰ , and values of $\delta^{13}\text{C}$ had increased from -1‰ to $+2.5\text{‰}$. Ratios of $^{87}\text{Sr}/^{86}\text{Sr}$ also increased during this interval (12 to 10 cal ka; Fig. 9). At that time (10 cal ka), the mineralogy and isotopic composition of bulk carbonate had approached values that would be typical of most of the Holocene (Figs. 5 and 9). A salinity increase, indicated by the higher aragonite concentrations, was accompanied by a drop in lake level by more than 25 m below modern level (Smoot and Rosenbaum, 2005). The marked increases in values of both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ (Fig. 9) and a distinct isotopic covariance in carbonates deposited after 12 cal ka (Fig. 10A) indicate increased residence time of the lake (Talbot, 1990).

Just after 9.5 cal ka, values of $^{87}\text{Sr}/^{86}\text{Sr}$ begin to decrease indicating that Bear River water was entering Bear Lake, and by 9.0 cal ka decreasing values of $\delta^{18}\text{O}$ show that the lake freshened rapidly (Fig. 9). Because the Sr concentration in the lake was so low, the effect of change on the isotopic composition of Sr would have been felt immediately, but it took a while longer for the change to be felt in the much larger oxygen–isotope system. During this freshening episode, the Mg/Ca ratio became smaller, but it did not become low enough for the mineralogy to shift from aragonite to calcite until after 8.5 cal ka (Fig. 5). This change in mineralogy is reflected in the HCl-leach chemistry by increases in Mg and Mn and decreases in Sr and Ba (Fig. 11). At about 8 cal ka, values of $^{87}\text{Sr}/^{86}\text{Sr}$ began to increase indicating that Bear Lake was beginning to detach from Bear River (Fig. 9). This detachment was not felt by the carbon and oxygen isotopes until about 7.5 cal ka, and by that time calcite was rapidly being replaced by aragonite.

The cause of the brief reconnection of Bear River with Bear Lake may have been tectonic, geomorphic, or climatic. Evidence from Elk Lake, Minnesota, suggests that the polar front in summer was south of northern Minnesota prior to about 8.3 cal ka, and was displaced northward at about 8.2 ka, permitting the introduction of strong westerly winds (Dean et al., 2002b). Although the displacement of the polar front and beginning of westerlies was at 8.2 cal ka, there was about a 500-year period centered on 9.5 cal ka when there is evidence (increased varve thickness and quartz content) that winds began to increase, but then the winds decreased, and the period between 9.2 and 8.2 cal ka contains some of the thinnest varves in the Elk Lake record (Dean et al., 2002b). This 1000-year interval is when *Cytherissa lacustris*, a boreal ostracode, was most abundant. *C. lacustris* lives in cold lakes in the boreal forests of Canada and Alaska, and cannot tolerate physical and chemical environmental variability (Dean et al., 2002b). This atmospheric perturbation may have permitted the temporary establishment of a polar low over Utah, Idaho, and Wyoming in winter, increasing the snow pack in the Uinta Mountains, the headwaters of the Bear River. Such a climatic event also would have increased the snow pack in the Bear River Range to the west of the lake thereby increasing the surface- and ground-water flow to the west side of the lake. The interpretation, then, is that when *C. lacustris* was at its greatest abundance in Elk Lake indicating the southernmost extent of the polar front in the early Holocene, increased snow melt in the Uinta Mountains and the Bear River Range increased the influx of surface and ground water into Bear Lake between 9.2 and 8.2 cal ka

(decrease in $^{87}\text{Sr}/^{86}\text{Sr}$). This resulted in a rise in lake level of about 10 m above modern level (Smoot and Rosenbaum, 2005). At 8.2 cal ka the abundance of *C. lacustris* in Elk Lake was greatly reduced indicating that the polar front had moved north and was replaced by the prairie ostracodes *Candona rawsoni* and *Limnocythere herricki*, marking the beginning of the altithermal or prairie period in Minnesota (Dean et al., 2002b). At that time, surface- and ground-water influx to Bear Lake decreased and the salinity of the lake began to increase (increase in $^{87}\text{Sr}/^{86}\text{Sr}$).

The increase in supply of water to Bear Lake, whatever its source, and associated lake-level rise probably corresponds to the Willis Ranch shoreline (Laabs and Kaufman, 2003). This shoreline is at an elevation of about 1814 masl (8 m above modern lake level) indicating that at that time Bear Lake occupied most of the Bear Lake Valley. The Willis Ranch shoreline was calibrated at 9.2 cal ka (Laabs and Kaufman, 2003), but this date is based on a conventional ^{14}C analysis of a large collection of shells (8.27 ^{14}C ka; Williams et al., 1962) and is probably too old, as suggested by the reservoir effect of 370 years used by Colman et al. (2005) to correct ostracode ages. With the reentry of Bear River, there is no increase in quartz (Fig. 5) to indicate that Bear River sediment also was entering the lake, so the sediment must have been trapped. Topographically, the most obvious place for Bear River to enter the lake is at the north end where it enters today through man-made canals. Perhaps by that time a marsh-lake complex, today represented by Dingle Marsh and Mud Lake (Fig. 1), had developed at the north end of the lake trapping Bear River sediment.

The lower salinity event lasted about 1000 years, and by 7.5 cal ka aragonite was again forming and values of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of carbonate were increasing. The mineralogy of carbonate in Bear Lake sediments has been relatively constant for the last 7 ky (Fig. 5), but values of $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, and $^{87}\text{Sr}/^{86}\text{Sr}$ continued to increase slightly (Fig. 9), indicating that perhaps salinity continued to increase after Bear River was again disconnected from Bear Lake at 7 cal ka, and evaporation dominated over precipitation. However, Bear Lake waters never reached highly evolved values of $\delta^{18}\text{O}$ (>0‰) in spite of an evaporation-dominated water balance.

5. Conclusions

A composite section from three piston cores from 30 to 50 m water depth, each up to 5 m long, was used to reconstruct the history of the Bear Lake for the last 31,000 years. This composite section reveals the com-

ing and going of the Bear River, and a complex carbonate mineralogy involving aragonite, low-Mg calcite, and dolomite.

During the last glacial interval, increased Pacific moisture that filled the large lake basins of the Great Basin region, provided sufficient glaciation, snow pack, and snow melt in the Uinta Mountains and the Bear River Range to connect Bear Lake with Bear River. Sediments deposited during the last glacial interval consist of detrital quartz and feldspar transported by the Bear River, with minor amounts of low-Mg calcite and dolomite. The quartz and feldspar are detrital, and most, if not all, of the low-Mg calcite and dolomite also are probably detrital, derived mainly from Paleozoic carbonate rocks in the Bear River Range to the west of the lake.

Between 18 and 16 cal ka, Bear River abandoned Bear Lake and endogenic carbonate, first as calcite, began to precipitate as air temperature increased, evaporation exceeded freshwater influx, and the salinity of the lake increased. The oxygen and carbon isotopic ratios and organic carbon content of the endogenic carbonate show that after 16 ka the salinity continued to increase, the water became warmer, and organic productivity increased. Because of this combination of environmental events, a threshold was reached at about 11 cal ka when carbonate mineralogy abruptly changed from calcite to aragonite. Aragonite formation in a large, deep, montane lake is highly unusual. The production of large quantities of aragonite under an evaporative climate indicates that ground-water discharge remained high during warm evaporative summers. Bear Lake hydrology combined with evaporation created an unusual situation that produced large amounts of aragonite, but no evaporite minerals.

Aragonite precipitation dominated the sediments for about 2 ky, then, just as abruptly, returned to calcite precipitation for another 1 ky, between 8.5 cal ka and 7.5 cal ka. This change in carbonate deposition probably was caused by reentry of the Bear River into Bear Lake, as suggested by more depleted oxygen, carbon, and strontium isotopic ratios.

By 7 cal ka, aragonite again dominated the sediments. This mode of carbonate sedimentation continued for the rest of the Holocene. Although the amount of aragonite did not increase during the last 7 ky, isotopic ratios of oxygen, carbon, and strontium continued to increase indicating that salinity continued to increase.

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