# Endogenic carbonate sedimentation in Bear Lake, Utah and Idaho, over the last two glacial-interglacial cycles

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## ABSTRACT

Sediments deposited over the past 220,000 years in Bear Lake, Utah and Idaho, are predominantly calcareous silty clay, with calcite as the dominant carbonate mineral. The abundance of siliciclastic sediment indicates that the Bear River usually was connected to Bear Lake. However, three marl intervals containing more than 50% $CaCO_3$ were deposited during the Holocene and the last two interglacial intervals, equivalent to marine oxygen isotope stages (MIS) 5 and 7, indicating times when the Bear River was not connected to the lake. Aragonite is the dominant mineral in two of these three high-carbonate intervals. The high-carbonate, aragonitic intervals coincide with warm interglacial continental climates and warm Pacific sea-surface temperatures. Aragonite also is the dominant mineral in a carbonate-cemented microbialite mound that formed in the southwestern part of the lake over the last several thousand years. The history of carbonate sedimentation in Bear Lake is documented through the study of isotopic ratios of oxygen, carbon, and strontium, organic carbon content, $CaCO_3$ content, X-ray diffraction mineralogy, and HCl-leach chemistry on samples from sediment traps, gravity cores, piston cores, drill cores, and microbialites.

Sediment-trap studies show that the carbonate mineral that precipitates in the surface waters of the lake today is high-Mg calcite. The lake began to precipitate high-Mg calcite sometime in the mid–twentieth century after the artificial diversion of Bear River into Bear Lake that began in 1911. This diversion drastically reduced the salinity and $Mg^{2+}:Ca^{2+}$ of the lake water and changed the primary carbonate precipitate from aragonite to high-Mg calcite. However, sediment-trap and core studies show that aragonite is the dominant mineral accumulating on the lake floor today, even though it is not precipitating in surface waters. The isotopic studies show that this aragonite is derived from reworking and redistribution of shallow-water sediment that is at least 50 yr old, and probably older. Apparently, the microbialite mound also stopped forming aragonite cement sometime after Bear River diversion. Because of reworking of old aragonite, the bulk mineralogy of carbonate in bottom sediments has not changed very much since the diversion. However, the diversion is marked by very distinct changes in the chemical and isotopic composition of the bulk carbonate.

After the last glacial interval (LGI), a large amount of endogenic carbonate began to precipitate in Bear Lake when the Pacific moisture that filled the large pluvial lakes of the Great Basin during the LGI diminished, and Bear River apparently abandoned

Bear Lake. At first, the carbonate that formed was low-Mg calcite, but ~11,000 years ago, salinity and $Mg^{2+}$:$Ca^{2+}$ thresholds must have been crossed because the amount of aragonite gradually increased. Aragonite is the dominant carbonate mineral that has accumulated in the lake for the past 7000 years, with the addition of high-Mg calcite after the diversion of Bear River into the lake at the beginning of the twentieth century.

## INTRODUCTION

Bear Lake occupies the southern half of the Bear Lake Valley in northeastern Utah and adjacent Idaho (Fig. 1). The lake is 32 km long and 6–13 km wide with an area of 280 $km^2$ at full capacity. Maximum depth is 63 m, and mean depth is 28 m (Birdsey, 1989). The present elevation of the lake when full is 1805 m above sea level, but this level has varied considerably through time (Reheis et al., this volume; Smoot and Rosenbaum, this volume). The natural watershed of the lake is quite small and has a basin-area:lake-area ratio of 4.8 (Wurtsbaugh and Luecke, 1997). Within historical times, the Bear River did not flow into Bear Lake. A series of canals was built beginning in 1911 to bring Bear River water into the lake and a pumping station was completed in 1918 to return Bear Lake water to the river (Birdsey, 1989). Apparently the first Bear River water was diverted into Mud Lake, and presumably overflowed into Bear Lake, in May 1911 (Mitch Poulsen, Bear Lake Regional Commission, 2004, personal commun.). Judging from sparse records, large volumes of Bear River water were not diverted into Bear Lake until at least 1913 (Connely Baldwin, PacifiCorp, 2005, personal commun.). The oldest chemical analysis from Bear Lake is of a water sample collected in 1912 (Kemmerer et al., 1923). I assume that the 1912 analysis is representative of the pre-diversion composition of the lake (Dean et al., 2007). Diversion of Bear River into Bear Lake created a reservoir to supply hydropower and irrigation water. This increased the basin-area:lake-area ratio considerably, to 29.5 (Birdsey, 1989). The mean annual surface hydrologic flux (including precipitation) to the lake is estimated to be $0.48 \times 10^9$ $m^3yr^{-1}$ (Lamarra et al., 1986). Outflow is estimated as $0.214 \times 10^9$ $m^3yr^{-1}$ which is ~3% of the lake volume ($7.86 \times 10^9$ $m^3$), giving an average residence time of ~37 yr. The magnitude of groundwater influx may be considerable, but is not known. However, Bright (this volume) concluded that the annual hydrologic balance of the lake is approximately zero. The topographic catchment of Bear Lake is small so that sustaining inflow during persistent dry climate intervals implies that there were (are) large extrabasinal sources of water, necessarily groundwater. The presence of extrabasinal sources of groundwater flow in this setting requires fracture flow along major faults in the highly faulted Bear Lake Valley (Colman et al., 2006; Reheis et al., this volume), and also flow through the highly karstic carbonate rocks of the Bear River Range. The supply of groundwater is primarily from snowmelt (Dean et al., 2007; Bright, this volume).

Acoustic profiling reveals a continuous layered sediment package that is at least 100 m thick and should contain records of at least two or more older glacial cycles (Colman, 2006; Kauf-

man et al., this volume). Such records are out of range of conventional lake coring systems. However, with the development of the Global Lake Drilling to 800 m (GLAD800) system (Dean, et al., 2002), lake cores up to 800 m long are theoretically possible. Preliminary testing of the GLAD800 system was conducted on Bear Lake in September 2000, and two closely spaced cores (holes BL00-1D and BL00-1E) up to 120 m long were collected at site BL00-1 (Fig. 1A; http://www.dosecc.org/html/utah_lakes.html).

The purposes of this paper are to examine the deposition of $CaCO_3$ in Bear Lake on five time scales. First, I examine modern $CaCO_3$ deposition based on evidence from sediment traps. Next, I examine the effect of the diversion of Bear River into Bear Lake on the chemistry, mineralogy, and isotopic composition of $CaCO_3$ deposited in the lake during the twentieth century. The "Rock Pile" (a microbial mound) at the southwest end of the lake documents an unusual development in space and time (past 2000 years) in Bear Lake's history. Then, I examine the initiation of $CaCO_3$ deposition in the lake following the last glacial interval, and changes in the chemistry, mineralogy, and isotopic composition of endogenic carbonate deposited during the latest Pleistocene and the Holocene. Finally, I document the changes in the amount and mineralogy of carbonate deposited in the lake over the last two glacial-interglacial cycles (220,000 cal yr).

Some of this material has been published elsewhere (Dean et al., 2006, 2007) but will be summarized here for a complete overview of all aspects of endogenic carbonate deposition in Bear Lake. However, much of the material is new, and the entire package emphasizes the diversity of the forms and environments of endogenic carbonate deposition that has occurred in Bear Lake over the past 220,000 years.

## MATERIALS AND METHODS

Piston cores were collected in 1996 from three localities in Bear Lake (Fig. 1A) using the Kullenberg coring system of the University of Minnesota's Limnological Research Center (UMN-LRC) (Dean et al., 2006). The piston cores recovered a maximum of 5 m of sediment each, but overlapping cores provide a 26,000 yr record. Unknown amounts of sediments were missing from the tops of the piston cores, so surface sediments (up to 50 cm) were collected with a gravity corer in 1998 (Fig. 1A; Dean et al., 2007). Piston cores and gravity cores were collected at five localities at the northern end of the lake in 2002 (Fig. 1A). To assess the seasonality of sedimentation, time-marking sediment traps that dispense Teflon granules in the collection tube every 30 days (Anderson, 1977) were suspended 10 m below the surface of the lake (referred to as surface traps) and 2 m above
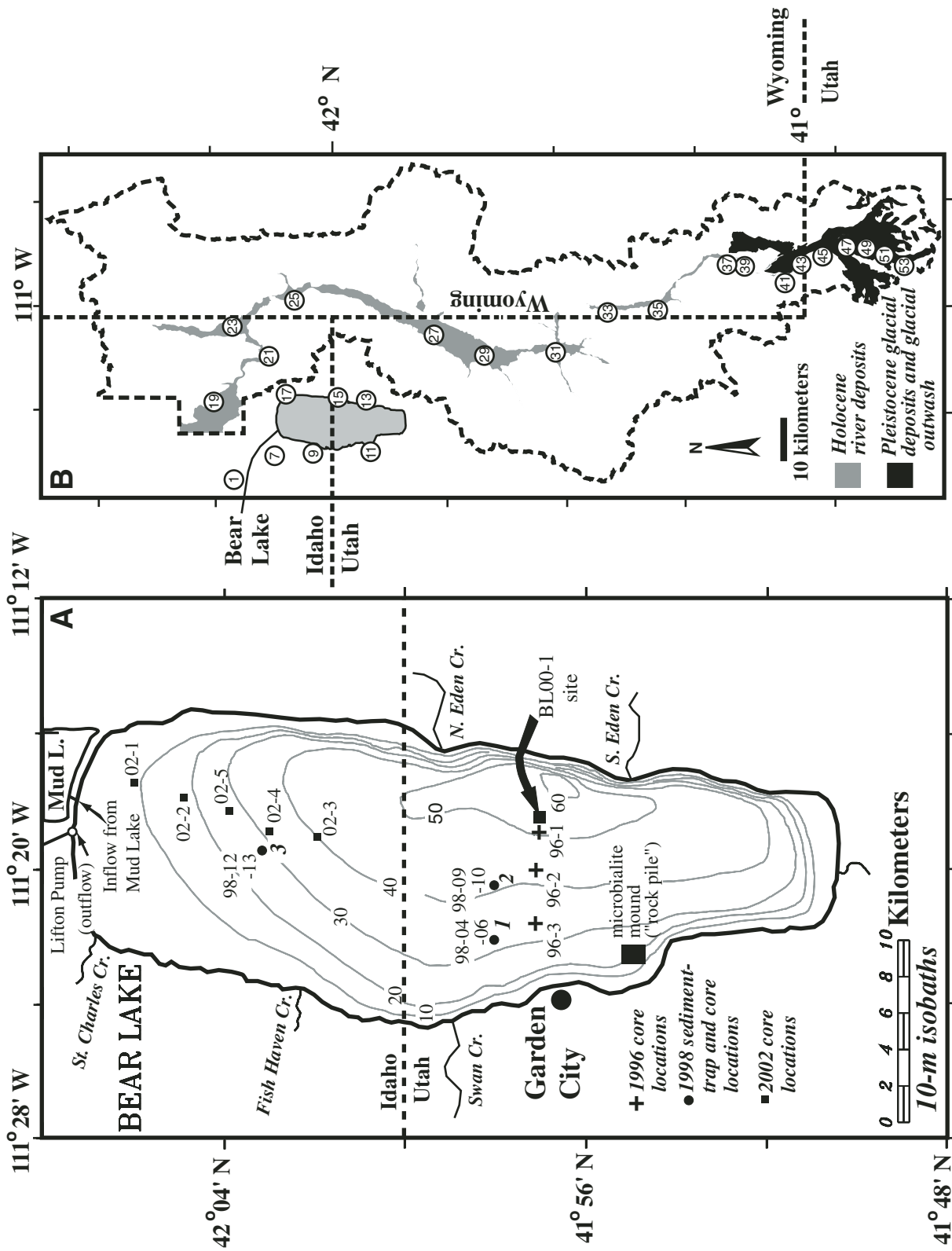
Figure 1. (A) Bathymetric map of Bear Lake, Utah and Idaho, showing the locations of 1996 piston cores, 1998 surface-sediment cores and sediment-trap deployments, 2002 piston and surface-sediment cores, and the microbialite mound ("rock pile"). (B) Locations of stream-sediment samples. Sample numbers in circles correspond to sample numbers in Table 3. Dashed line corresponds to the catchment of the upper reach of the Bear River to its headwaters in the Uinta Mountains. 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. See Rosenbaum et al. (this volume) for mineralogy, geochemistry, and magnetic properties of these stream-sediment samples.

the bottom of the lake (bottom traps) at three localities where multiple gravity cores were collected (Fig. 1A) for up to three years (1998–2000).

The GLAD800 cores (from holes BL00-1D and BL00-1E) were shipped under refrigeration to the UMN-LRC. There, the cores were scanned with a Geotek multisensor logger (MSL) that measures porosity, wet bulk density (WBD), magnetic susceptibility (MS), and P-wave velocity. The correlation of sediment recovered in holes BL00-1D and -1E is based on MSL-MS profiles (Colman et al., 2006; Heil et al., this volume). Sampling and analyses followed core curation specifications and protocols that were in place before drilling. The photographs and initial core descriptions (ICD) are available at the National Oceanic and Atmospheric Administration/National Geophysical Data Center, Boulder, Colorado (http://www.ngdc.noaa.gov/mgg/curator/laccore.html). Initial samples for magnetic properties, mineralogy, and geochemistry were taken from the core catcher samples. Additional samples were collected at the same intervals that smear-slide samples were taken at the UMN-LRC in preparing the ICDs.

Concentrations of total carbon (TC) and inorganic carbon (IC) were determined by coulometric titration of $CO_2$ following extraction from the sediment by combustion at 950 °C and acid volatilization, respectively (Engleman et al., 1985), in U.S. Geological Survey (USGS) laboratories, Denver, Colorado. Weight percent (wt%) IC was converted to wt% $CaCO_3$ by dividing by 0.12, the fraction of carbon in $CaCO_3$. Organic carbon (OC) was determined as the difference between TC and IC. The accuracy and precision for both TC and IC, determined from hundreds of replicate standards usually are better than 0.10 wt%.

Semiquantitative estimates of mineral contents were determined by X-ray diffraction (XRD) techniques (e.g., Moore and Reynolds, 1989) in USGS laboratories, Denver, Colorado. For each sample, raw XRD peak intensities for the main peaks of minerals detected in each sample were converted to semiquantitative percentages by dividing the main peak intensity of a mineral by the sum of the main peak intensities of all minerals. More quantitative estimates of aragonite and calcite, the dominant minerals, were calculated by partitioning the percentage of total $CaCO_3$, determined by coulometry, using the intensity ratios of the main XRD peaks of aragonite to calcite (I-aragonite/I-calcite) and curves of percent aragonite (of total $CaCO_3$) versus I-aragonite/I-calcite determined by Chave (1954) and Lowenstam (1954).

Measurements of ratios of stable isotopes of carbon and oxygen were made on aliquots of the carbon samples (see Dean et al., 2006, 2007 for methods). Isotope measurements on samples from the 1996 cores, 1998 cores, and sediment traps were made in the stable isotope laboratory at the University of Minnesota. Isotope measurements on samples from the 2002 cores were made in the stable isotope laboratory at the University of Arizona. Results of analyses are reported in the usual per mil (‰) δ-notation relative to the Vienna Pee Dee Belemnite (VPDB) marine-carbonate standard for carbon and oxygen:

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

where R is the ratio ($^{13}C$:$^{12}C$) or ($^{18}O$:$^{16}O$). Precision is ± 0.1‰ for oxygen and ± 0.06‰ for carbon.

Measurements of dissolved strontium isotope ratios ($^{87}Sr$/$^{86}Sr$) were made on filtered, unacidified water samples and on bulk sediment samples leached in 5 mol $L^{-1}$ acetic acid (see Dean et al., 2007, for locations of water samples and details of methods). Precision is usually ±0.00001 for both waters and sediments.

Gas bubbles in Bear Lake were collected with an inverted funnel and water-filled bottle. The chemical composition of the gas was determined by using a Hewlett Packard 6890 series gas chromatograph in the USGS Organic Geochemistry Laboratory, Denver, Colorado. The isotopic composition of gas samples was measured with a Micromass Optima continuous flow Isotope Ratio Mass Spectrometer coupled with an Agilent 6890 Gas Chromatograph and a combustion furnace in the USGS Organic Geochemistry Laboratory, Denver, Colorado. The precision of the method is ±0.2‰.

Chronologies for the 1998 gravity cores are provided by $^{210}Pb$ dating (Smoak and Swarzenski, 2004). This technique gives approximate ages, at least for the past 100 years. Accelerator mass spectrometer (AMS) $^{14}C$ ages were obtained from samples of various materials including pollen concentrates, ostracodes, and bulk-sediment OC from piston cores (Colman et al., this volume). Reservoir-corrected and calibrated radiocarbon ages are expressed in kiloannum B.P. (e.g., 26.6 cal ka), and time intervals in thousands of calendar years (cal k.y.). The age model for the 120-m-long sediment sequence in BL00-1 is presented by Colman et al. (2006).

## RESULTS AND INTERPRETATION

### Sediment Traps

Summer carbonate sediment collected in all surface traps consists of almost pure high-Mg calcite (~10 mol % Mg) (Table 1, Figs. 2 and 3). The bottom trap at the deep site contained aragonite as the dominant mineral, but also had considerable amounts of high-Mg calcite, low-Mg calcite, and quartz, and minor amounts of dolomite (site 2; Table 1, Figs. 1A and 2). In contrast, high-Mg calcite was the dominant mineral in sediments collected in the bottom trap deployed at 27 m below the surface at the north end of the lake, along with considerable amounts of low-Mg calcite, quartz, and aragonite (site 3; Table 1, Figs. 1A and 3). The aragonite occurs as subrounded, needle-shaped crystals ~5 μm long and <1 μm in diameter (Fig. 4A). Most of the calcite occurs as equant, subrounded rhombohedral grains ~4–5 μm in diameter (Fig. 4B).

Three key questions posed by these observations are (1) Why does unstable high-Mg calcite form in the epilimnion? (2) Does high-Mg calcite dissolve and reprecipitate as low-Mg calcite on its trip through the water column? And (3) what is the origin of the aragonite in the hypolimnion?

To my knowledge, the first mention of either high-Mg calcite or aragonite from a freshwater environment is from Lake Balaton, Hungary (Müller, 1970, 1971). In that lake, high-Mg calcite

TABLE 1.  BEAR LAKE TRAP SAMPLES FROM LAKE CENTER (SITES 1 AND 2) AND NORTH END (SITE 3).

| Site | Surface or Bottom* | Date Deployed | Date Removed | Sample ID | Approximate Date Deposited | XRD Mineralogy[#] | CaCO$_3$ (%) | OC (%) | $^{87}$Sr/$^{86}$Sr | d$^{13}$C (VPDB) (‰) | d$^{18}$O (VPDB) (‰) |
|---|---|---|---|---|---|---|---|---|---|---|---|
| 1 | bottom | 7/14/1998 | | 0 cm | Summer, 1998 | | 63.2 | 2.34 | | | |
| 1 | bottom | 7/14/1998 | 9/18/1999 | 50 cm | Summer, 1998 | A>HMC>Q>LMC>>D | 64.3 | 2.63 | | | |
| 1 | bottom | 7/14/1998 | 9/18/1999 | 50–100 cm | Summer, 1998 | HMC>A>Q>LMC>>D | 68.4 | | 0.7093 | | |
| 2 | surface | 7/14/1998 | 9/18/1999 | 5/7/1999 | 5/7/1999 | HMC>>A-C-Q | 80.9 | 2.34 | 0.7093 | -1.66 | -9.24 |
| 2 | surface | 7/14/1998 | 9/18/1999 | 7/6/1999 | 7/6/1999 | HMC>>A-C-Q | 85.1 | 2.46 | | | |
| 2 | surface | 7/14/1998 | 9/18/1999 | 8/5/1999 | 8/5/1999 | HMC>>A-C-Q | 86.7 | | 0.7092 | | |
| 2 | surface | 7/14/1998 | 9/18/1999 | 9/4/1999 | 9/4/1999 | HMC>>A-C-Q | 87.5 | | | -1.73 | -9.72 |
| 2 | bottom | 7/14/1998 | 9/18/1999 | top | Sept., 1999 | | 58.1 | 5.05 | | | |
| 2 | bottom | 7/14/1998 | 9/18/1999 | 20–30 cm | Summer, 1999 | A>HMC>Q>LMC>>D | 65.8 | | | | |
| 2 | bottom | 7/14/1998 | 9/18/1999 | 30–32 cm | Summer, 1999 | A>Q=HMC>LMC>>D | 65.3 | 2.18 | 0.7094 | 0.36 | -7.94 |
| 2 | bottom | 7/14/1998 | 9/18/1999 | 60 cm | Early Summer, 1999 | HMC>A>Q>LMC | 61.9 | 4.79 | | | |
| 2 | bottom | 7/14/1998 | 9/18/1999 | bot. 10 cm | Late Summer, 1998 | A>>Q>HMC>LMC>>D | 61.8 | 3.84 | | -0.19 | -7.92 |
| 2 | bottom | 9/22/1999 | 9/10/2000 | 1 | Sept., 2000 | | 69.7 | 1.99 | | | |
| 2 | bottom | 9/22/1999 | 9/10/2000 | 2 | Summer, 2000 | | 63.3 | 2.40 | | | |
| 2 | bottom | 9/22/1999 | 9/10/2000 | 3 | Summer, 2000 | A>HMC>Q>LMC>>D | | | 0.7094 | | |
| 2 | bottom | 9/22/1999 | 9/10/2000 | 4 | Spring, 2000 | | 66.9 | 1.89 | | | |
| 2 | bottom | 9/22/1999 | 9/10/2000 | 5 | Summer, 1999 | | 66.8 | 1.91 | | | |
| 2 | bottom | 9/22/1999 | 9/10/2000 | 6 | Summer, 1999 | A>HMC>Q>LMC>>D | 67.2 | 1.88 | | | |
| 2 | bottom | 9/22/1999 | 9/10/2000 | 7 | Summer, 1999 | | 69.1 | 1.81 | | | |
| 2 | bottom | 9/22/1999 | 9/10/2000 | 8 | | | 68.9 | 1.85 | | | |
| 2 | bottom | 9/22/1999 | 9/10/2000 | 9 | | HMC>A>Q>LMC>>D | 71.5 | 1.74 | | | |
| 2 | bottom | 9/22/1999 | 9/10/2000 | 10 | Summer, 1999 | A=HMC>Q>LMC>>D | | | 0.7094 | | |
| 2 | bottom | 9/22/1999 | 9/10/2000 | 11 | Sept.,1999 | | | | | | |
| 3 | surface | 7/14/1998 | 9/8/2000 | 1 | Summer, 2000 | all hi-Mg cal | | | 0.7092 | -0.50 | -8.73 |
| 3 | surface | 7/14/1998 | 9/8/2000 | 2 | Summer, 2000 | | 74.9 | 2.53 | | | |
| 3 | surface | 7/14/1998 | 9/8/2000 | 3 | Summer, 2000 | HMC>>A>Q | 69.3 | 3.12 | | | |
| 3 | surface | 7/14/1998 | 9/8/2000 | 4 | Spring, 2000 | all hi-Mg cal | | | 0.7093 | -1.26 | -9.24 |
| 3 | surface | 7/14/1998 | 9/8/2000 | 5 | Summer, 1999 | | 84.5 | 1.67 | | | |
| 3 | surface | 7/14/1998 | 9/8/2000 | 6 | Summer, 1999 | HMC>>A=Q | 82.4 | | | | |
| 3 | surface | 7/14/1998 | 9/8/2000 | 7 | Summer, 1999 | all hi-Mg cal | | | 0.7093 | -0.83 | -8.36 |
| 3 | surface | 7/14/1998 | 9/8/2000 | 8 | Summer, 1998 | | 83.7 | 1.72 | | | |
| 3 | surface | 7/14/1998 | 9/8/2000 | 9 | Summer, 1998 | HMC>>Q>A | 78.4 | 2.20 | | | |
| 3 | surface | 7/14/1998 | 9/8/2000 | 10 | July, 1998 | all hi-Mg cal | | | 0.7093 | -0.38 | -7.87 |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 1 | Late Summer, 2000 | | 64.3 | 1.79 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 2 | Summer, 2000 | | 63.4 | 2.03 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 3 | Summer 2000 | HMC>A=Q>LMC>>D | 64.5 | | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 4 | Summer, 2000 | | 66.0 | 1.72 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 5 | Summer, 2000 | | 65.1 | 1.83 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 6 | Summer, 2000 | | 61.8 | 2.27 | 0.7094 | -0.27 | -8.02 |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 7 | Spring, 2000 | | 65.1 | 1.97 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 8 | Dec.,1999 | HMC>A>Q>LMC>>D | | | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 9 | Nov-Dec., 1999 | | 61.3 | 2.10 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 10 | Nov., 1999 | HMC>A>LMC>Q>>D | 65.1 | 2.10 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 11 | Oct.-Nov., 1999 | | 65.8 | 2.04 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 12 | Oct.? 1999 | | 66.3 | 1.59 | 0.7094 | -0.50 | -8.12 |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 13 | Sept., 1999 | | 67.1 | 2.36 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 14 | Aug., 1999 | | 63.0 | 2.35 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 15 | July, 1999 | | 60.9 | 1.49 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 16 | July, 1999 | | 71.3 | 1.08 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 17 | June, 1999 | | 68.3 | 1.40 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 18 | June, 1999 | | 67.1 | 2.07 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 19 | Jan-Aprl-99 | | 39.4 | 5.38 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 20 | Jan.? 1999 | | 70.3 | 1.45 | 0.7094 | -0.03 | -7.80 |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 21 | Dec., 1998 | | 68.5 | 1.93 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 22 | Oct.-Nov., 1998 | | 71.1 | 1.96 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 23 | Sept., 1998 | | 54.3 | 3.70 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 24 | August, 1998 | | 65.8 | 2.47 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 25 | August, 1998 | | 71.8 | 1.88 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 26 | August, 1998 | HMC>>Q>A>LMC>>D | | | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 27 | August, 1998 | | 62.3 | 3.11 | | | |
| 3 | bottom | 7/14/1998 | 9/8/2000 | 28 | August, 1998 | | 56.1 | 3.33 | 0.7094 | -0.53 | -8.18 |

*Surface—trap 10 m below surface of water.  Bottom—trap 2 m above lake bottom.
[#]XRD mineralogy: HMC—high-Mg calcite; LMC—low-Mg calcite; A—aragonite; D—dolomite; Q—quartz.

Lat. 41° 57' 55.39"
Long. 111° 20' 18.30"

Bear Lake Site 2
lake center

Deployed:
14 July 1998
Water Depth:
150 Ft.
(43 m)

Bear Lake water @ 4 m:
Ca = 32 ppm   Alk. = 293 ppm
Mg = 52 ppm   SO4 = 68 ppm
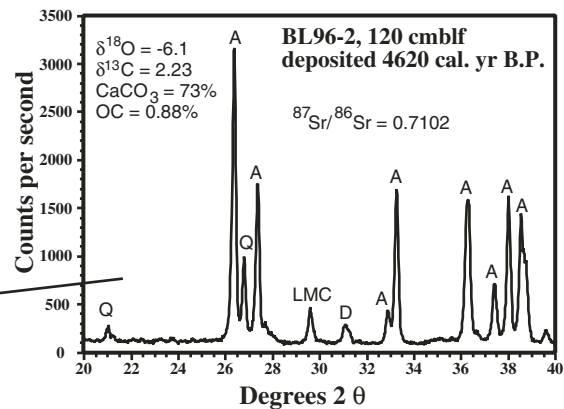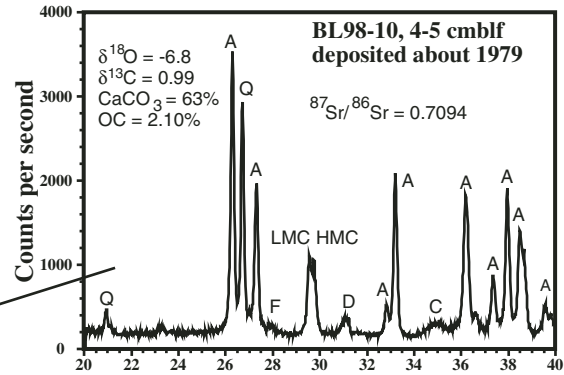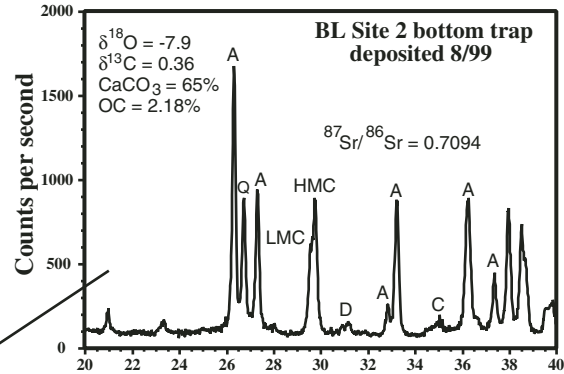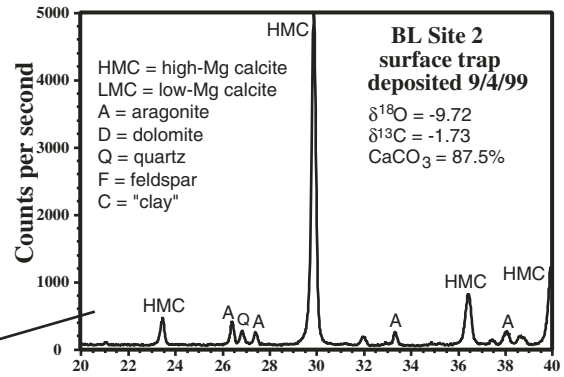Na = 41 ppm   Cl = 43 ppm
K = 4.6 ppm
$^{87}Sr/^{86}Sr = 0.70920$

10 m

Bear Lake water @ 23 m:
Ca = 28 ppm   Alk. = 277 ppm
Mg = 53 ppm   SO4 = 70 ppm
Na = 40 ppm   Cl = 45 ppm
K = 5.0 ppm
$^{87}Sr/^{86}Sr = 0.70917$

Bear Lake water @ 50 m:
Ca = 28 ppm   Alk. = 284 ppm
Mg = 53 ppm   SO4 = 71 ppm
Na = 41 ppm   Cl = 46 ppm
K = 4.8 ppm
$^{87}Sr/^{86}Sr = 0.70920$

40 m

-5 cm

-120 cm

**BL Site 2 surface trap deposited 9/4/99**
HMC = high-Mg calcite
LMC = low-Mg calcite
A = aragonite
D = dolomite
Q = quartz
F = feldspar
C = "clay"
$\delta^{18}O = -9.72$
$\delta^{13}C = -1.73$
CaCO3 = 87.5%

**BL Site 2 bottom trap deposited 8/99**
$\delta^{18}O = -7.9$
$\delta^{13}C = 0.36$
CaCO3 = 65%
OC = 2.18%
$^{87}Sr/^{86}Sr = 0.7094$

**BL98-10, 4-5 cmblf deposited about 1979**
$\delta^{18}O = -6.8$
$\delta^{13}C = 0.99$
CaCO3 = 63%
OC = 2.10%
$^{87}Sr/^{86}Sr = 0.7094$

**BL96-2, 120 cmblf deposited 4620 cal. yr B.P.**
$\delta^{18}O = -6.1$
$\delta^{13}C = 2.23$
CaCO3 = 73%
OC = 0.88%
$^{87}Sr/^{86}Sr = 0.7102$

Counts per second

Degrees 2 θ

Figure 2. Diagram showing the placement of surface (10 m) and bottom (40 m) sediment traps at site 2 (Fig. 1A) in the center of Bear Lake in a water depth of 43 m. Results of chemical analyses and $^{87}Sr/^{86}Sr$ of water from depths of 4 m, 23 m, and 50 m are shown to the left of the trap diagram (data from Dean et al., 2007). X-ray diffractograms of samples of bulk sediment from the two traps, from 4 to 5 cm in short core BL98-10, and from 150 centimeters below lake floor (cmblf) in piston core BL96-2 are shown to the right of the trap diagram.
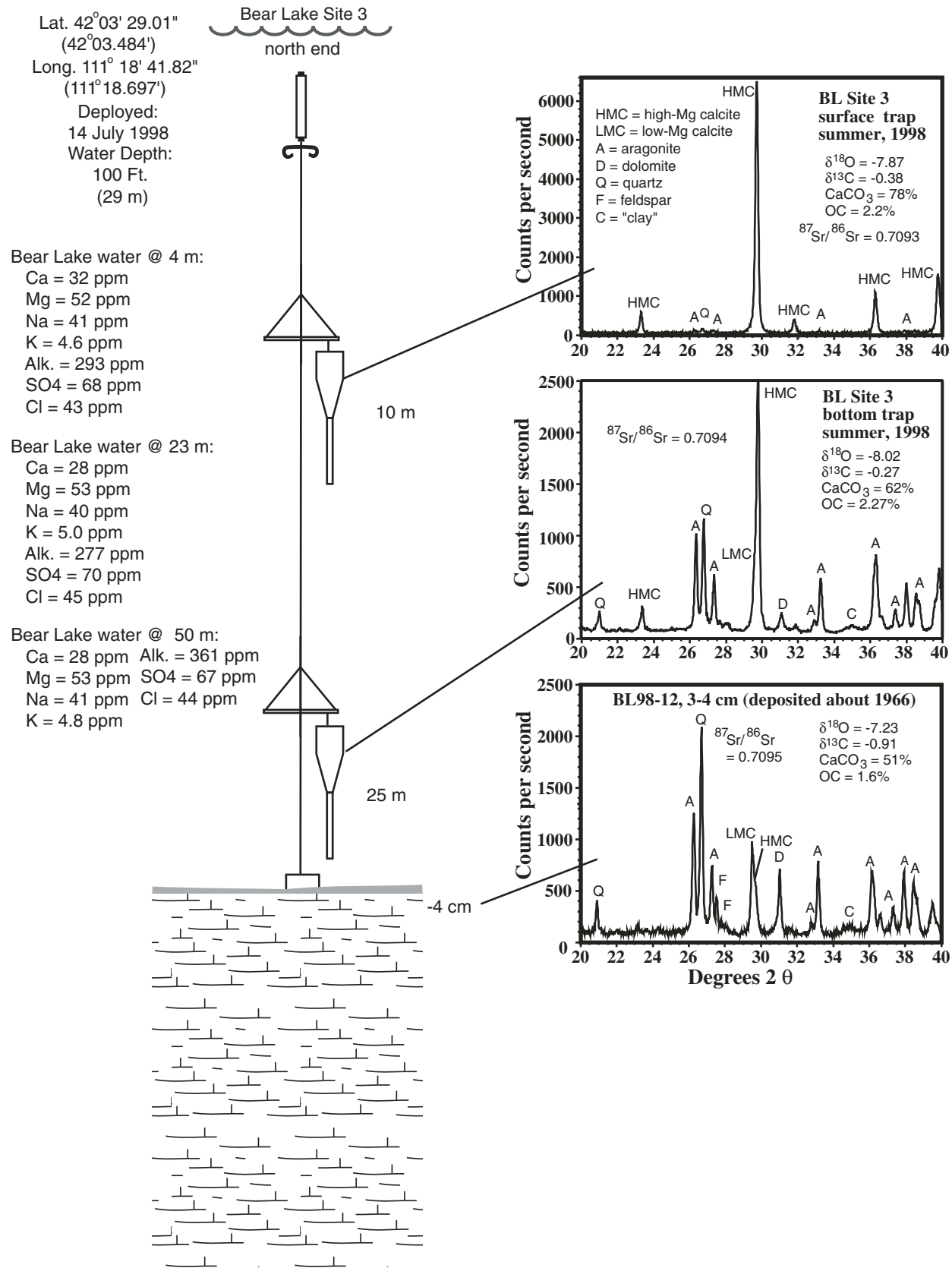
Figure 3. Diagram showing the placement of surface (10 m) and bottom (25 m) sediment traps at site 3 (Fig. 1A) at the north end of Bear Lake in a water depth of 29 m. Results of chemical analyses and $^{87}Sr/^{86}Sr$ of water from depths of 4 m, 23 m, and 50 m are shown to the left of the trap diagram (data from Dean et al., 2007). X-ray diffractograms of samples of bulk sediment from the two traps, and from 3 to 4 cm in short core BL98-12 are shown to the right of the trap diagram.

precipitates in the summer months during strong phytoplankton blooms, and aragonite forms crusts on the leaves of *Potomogeton*, a rooted aquatic plant. The reason for the difference "remains to be solved" (Müller, 1971). On the basis of empirical observations, such as those from Lake Balaton and elsewhere, Müller et al. (1972) suggested that low-Mg calcite forms in lakes having a $Mg^{2+}$:$Ca^{2+}$ of <2; high-Mg calcite forms in lakes having a $Mg^{2+}$:$Ca^{2+}$ of 2–12; and aragonite precipitates in lakes having a $Mg^{2+}$:$Ca^{2+}$ of >12. Last (1982) found that modern bottom sediments in the south basin of Lake Manitoba, Canada, consist mainly of high-Mg calcite with minor amounts of low-Mg calcite, dolomite, and aragonite. He concluded that the low-Mg calcite and dolomite were detrital and that the aragonite came from shell fragments. The high-Mg calcite is derived from precipitation in the water column triggered by photosynthetic removal of $CO_2$ and concomitant increase of pH. Last (1982) thought that the high $Mg^{2+}$:$Ca^{2+}$ in the water (1.7) was responsible for generating high-Mg calcite rather than low-Mg calcite.

Positive values of the saturation index (SI; Table 2) show that Bear Lake water is oversaturated with respect to calcite, aragonite, and dolomite at all depths, and $Mg^{2+}$:$Ca^{2+}$ is ~1.9 (Dean et al., 2007). The 1912 analysis of lake water (Kemmerer et al., 1923; Dean et al., 2007) indicates that the lake was also saturated with respect to all three minerals at that time, but $Mg^{2+}$:$Ca^{2+}$ was 38, definitely favoring aragonite.

I assume, but have no direct proof, that the precipitation of high-Mg calcite in the epilimnion of Bear Lake is biologically mediated through photosynthetic removal of $CO_2$ and an increase in pH during the warm summer months (Otsuki and Wetzel, 1974; Kelts and Hsü, 1978; Stabel, 1986; Dean and Megard, 1993; Dean, 1999). The average July temperature of surface water from 1989 to 2004 was 20.4 °C (Dean et al., this volume). The average pH at the surface from 1989 to 2004 was 8.43 ± 0.25 and that at 40 m was 8.36 ± 0.27 (V. Lamarra, 2004, personal commun.). Algal cells can also serve as nuclei for formation of particulate $CaCO_3$ (Stabel, 1986; Wetzel, 2001). Bear Lake is oligotrophic and generally contains low surface concentrations of chlorophyll *a* in summer (Wurtsbaugh and Luecke, 1997; Dean et al., this volume). Diatoms constitute ~80% of the algal population in Bear Lake (Birdsey, 1989), but algal blooms rarely occur (S. Tolentino, C. Luecke, and V. Lamarra, 2000, personal commun.). However, the sediments collected in the surface trap at site 2 in April 1999 included a considerable amount of green algal debris (Fig. 5). Therefore the algal bloom may have mediated the initiation of carbonate precipitation, which continued through the summer.

The dominant carbonate mineral in the sediments deposited in all surface traps during 1998, 1999, and 2000 was high-Mg calcite (Table 1). A layer ~1 cm thick of carbonate sediment accumulated in the trap during late July and early August 1998, but little sediment accumulated in August and September of that year (Fig. 5). In contrast, a much larger amount of carbonate-rich sediment accumulated in July, August, and September 1999, perhaps in response to the April 1999 algal bloom. With a trap amplification ratio of 225 (ratio of the area of the trap opening to the cross sectional area of the collection tube) a 15-cm-thick layer in the collection tube (e.g., 06 July to 05 August 1999, Fig. 5) translates to about a 0.7-mm-thick layer of sediment on the lake floor if all of the carbonate survived as it settled and accumulated on the lake floor.

Calculations using the chemistry of Bear Lake water (Dean et al., 2007) and the stability of Mg-calcites (Bischoff et al., 1987) indicate that at 20 °C the water is supersaturated with respect to aragonite and all Mg-calcites up to 11 mol % Mg, and
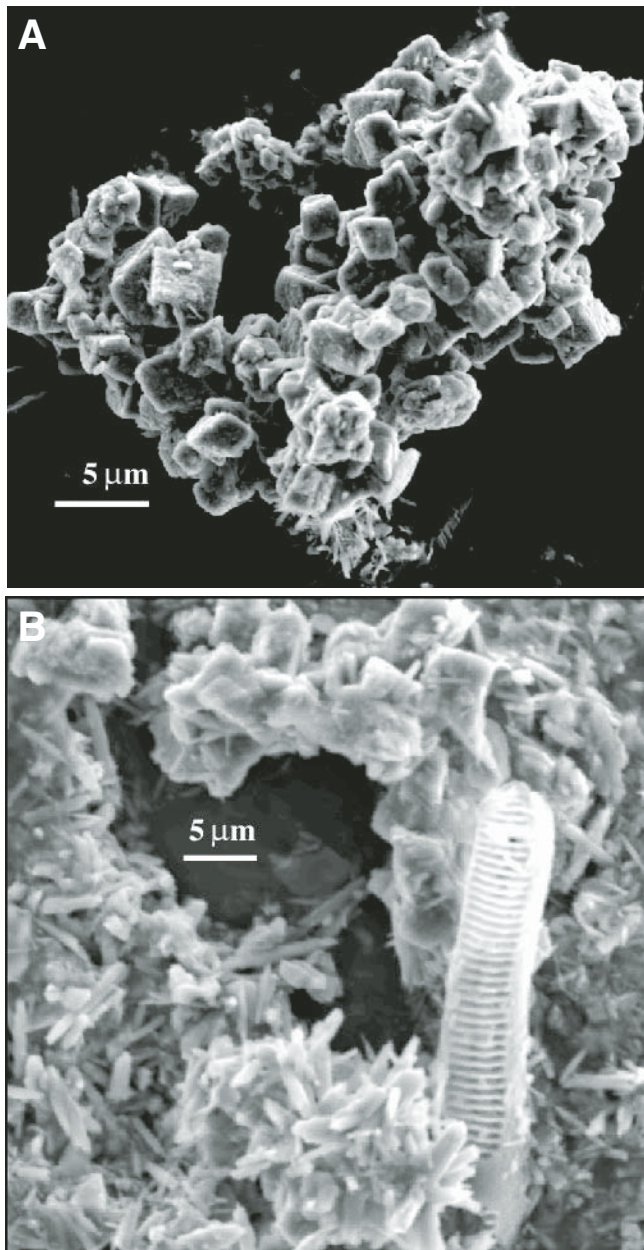
Figure 4. Scanning electron micrographs of bulk sediment showing equant blocks (rounded rhombohedrons) of high-Mg calcite from a surface sediment trap (A), and subrounded needles of aragonite (bottom-left), a pennate diatom (right), and equant blocks of calcite (top-right) from a bottom sediment trap (B).

TABLE 2. SATURATION INDEX (SI), LOG ION ACTIVITY PRODUCT (IAP), AND LOG EQUILIBRIUM
SOLUBILITY PRODUCT (KT) FOR CALCITE, ARAGONITE, AND DOLOMITE IN WATERS FROM BEAR RIVER AND BEAR LAKE

| Location | Calcite | | | Aragonite | | | Dolomite | | |
|---|---|---|---|---|---|---|---|---|---|
| | SI | log IAP | log KT | SI | log IAP | log KT | SI | log IAP | log KT |
| Bear River at gauging station, Idaho | 0.99 | −7.49 | −8.48 | 0.85 | −7.49 | −8.34 | 2.03 | −15.06 | −17.09 |
| Bear Lake, east shore, Utah | 1.29 | −7.19 | −8.48 | 1.14 | −7.19 | −8.34 | 3.21 | −13.88 | −17.09 |
| Bear Lake at 4 m | 0.44 | −8.04 | −8.48 | 0.30 | −8.04 | −8.34 | 1.45 | −15.64 | −17.09 |
| Bear Lake at 50 m | 0.68 | −7.71 | −7.87 | 0.52 | −7.71 | −8.24 | 1.64 | −14.96 | −16.59 |
| Bear Lake, NE of rock pile | 0.37 | −8.08 | −8.45 | 0.22 | −8.08 | −8.31 | 1.27 | −15.70 | −16.97 |
| Bear Lake, 1979 (Lamarra et al., 1986) | −0.67 | −9.12 | −8.45 | −0.82 | −9.12 | −8.31 | −0.57 | −17.55 | −16.97 |
| Bear Lake, 1912 (Kemmerer et al., 1923) | −0.21 | −8.67 | −8.45 | −0.36 | −8.67 | −8.31 | 1.44 | −15.54 | −16.97 |

*Note:* Data from Dean et al. (2007).



# Bear Lake
## Utah/Idaho

### Shallow Trap
### 10 m
### Below Water Surface
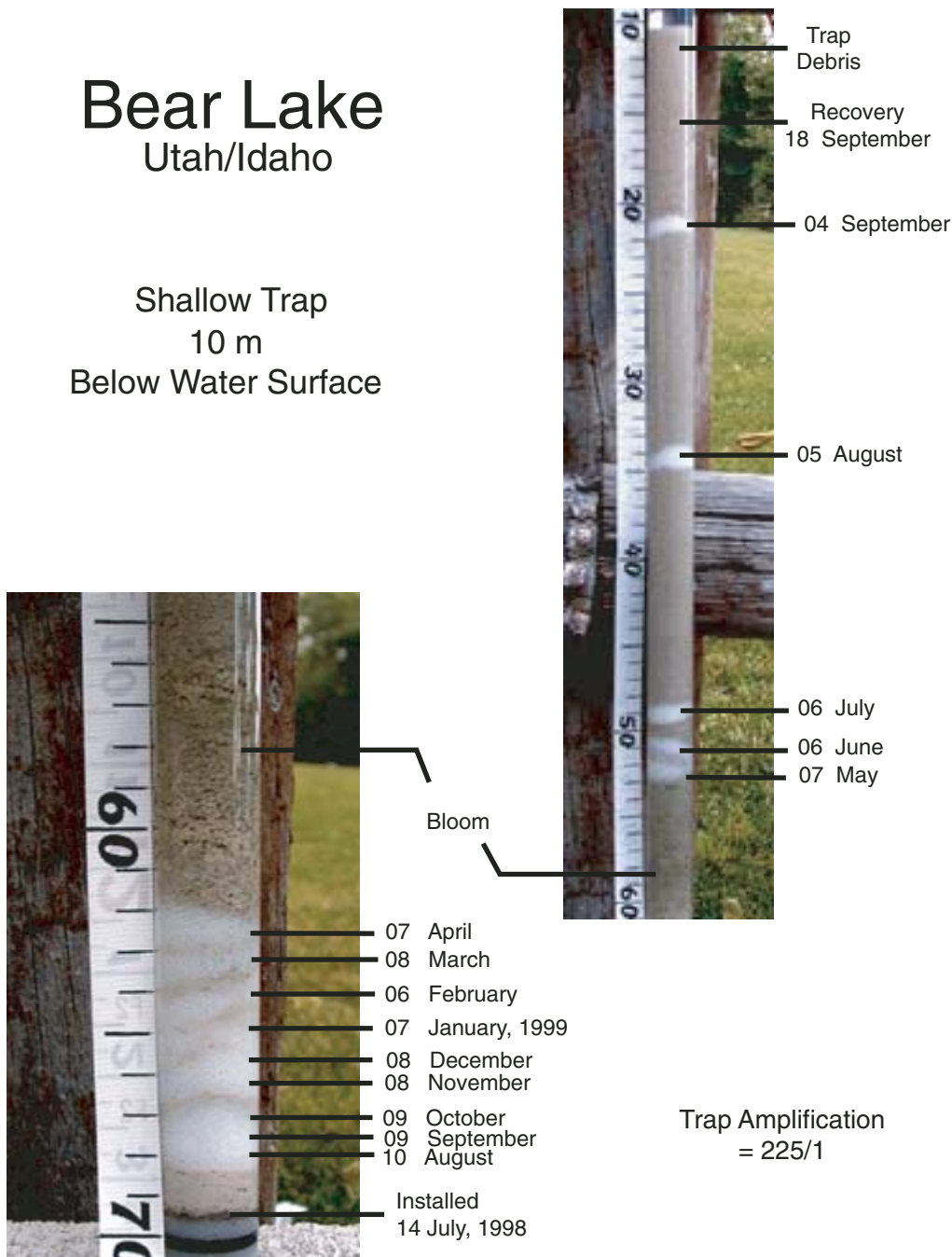
Figure 5. Photograph of the collection tube from a surface sediment trap (10 m) at site 2 (Figs. 1 and 2), deployed on 14 July 1998 and recovered 18 September 1999. White layers are composed of Teflon granules dispensed every 30 days by an automatic timer in the cone of the trap. Note that little sediment accumulated between 10 August 1998 and 7 April 1999. The tan marl deposited between 7 April and 7 May contains abundant phytoplankton debris produced by an algal bloom that was detected in the water column by large increases in chlorophyll *a* (Dean et al., this volume; Fig. 9). The collection tube contained buffered formalin to help preserve any organic matter. All of the CaCO$_3$ deposited in 1999 consists of high-Mg calcite.

Trap
Debris

Recovery
18 September

04 September

05 August

06 July
06 June
07 May

Bloom

07 April
08 March
06 February
07 January, 1999
08 December
08 November
09 October
09 September
10 August

Installed
14 July, 1998

Trap Amplification
= 225/1

that calcite with 4 mol % Mg is the least soluble (i.e., the most stable) calcite composition (Bischoff et al., 2005). At 4 °C, low-Mg calcite is more stable than high-Mg calcite, and aragonite is less stable than either (Bischoff et al., 2005). This, together with the appearance of low-Mg calcite and high-Mg calcite in the bottom traps (Figs. 2 and 3), suggest that metastable high-Mg is being dissolved, and recrystallized as low-Mg calcite, in the cold hypolimnion. Such a stabilization reaction likely occurs as a microsolution-reprecipitation reaction involving calcitization of the original high-Mg calcite (Mackenzie et al., 1983). Thus, high-Mg calcite formed in the epilimnion would dissolve in the water column and reprecipitate as low-Mg calcite. Of course, some of the low-Mg calcite may be detrital.

The above considerations answer the first and second questions posed above, but fail to explain the appearance of aragonite in the hypolimnion (question 3). Calculations by Bischoff et al. (2005) show that in Bear Lake, high-Mg calcite cannot convert to aragonite. In lakes in the United States, aragonite precipitation usually occurs in saline prairie lakes (e.g., Shapley et al., 2005) and does not occur in large, cold, deep, oligotrophic, high-altitude, north temperate lakes fed by snowmelt. Aragonite is a major component in the sediments of Beaver Lake, a marl lake in southeastern Wisconsin, but there the aragonite is derived from mollusk shell fragments (Brown et al., 1992). Dolomite and aragonite are common in the sediments of the saline lakes of the northern Great Plains (North Dakota, Montana, Saskatchewan, and Alberta) where the lake chemistry is dominated by $Na^+-Mg^{2+}-SO_4^{2-}$ waters (Last and Schweyen, 1983). In most lakes where evaporation exceeds precipitation, the volume of groundwater and surface-water inflow is inversely proportional to evaporation. That is, as evaporation increases, the inflow of water decreases, because the lake's hydrologic budget is linked to climate. Therefore, increased evaporation results in less inflow and a saline lake precipitating evaporite minerals or a dry lake. The conditions that form aragonite in saline lakes of the northern Great Plains do not apply to Bear Lake. To help answer the question of where the aragonite in the bottom-trap and post-diversion sediments came from, we need to take a close look at the geochemistry and mineralogy of pre- and post-diversion carbonate, particularly the O, C, and Sr isotopic composition of those sediments.

## Surface Sediments

### *Isotope Geochemistry*

Values of $\delta^{18}O$, $\delta^{13}C$, and $^{87}Sr/^{86}Sr$ in bulk carbonate decrease abruptly and markedly at ~12 cm in basin-center core BL98-10 (Fig. 6). I interpret these changes to represent the chemical signature of Bear River diversion early in the twentieth century, reflecting the more $^{18}O$-, $^{13}C$-, and $^{87}Sr$-depleted waters of Bear River (Dean et al., 2007). If this interpretation is correct, the $^{210}Pb$ ages in years AD in Figure 6 must be too old, particularly the older ones. Smoak and Swarzenski (2004) state that "sediments deposited prior to 100 years ago are assigned artificially 'too old' dates by the CRS [constant rate of supply] model" and that "error

terms are large on the older dates." Sediments deposited prior to diversion have values of $\delta^{18}O$ of about −4‰, whereas the carbonates that accumulated in the less saline waters of the lake after diversion have values of $\delta^{18}O$ of −7‰ to −7.5‰, slightly more enriched in $^{18}O$ than in sediments from bottom traps and much more enriched than in sediments from surface traps (Table 1, Figs. 2, 3, and 6). Values of $\delta^{13}C$ in pre-diversion sediments are 2.5‰–3‰, whereas those of post-diversion sediments are <1‰ (Figs. 2, 3, and 6), slightly more enriched in $^{13}C$ than in sediments from the bottom traps and much more enriched than in sediments from surface traps (Table 1, Figs. 2, 3, and 6).

The Sr isotopic ratio of any endogenic carbonate (precipitate or shell) will be the same as in the water from which it formed (e.g., Capo and DePaolo, 1990; Hart et al., 2004). Therefore, the Sr isotopic composition of carbonate that formed in the epilimnion of Bear Lake will be the same as that of the lake water, and we can use $^{87}Sr/^{86}Sr$ to interpret the source of pre-diversion Bear Lake water, and the origin of the mixtures of minerals in bulk carbonate in the trap sediments. Values of $^{87}Sr/^{86}Sr$ in carbonates deposited before diversion are around 0.7102 (Fig. 6), similar to values in stream and spring waters today on the west side of the lake (Fig. 6; Dean et al., 2007). The west-side stream sediments contain a considerable amount of dolomite and calcite (Table 3), presumably derived mainly from Paleozoic carbonate rocks in the Bear River Range to the west of the lake (Oriel and Platt, 1980; Dover, 1985). Values of $^{87}Sr/^{86}Sr$ in carbonates deposited in the lake after diversion are around 0.7094, slightly more enriched in $^{87}Sr$ than Bear Lake water today ($^{87}Sr/^{86}Sr$ = 0.7092; range 0.70917–0.70922; Dean et al., 2007), and considerably more enriched than Bear River water (~0.7085). Values of $^{87}Sr/^{86}Sr$ in the present lake are intermediate between the pre-diversion lake and Bear River. Following the diversion, values of $^{87}Sr/^{86}Sr$ in Bear Lake (i.e., in post-diversion sediments) started to follow an $^{87}Sr$-depletion trend toward values in Bear River (12–8 cm in core BL98-10, Fig. 6), but sediments deposited after AD 1936 ($^{210}Pb$, Fig. 6) are still enriched in $^{87}Sr$ ($^{87}Sr/^{86}Sr$ = 0.7094) relative to present lake water ($^{87}Sr/^{86}Sr$ = 0.7092). Therefore, the Sr isotopic composition of the most recent sediments does not reflect the Sr isotopic composition of modern lake water.

### *Mineralogy*

Prior to diversion of Bear River into Bear Lake in the early part of the twentieth century, there is no record of the Bear River entering Bear Lake during historic times. The only information about the water chemistry of Bear Lake prior to diversion is an analysis of water collected in 1912, which showed concentrations of total dissolved solids (TDS), $Mg^{2+}$, and $Ca^{2+}$, of 1060, 152, and 4.1 mg $L^{-1}$, respectively (Kemmerer et al., 1923). Corresponding values in Bear Lake today are 500, 53, and 31 mg $L^{-1}$, respectively. Such changes should have had a large effect on carbonate minerals precipitating in the lake and on the chemistry of those minerals.

Analyses of cores of surface sediments (up to 30–50 cm long) indicate that prior to diversion the dominant mineral in Bear Lake sediments was aragonite (Dean et al., 2007). After diversion, aragonite continued to be the dominant polymorph of $CaCO_3$ even

though $Mg^{2+}:Ca^{2+}$ in the water was much lower. High-Mg calcite began to appear in sediments deposited shortly after Bear River diversion (Fig. 6; Dean et al., 2007).

**Source of the Aragonite**

The Bear River diversion caused profound changes in the isotopic compositions of C, O, and Sr and the concentrations of

Mg and Sr in carbonate minerals (Fig. 6). The ionic radius of Sr is larger, and that of Mg is smaller, than that of Ca. Therefore, Sr more readily substitutes for Ca in the more open crystal lattice of aragonite, and Mg more readily substitutes for Ca in calcite, so that calcite typically has a lower concentration of Sr and a higher concentration of Mg relative to aragonite. The higher concentration of Sr and lower concentration of Mg in bulk carbonate when aragonite precipitation ceased after the introduction of Bear River water
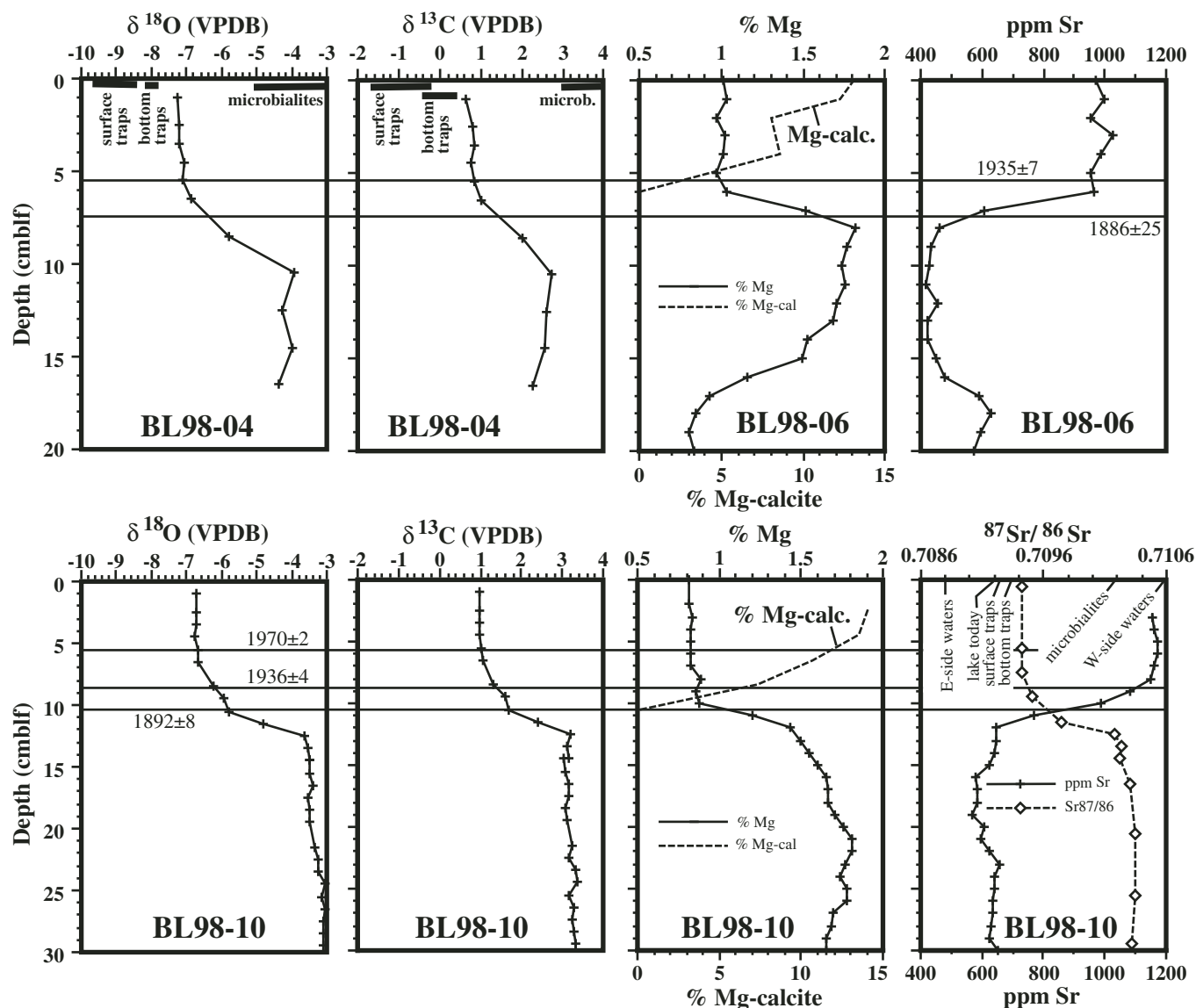


Figure 6. Profiles of values of $\delta^{18}O$ and $\delta^{13}C$ in bulk-sediment samples from cores BL98-04 and BL98-10, concentrations of HCl-soluble Mg and Sr (Bischoff et al., 2005), and abundance of high-Mg calcite calculated from XRD peak intensities (dashed lines) in cores BL98-06 and BL98-10, all versus depth in centimeters below lake floor (cmblf). Values of $^{87}Sr/^{86}Sr$ in carbonate in core BL98-10 also are given (dashed line with diamond symbols). Ranges of values of $\delta^{18}O$ and $\delta^{13}C$ in surface and bottom sediment traps and in microbialites are shown as bars at the tops of the profiles of $\delta^{18}O$ and $\delta^{13}C$ for core BL98-04. Average values of $^{87}Sr/^{86}Sr$ in Bear Lake, east-side springs and creeks, west-side springs and creeks, surface-sediment traps, bottom-sediment traps, and microbialites are shown at the top of the profile of $^{87}Sr/^{86}Sr$ for core BL98-10. Values of $^{87}Sr/^{86}Sr$ for waters are from Dean et al. (2007). Selected $^{210}Pb$ dates (in years A.D.; Smoak and Swarzenski, 2004) are shown for each core. As discussed in the text, these dates, especially the older ones, are probably too old. XRD—x-ray diffraction; VPDB—Vienna Pee Dee Belemnite.

(Fig. 6) is counterintuitive. This result is due to the much higher concentration of Sr, and much lower concentration of Mg, in Bear River water relative to the pre-diversion lake (Dean et al., 2007).

The sediments were much more enriched in $^{13}$C, $^{18}$O, and $^{87}$Sr prior to diversion. Assuming that $^{87}$Sr/$^{86}$Sr in pre-diversion bulk carbonate is representative of the ratio in pre-diversion lake water, then the 1912 Bear Lake had a ratio of ~0.7102 (Fig. 6), which is within the range of west-side waters (0.7101–0.7132), indicating that the west-side waters probably dominated the lake's hydrologic budget. A calculation considering only stream input indicates that ~99% of the 1912 solutes came from west-side waters (Dean et al., 2007).

As pointed out earlier, the Sr isotopic composition of sediments deposited after AD 1936 (by $^{210}$Pb) does not reflect the Sr isotopic composition of modern lake water. Any carbonate mineral that precipitates in Bear Lake today should have an $^{87}$Sr/$^{86}$Sr of 0.7092, the ratio in lake water today (Fig. 6). The average $^{87}$Sr/$^{86}$Sr in bulk carbonate in surface traps is ~0.70925, and that in bottom traps is ~0.70937 (Table 1, Figs. 2, 3, and 6). The bulk carbonate in the trap sediments, especially in the bottom traps, and in the most recently deposited profundal sediments is, therefore, a mixture of carbonate precipitated in the lake today and carbonate more enriched in $^{87}$Sr (higher $^{87}$Sr/$^{86}$Sr), and also more enriched in $^{18}$O and $^{13}$C (Fig. 6).

X-ray diffraction patterns (Figs. 2 and 3) show that the surface traps contain a small amount of aragonite in addition to the dominant high-Mg calcite. If that enriched aragonite had a $^{87}$Sr/$^{86}$Sr of 0.7094 (that of twentieth-century carbonate; Fig. 6), a simple calculation, based on mixing a 0.7094 twentieth-century carbonate

(mostly aragonite) with a 0.7092 endogenic high-Mg calcite, and ignoring differences in Sr contents, shows that the surface traps (with a $^{87}$Sr/$^{86}$Sr of 0.70925), on average, contain 75% modern carbonate (high-Mg calcite) and 25% twentieth-century carbonate (mostly aragonite). Because aragonite contains a higher concentration of Sr relative to calcite, this is a liberal estimate for the contribution of high-Mg calcite, which might be less than 75%. The same calculation for the bottom traps shows that, on average, they contain 15% modern carbonate and 85% twentieth-century carbonate. I conclude, therefore, that the aragonite collected in the traps did not precipitate from present-day Bear Lake water, but is reworked aragonite that is at least 50 yr old.

The aragonite in sediments deposited after diversion must have continued to form in the lake for some period of time. If it was reworked pre-diversion aragonite, it would have the isotopic and chemical signatures of pre-diversion aragonite (Fig. 6). Instead, the post-diversion aragonite recorded the changing isotopic and chemical composition of the lake. High-Mg calcite started forming shortly after diversion, and sediments deposited during the latter part of the twentieth century have increasing proportions of high-Mg calcite (Fig. 6). The horizon in the sediments that indicates environmental change due to Bear River diversion is well defined by the C, O, and Sr isotope data (Fig. 6). The differences in isotopic values between the high-Mg calcite in surface traps and the aragonite in pre-diversion sediments represents the actual environmental isotopic change in lake water (Fig. 6). The isotopic values of the twentieth-century surface sediments do not represent the full range of change in the lake water because of mixing of modern endogenic high-Mg calcite and older aragonite that is more enriched in $^{87}$Sr (and $^{18}$O and $^{13}$C).

TABLE 3. 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 (SEE FIG. 1B FOR LOCATIONS OF SAMPLES)

| Sample number | Stream | Quartz (%)* | Dolomite (%)* | Calcite (%)* | Feldspar (%)* | CaCO$_3$ (%) | OC (%) | δ$^{18}$O-carbonate (per mil, VPDB) | δ$^{13}$C-carbonate (per mil, VPDB) | $^{87}$Sr/$^{86}$Sr (carbonate) | $^{87}$Sr/$^{86}$Sr (residue) |
|---|---|---|---|---|---|---|---|---|---|---|---|
| West-side creeks | | | | | | | | | | | |
| 1 | Paris Creek | 42 | 56 | 0 | 3 | 0.0 | 5.74 | −14.3 | −6.1 | 0.71001 | 0.72098 |
| 7 | St. Charles Creek | 52 | 36 | 6 | 6 | | | −13.0 | −5.7 | 0.71004 | 0.71663 |
| 9 | Fish Haven Creek | 58 | 26 | 11 | 5 | | | | | 0.71030 | 0.72283 |
| 11 | Hodges Canyon | 81 | 3 | 5 | 11 | 0.0 | 0.81 | −14.9 | −8.2 | 0.71025 | 0.71953 |
| East-side creeks | | | | | | | | | | | |
| 13 | South Eden Creek | 63 | 4 | 28 | 4 | 25.6 | 1.85 | −13.4 | −7.3 | 0.70892 | 0.71264 |
| 13 | South Eden Creek | | | | | | | −11.4 | −5.0 | | |
| 15 | North Eden Creek | 85 | 3 | 8 | 4 | 10.8 | 0.65 | −13.5 | −4.8 | 0.70938 | 0.71489 |
| 15 | North Eden Creek | | | | | | | −12.4 | −5.0 | | |
| 17 | Indian Creek | 69 | 3 | 21 | 7 | | | −11.6 | −4.5 | 0.70768 | 0.71223 |
| Bear River north of Uintas | | | | | | | | | | | |
| 19 | Bear River | 85 | 3 | 9 | 3 | 9.9 | 0.49 | −12.7 | −5.0 | 0.70907 | 0.72417 |
| 21 | Bear River | 75 | 8 | 11 | 5 | | | | | 0.70907 | 0.71737 |
| 23 | Bear River | 78 | 4 | 11 | 6 | 12.1 | 1.34 | −11.5 | −4.0 | 0.70929 | 0.71646 |
| 25 | Bear River | 81 | 5 | 7 | 7 | 11.3 | 0.41 | −11.9 | −5.4 | 0.71006 | 0.71942 |
| 27 | Bear River | 75 | 6 | 16 | 3 | | | | | 0.71067 | 0.72131 |
| 29 | Bear River | 85 | 3 | 10 | 2 | 13.2 | 0.89 | −12.8 | −5.7 | 0.71027 | 0.72169 |
| 31 | Bear River | 77 | 6 | 17 | 4 | | | | | 0.71078 | 0.71911 |
| 33 | Bear River | 85 | 5 | 9 | 3 | 4.5 | 0.38 | −12.2 | −5.0 | 0.71066 | 0.74241 |
| 35 | Bear River | 80 | 6 | 12 | 8 | 11.6 | 2.39 | −11.4 | −5.9 | 0.71112 | 0.72474 |
| 37 | Bear River | 83 | 6 | 10 | 9 | | | | | | |
| 39 | Bear River | 85 | 6 | 8 | 4 | 8.0 | 2.26 | −10.0 | −5.0 | 0.71190 | 0.72881 |
| 41 | Bear River | 77 | 7 | 15 | 6 | 12.9 | 1.00 | −10.1 | −5.3 | 0.71050 | 0.73716 |
| 43 | Bear River | 70 | 9 | 20 | 10 | 22.2 | 1.64 | −9.3 | −3.5 | 0.70990 | 0.72142 |
| 45 | Bear River | 72 | 5 | 22 | 12 | | | | | | |
| Bear River in Uintas | | | | | | | | | | | |
| 51 | Bear River | 83 | 0 | 0 | 17 | | | | | | |
| 53 | Bear River | 88 | 0 | 0 | 12 | | | | | 0.71910 | 0.74348 |

*Note:* OC—organic carbon; VPDB—Vienna Pee Dee Belemnite.
*XRD percentages were calculated from peak intensities.

## The Rock Pile

Another part of the endogenic carbonate story in Bear Lake is a mound of carbonate-cemented cobbles and pebbles, known locally as the "rock pile," at the southwestern end of the lake (Fig. 1A), notorious for snagging fishing lines and anchor ropes. This mound rises from a depth of ~18 m below the surface on the north end of the mound to a hummocky surface at a depth of ~12 m. Samples from the rock pile were first provided by Bryce Nielson and Scott Tolentino of the Utah State Division of Wildlife Resources in Garden City, Utah, who brought them up in their gill nets. Each sample had a rounded cobble as a nucleus covered by a thin (<1 cm) layer of carbonate. X-ray diffraction analyses of these samples showed that the carbonate layers were composed of aragonite, and the rounded cobble nuclei were composed of dolomite, suggesting that they came from the Paleozoic carbonate bedrock in the Bear River Range. Limited sampling, underwater video, and SCUBA-based observations indicate that the microbialite mound is made up of many rounded cobbles and pebbles that have been coated and cemented together. Divers described the entire surface of the mound as being coated with a layer of algae. This was subsequently confirmed by underwater video. For this reason I used the term algal mound, or, more correctly, microbialite mound for the rock pile.

Microbialites, or what were formally referred to as "stromatolites," are "organosedimentary deposits that have accreted as a result of a benthic microbial community trapping and binding detrital sediment and/or forming the locus of mineral precipitation" (Burne and Moore, 1987). The role of microbes might simply be through trapping and binding of sediment particles or also through induced or mediated precipitation of $CaCO_3$ by increasing the pH in the micro-environment surrounding the microbes. Once formed, a microbialite may become the locus of passive inorganic cementation. The term microbialite is also used here for individual samples of aragonite-coated cobbles and pebbles.

The largest reported microbialites in lakes are towers up to 40 m high in alkaline (pH>9.7) Lake Van, Turkey (Kempe et al., 1991). The Lake Van microbialites occur where Ca- and $CO_2$-rich groundwater seeps into the lake, precipitating aragonite. The soft aragonite crusts are then colonized by cyanobacteria, which precipitate more aragonite and stabilize the growing mound.

Although the microbial coating suggests that the microbialite mound in Bear Lake is still forming, radiocarbon ages on four samples of the mound range in age from 3000 to 6500 ^{14}C yr B.P. (Table 4). The difference in age between the bottom of the carbonate layer (just above the dolomite cobble nucleus) and the top of the carbonate layer (outer edge) in the initial three small samples collected in gill nets was ~2000 yr (Table 4, microbialites A, B, and C). A larger sample of the microbialite mound collected by divers in 2002 had a thick rind of aragonite, and a radiocarbon age difference from top to bottom of 3500 yr (Table 4, Fig. 7). The radiocarbon ages suggest that the microbialite mound is not modern but formed during the middle to late Holocene.

Samples of carbonate cement were drilled along transects across the thick carbonate rind of the 2002 microbialite (Fig. 7) for O, C, and Sr isotope analyses. Values of $^{87}Sr/^{86}Sr$ in three samples from the 2002 microbialite (Fig. 8, transect A1–C) and from two other microbialites are remarkably constant, ranging from 0.7101 to 0.7103. These values are similar to those in pre-diversion sediments in cores BL98-10 (Fig. 6) and BL02-4 (Fig. 8). Values of $\delta^{18}O$ and $\delta^{13}C$ range from −3‰ to −5‰, and from 2.5‰ to 4‰, respectively. Analyses of samples drilled from the 2002 microbialite on parallel transects (transects X, R, and G, Fig. 7) produced essentially identical results. Values of $\delta^{18}O$ >−4‰ and values of $\delta^{13}C$ >3‰ (vertical lines on profiles in Fig. 8) occur only in sediments deposited during the past 1500 yr, but before Bear River diversion (Fig. 8). The high values of $\delta^{18}O$ and $\delta^{13}C$ indicate that the microbialite mound, as represented by the 2002 sample (Fig. 7), was not precipitating aragonite during the mid-Holocene because the microbialite samples are much more enriched in $^{18}O$, $^{13}C$, and $^{87}Sr$ than mid-Holocene sediments (discussed below). The core and microbialite C and O isotope data presented in Figure 8 indicate that the microbialite mound, at least the top of the mound represented by these few samples, is only several thousand years old, and was not precipitating aragonite cement during the latter part of the twentieth century. Therefore, the radiocarbon ages (Table 4, Fig. 7) must be too old, and this gives some measure of the radiocarbon reservoir effect in Bear Lake.

Values of $\delta^{13}C$ on the outer edge of the 2002 microbialite sample are more than 1‰ enriched in $^{13}C$ ($\delta^{13}C$ = 4‰–4.5‰, Fig. 8, transects E1–E8 and R1–R6) than Holocene core carbonate ($\delta^{13}C$ = 3‰, Fig. 8, core BL02-4). The higher microbialite values

TABLE 4. RADIOCARBON AGES OF BULK CARBONATE
FROM FOUR MICROBIALITE SAMPLES (A, B, C, AND 2002)

| Sample | Position | [14]C age (yr) | Error (yr) | USGS No. |
|---|---|---|---|---|
| Microbialite A | Top* | 3200 | 40 | 3383 |
| Microbialite A | Bottom[†] | 5810 | 40 | 3384 |
| Microbialite B | Top | 5270 | 40 | 3385 |
| Microbialite B | Bottom | 5400 | 40 | 3386 |
| Microbialite C | Top | 3920 | 40 | 3387 |
| Microbialite C | Bottom | 5550 | 40 | 3388 |
| Microbialite 2002 | Top | 3040 | 40 | 4145 |
| Microbialite 2002 | Middle | 5265 | 40 | 4143 |
| Microbialite 2002 | Bottom | 6550 | 40 | 4144 |

*Top—outer edge of microbialite.
[†]Bottom—just above cobble nucleus.

may represent a slight photosynthetic enrichment by the microbes involved in microbialite formation (Burne and Moore, 1987). All of the carbon isotope results suggest that during the early part of microbialite formation, aragonite cementation was predominantly inorganic, precipitated in isotopic equilibrium with lake water ($\delta^{13}C = 3‰$), and only in later stages of microbialite formation was there a slight photosynthetic enrichment in $^{13}C$ in aragonite cement ($\delta^{13}C >3‰$ in transects E1–E8 and R1–R6; Fig. 8).

A mound of aragonite-cemented cobbles and pebbles at one particular area of Bear Lake is suspicious. Mounds of inorganically precipitated $CaCO_3$ ("tufa mounds"; e.g., at Mono Lake, California) commonly form where $CO_2$-charged groundwaters reach the surface as springs, whereupon they degas, increasing the pH of the water, which triggers the precipitation of $CaCO_3$ (e.g., Mono Lake, California; Lake Van, Turkey). This $CaCO_3$ precipitation may or may not be mediated by the photosynthetic activity of cyanobacteria or other microbes. Because we are unable to balance the present lake ionic composition by using that of inflowing surface waters, I suspect that there is a fairly large extrabasinal supply of groundwater, and that groundwater is enriched in $^{18}O$, $^{13}C$, and $^{87}Sr$ relative to the lake and surface-water sources (Dean et al., 2007).

Persistent sublacustrine "springs" occur in at least 25 localities in Bear Lake, mostly on the west side and in a north-south line (S. Tolentino and B. Nielson, 2002, personal commun.). They are manifested at the surface of the lake as patches of bubbles. Bubble trains in the water column from some of these springs are strong enough to be seen on sonar images (Bright, this volume). In the spring of 2002, I collected the gas bubbles from three of these springs. Upon
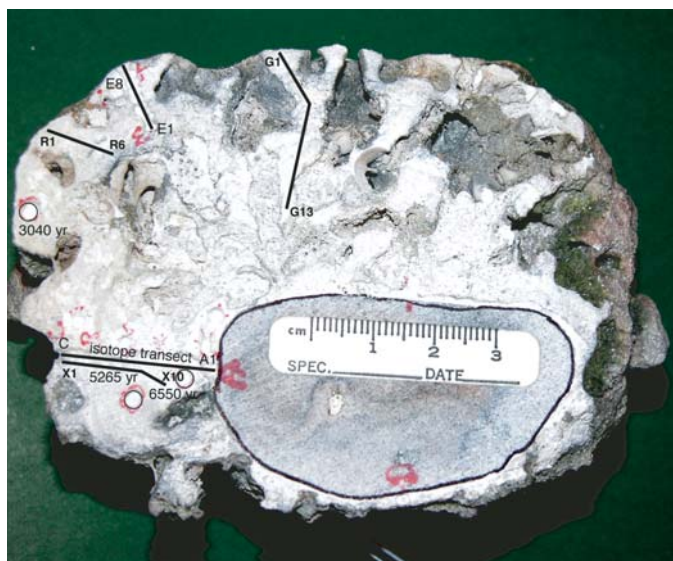


Figure 7. Photograph of a cut half of a 2002 sample of a microbialite containing a rounded dolomite cobble nucleus and a rind of varying thickness of aragonite. Locations of three radiocarbon samples (Table 4) are shown (circles with dates). Letter-number lines are transects of drilled samples for C and O isotope analyses. Results of $\delta^{18}O$ and $\delta^{13}C$ for transects A1–C and E1–E8 are shown in Figure 8.

analyses, all three proved to be methane with $\delta^{13}C$ values of $-68‰$ to $-70‰$ indicating a biogenic origin (Clark and Fritz, 1997). Later that summer, divers observed that there was no water coming from the methane seeps, at least in the summer of 2002.

Methanogenesis is occurring in the sediments of Bear Lake. Coring at site BL00-1 (Fig. 1A) was begun with 3 m drives. However, below ~50 m the sediments were very gassy and expanded in the liner, which necessitated coring with 2 m drives to allow for expansion. Presumably the gas was biogenic methane, although I do not have the appropriate analyses to prove this. However, methane in sediments cannot explain the large volumes of methane originating from vents that are present in the same place year after year and occur in a north-south line parallel to secondary faults that parallel the main bounding fault (East Bear Lake fault; Colman, 2006). Some of these secondary faults cut the youngest sediments and intersect the lake floor. This methane must be coming from a deep source along these secondary faults, and these faults might also be conduits for deep, extrabasinal groundwater.

My current working hypothesis for the origin of the rock pile is that $CO_2$-charged groundwater entered the lake under the microbialite mound, discharged $CO_2$, and precipitated $CaCO_3$, perhaps mediated by cyanobacteria and other microbes in much the same way that microbialite mounds are formed in lakes in many parts of the world (Burne and Moore, 1987). Active building of the microbialite mound apparently began about several thousand years ago and apparently ended when Bear River was diverted into Bear Lake. No carbonate is forming on the top of the microbialite mound where the samples discussed here came from, but if my working hypothesis is correct, groundwater may still be entering from below the mound and precipitating carbonate of some unknown mineralogy.

## Glacial through Holocene Changes in Carbonate Deposition

The sedimentary sequence in Bear Lake consists of a series of eastward dipping strata, the shallowest of which pinch out to the west (Colman, 2006). Therefore, the three Kullenberg cores that were collected (BL96-1, BL96-2, and BL96-3, Fig. 1A) contain overlapping sequences. This can be seen in the profiles of calcite and aragonite versus depth shown in Figure 9. Core BL96-2 contains all of the lithologic units deposited over the last 19 cal k.y. in Bear Lake condensed into 400 cm of section. Therefore, I will focus most of the discussion on that core, supplemented with some isotope data obtained at higher temporal resolution from core BL02-4 from the north end of the lake (Fig. 1A). According to the age model of Colman et al. (2005; this volume), the section collected in core BL96-2 appears to be nearly continuous and intact. However, pyritized tubes in some horizons in BL96-2 are interpreted as root casts and, if so, would appear to indicate very shallow water or subaerial exposure (Smoot, this volume). Shell layers associated with hiatuses in core BL02-4, collected in shallower water than core BL96-2, suggest subaerial exposure and erosion or nondeposition, and lake levels interpreted from

grain-size data in BL02-4 and -3 indicate times of lower lake level (Smoot and Rosenbaum, this volume; discussed below).

### Mineralogy

Mineralogy data for core BL96-2 (Dean et al., 2006) show that the carbonate-poor section deposited prior to 16 cal ka (unit 1) consists of ~75% quartz, 10%–20% $CaCO_3$ as calcite, and minor feldspar and dolomite. Some of the calcite and all of the dolomite deposited during that time probably is detrital. These sediments deposited during the last glacial interval (LGI) consist of red, calcareous, silty clay that are in marked contrast to the light-tan carbonate-rich sediments deposited during the glacial-Holocene transition and the Holocene, regardless of mineralogy.

Several abrupt changes in mineralogy occur within carbonate-rich sediments deposited since 16 cal ka (Fig. 10).

These changes in mineralogy do not always coincide with changes in percentages of $CaCO_3$ (Fig. 10). The percentage of $CaCO_3$, initially as calcite, began to increase ca. 15.5 cal ka (unit 2; Fig. 10), and reached a plateau of ~40% between 15 and 12 cal ka (unit 3). During this interval, minor amounts of aragonite also formed. The percentage of $CaCO_3$ increased to ~70% by 11 cal ka and remained at that level throughout the Holocene and through changes in mineralogy. Between 11 and 9 cal ka, aragonite was the dominant polymorph of $CaCO_3$ formed in the lake (unit 4). A return of calcite precipitation between 9 and 7.5 cal ka (unit 5) indicates a brief freshening event, probably re-entry of Bear River into Bear Lake (see isotope discussion below), which lowered $Mg^{2+}$:$Ca^{2+}$ out of the aragonite stability field. The sediments deposited over the last 7.5 k.y. (unit 6) consist of ~70% $CaCO_3$, mostly as aragonite with only ~5% calcite (Fig. 10).
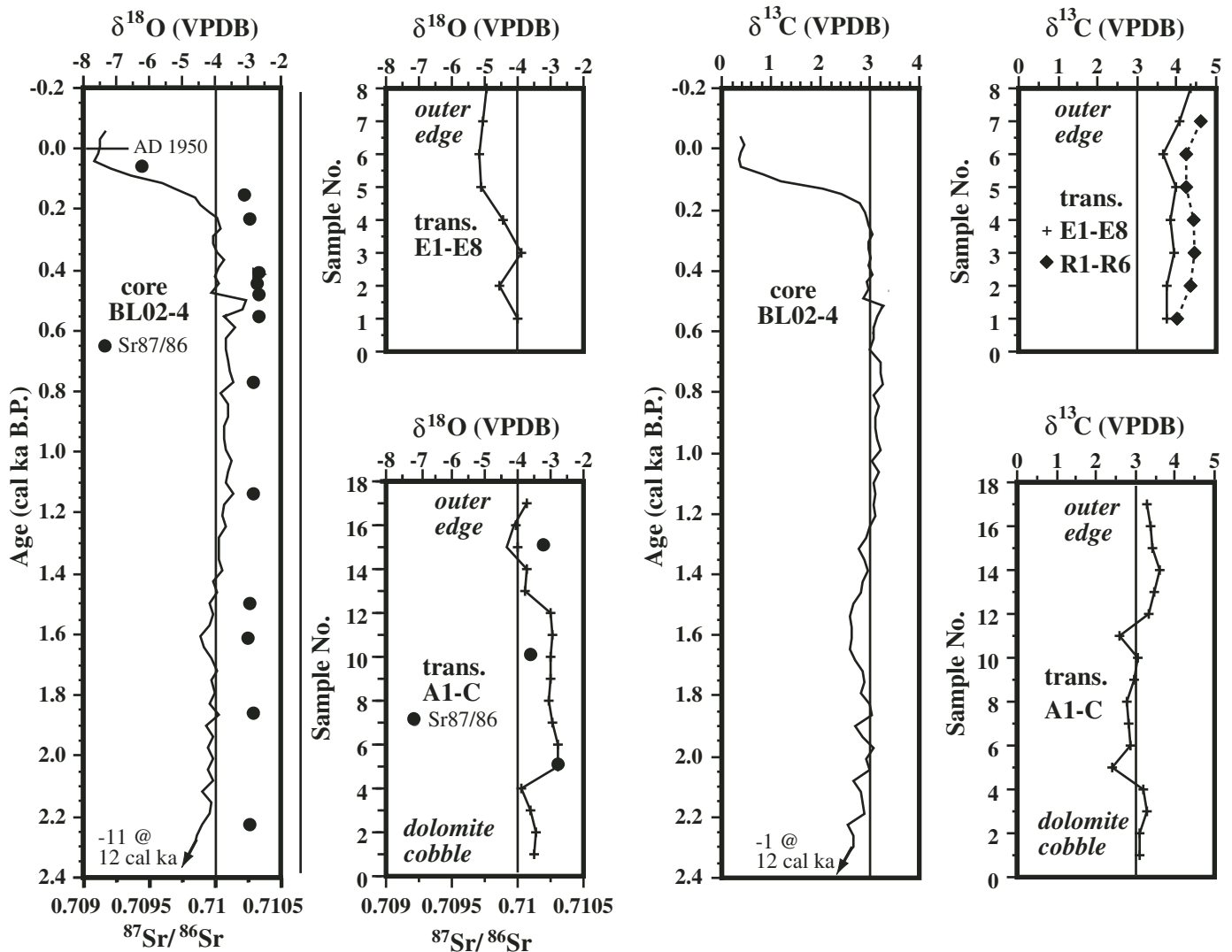


Figure 8. Profiles of $\delta^{18}O$ and $\delta^{13}C$ for core BL02-4 and microbialite transects A1–C and E1–E8 (see Fig. 7 for locations of the transects on the 2002 microbialite sample). Values of $^{87}Sr/^{86}Sr$ for core BL02-4 and microbialite transect A1–C are shown as closed circles on the profiles of $\delta^{18}O$. VPDB—Vienna Pee Dee Belemnite.
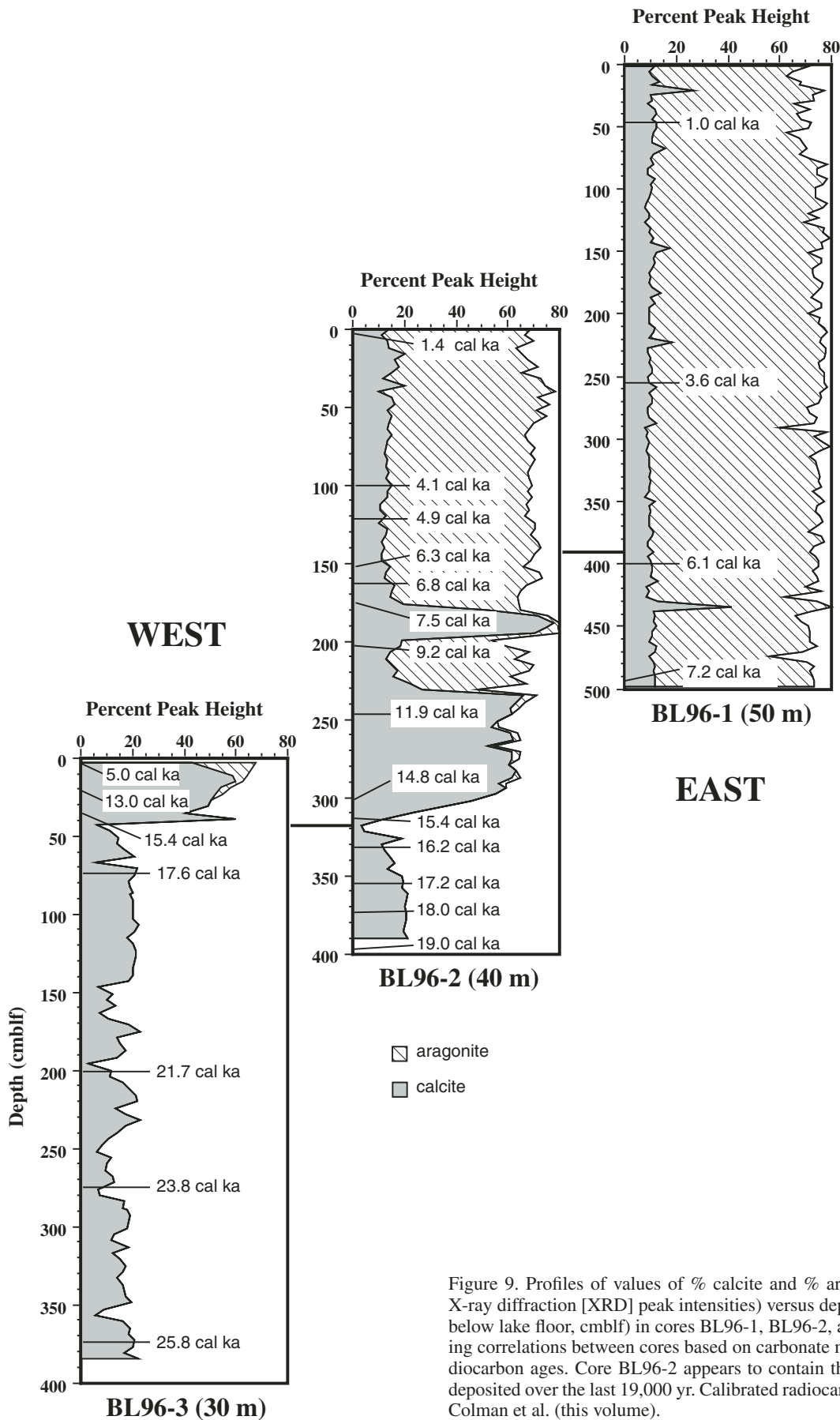
Figure 9. Profiles of values of % calcite and % aragonite (based on X-ray diffraction [XRD] peak intensities) versus depth (in centimeters below lake floor, cmblf) in cores BL96-1, BL96-2, and BL96-3 showing correlations between cores based on carbonate mineralogy and radiocarbon ages. Core BL96-2 appears to contain the entire sequence deposited over the last 19,000 yr. Calibrated radiocarbon ages are from Colman et al. (this volume).

### Isotope Geochemistry and Development of Bear Lake over the Last 26 k.y.

According to the strontium isotopic composition of pre- and post-diversion carbonates (Fig. 6), the strontium isotopic composition of twentieth-century carbonates in Bear Lake sediments can be explained by mixing of Bear River water and some pre-diversion lake water having an isotopic composition defined by pre-diversion, late Holocene endogenic carbonate (aragonite; Fig. 6). Therefore, if Bear River entered Bear Lake in the past, the C, O, and Sr isotopic composition of endogenic carbonates should, as a first approximation, provide evidence of such an event. In particular, the very low values of $^{87}Sr/^{86}Sr$ in Bear River water (0.7085) should be an excellent tracer for times when Bear River water entered Bear Lake (Bouchard et al., 1998). When Bear Lake was not connected to Bear River, the hydrology of the lake can be assumed, as a first approximation, to be the main control on isotopic composition. The balance between precipitation and evaporation is perhaps the largest hydrologic factor, but, as discussed earlier, there is an unknown groundwater source that may have varied considerably over time (Dean et al., 2007).

Values of $\delta^{18}O$ and $\delta^{13}C$ in calcite in the calcareous, quartz-rich LGI sediments (unit 1) are lower (Fig. 10) than in Holocene carbonate and in carbonate collected in sediment traps (Fig. 6). The LGI sediments are even more depleted than carbonate rocks

(limestones and dolomites) in the Bear River Range to the west of the lake and in the Bear Lake Plateau east of the lake, as judged by an average value of $\delta^{18}O$ of $-8.2 \pm 2.7‰$ and an average value of $\delta^{13}C$ of $0.4 \pm 1.4‰$ in nine samples of those rocks (Bright, this volume; Fig. 11A). However, the LGI sediments are not as depleted as the fine fractions (<63μm) of stream sediments, particularly those in west-side streams (Table 3, Fig. 11A). Apparently the isotopic composition of bulk carbonate in stream sediments does not reflect the isotopic composition of carbonate bedrock. This is because stream sediments contain grains of endogenic calcite (tufa), which probably formed in the streams. Isotope analyses of several hand-picked grains show that they are depleted in both $^{18}O$ and $^{13}C$ (Kaufman et al., this volume; Fig. 11A). It appears that modern stream sediments are mixtures of detritus from carbonate bedrock and endogenic tufa (Fig. 11A). There is no evidence that Bear River and other streams were producing endogenic tufa during the LGI, but the isotopic composition of LGI bulk carbonate (Fig. 10) suggests that some isotopically depleted carbonate was being added to detrital bedrock carbonate, although nine samples are hardly representative of the stratigraphic and geographic distribution of the isotopic composition of carbonate rocks in the Bear River Range and Bear Lake Plateau.

The timing of the isotope excursions might be better defined by the higher resolution isotope data from core BL02-4 from
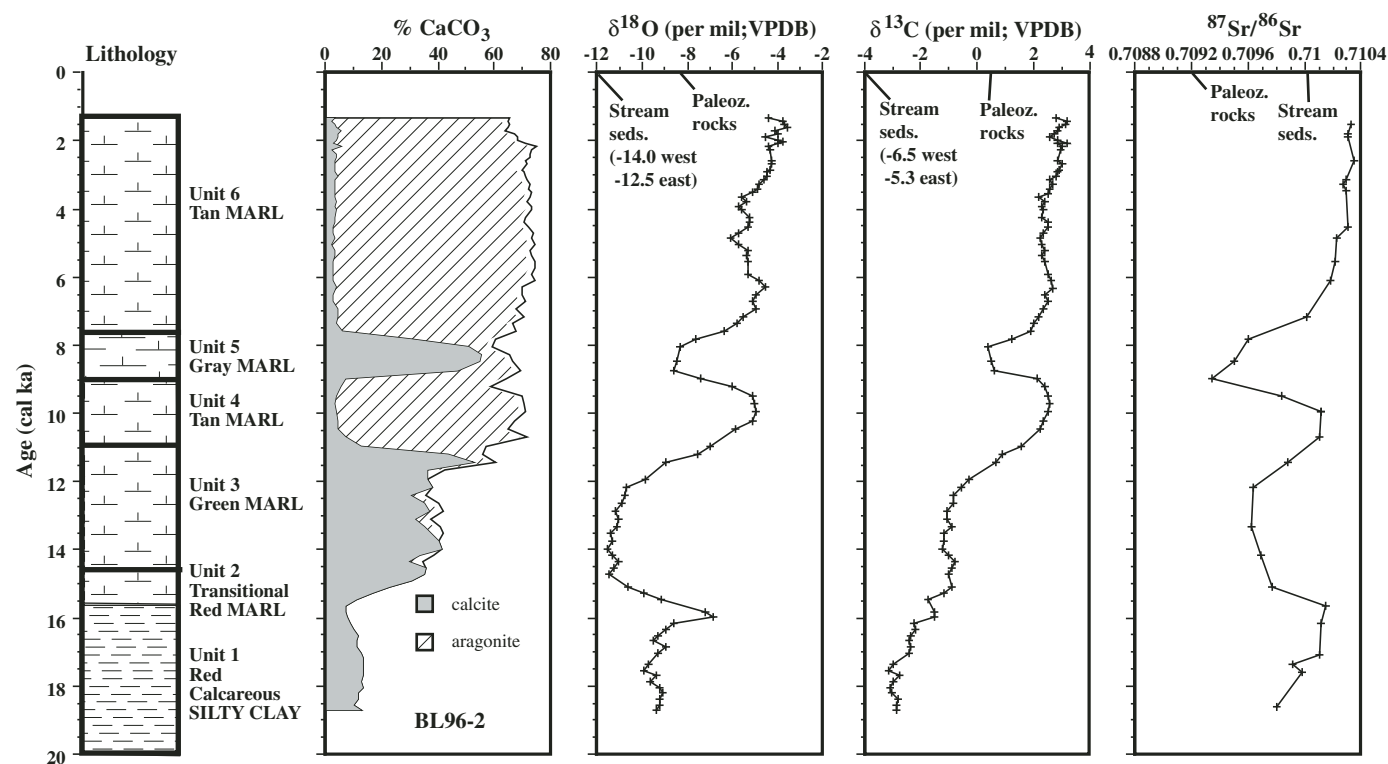


Figure 10. Lithology and profiles of percent CaCO₃ as aragonite and calcite (see text for method of partitioning total CaCO₃ into aragonite and calcite) and values of $\delta^{18}O$, $\delta^{13}C$, and $^{87}Sr/^{86}Sr$ in bulk carbonate versus age in core BL96-2. Values in stream sediments (Table 3) and Paleozoic carbonate rocks (Bright, this volume) are shown at the top of each profile. The age model was developed by Colman et al. (this volume).
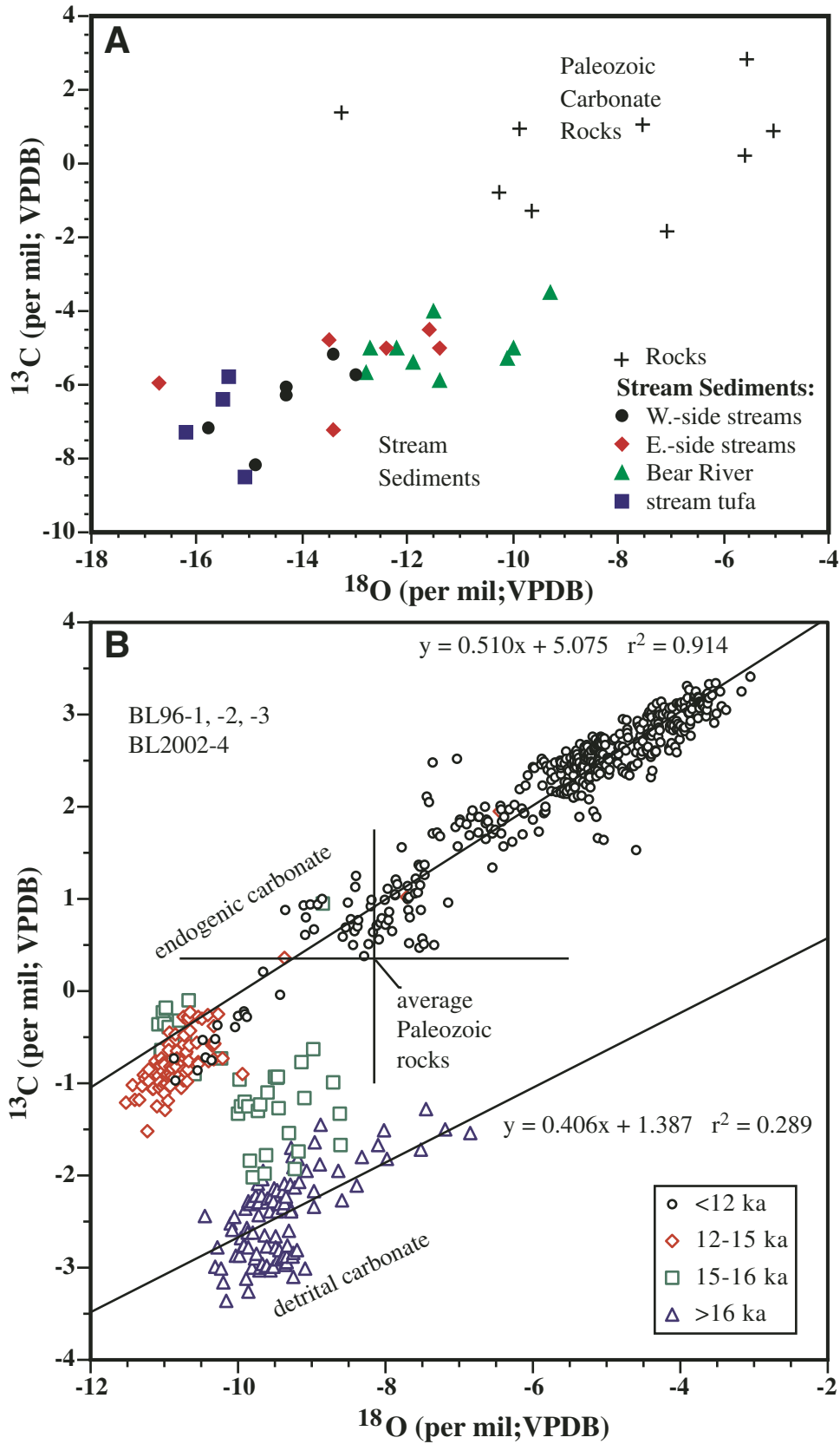
Figure 11. (A) Cross plot of values of $\delta^{18}O$ and $\delta^{13}C$ in samples from Paleozoic bedrock in the Bear River Range, in the <63 μm fractions of stream sediments, and in hand-picked samples of endogenic stream calcite (tufa). Stream-sediment values are from Table 3. Rock values are from Bright (this volume). (B) Cross plot of values of $\delta^{18}O$ and $\delta^{13}C$ in samples from cores BL96-1, -2, -3, and BL02-4. Horizontal and vertical bars represent the average and one standard deviation of the carbonate-rock values in A. VPDB—Vienna Pee Dee Belemnite.

shallower water at the north end of the lake (Figs. 1A and 12). Measurements of oxygen and carbon isotopes in this core were made every centimeter, whereas those in core BL96-2 were made every 4 cm (Dean et al., 2006). Unfortunately, the isotope records in core BL02-4 are interrupted by two hiatuses represented by shell layers. The shell layers mark the culmination of lower lake levels of as much as 25 m (Smoot and Rosenbaum, this volume). Nevertheless, the higher-resolution records should help to define the timing of isotope excursions better.

Between 15.5 and 14.5 cal ka (unit 2), values of $\delta^{18}O$ in core BL96-2 declined to −11‰, the lowest value in the entire glacial-Holocene sequence, where they remained for ~4 k.y. (Fig. 10). The decline in core BL02-4 was more abrupt, occurring ca. 16 cal ka, and there is no peak in values of $\delta^{18}O$ between 16 and 15 cal ka (Fig. 12) as there is in core BL96-2 (Fig. 10). If the peak in $\delta^{18}O$ in core BL96-2 from the center of the lake is related to the disconnection of Bear River from Bear Lake, and to content of detrital material from Bear Lake catchment streams, this event was not recorded at the north end of the lake. These lower values of $\delta^{18}O$ in sediments deposited between 14.5 and 12 cal ka (unit 3 in Fig. 10;

Fig. 12) might be due to greater influx of Bear River water, but the influx of detrital clastic material was decreasing at that time (Rosenbaum et al., this volume), and there are no corresponding decreases in values of $\delta^{13}C$ (Fig. 10). The low values of $\delta^{18}O$ in calcite deposited between 14.5 and 12 cal ka are therefore most likely due to continued abundance of cold-season precipitation (winter snow).

For most of the sediments in cores BL96-1, -2, -3, and BL02-4, there is a strong covariation between values of $\delta^{18}O$ and $\delta^{13}C$ (Fig. 11B). There is a positive covariation between $\delta^{18}O$ and $\delta^{13}C$ in samples of endogenic carbonate, regardless of mineralogy, younger than ca. 15 cal ka (Fig. 11B), but there are some trends in samples of older sediments that may provide clues for the isotope excursions in the transition between the detrital sediments of the LGI (>16 cal ka) and the endogenic carbonate sediments deposited after 15 cal ka.

A positive covariance between values of $\delta^{18}O$ and $\delta^{13}C$ 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, 1994; Talbot and Kelts, 1990; Li and Ku, 1997). Covariant trends may have remarkably long-term persistence through major
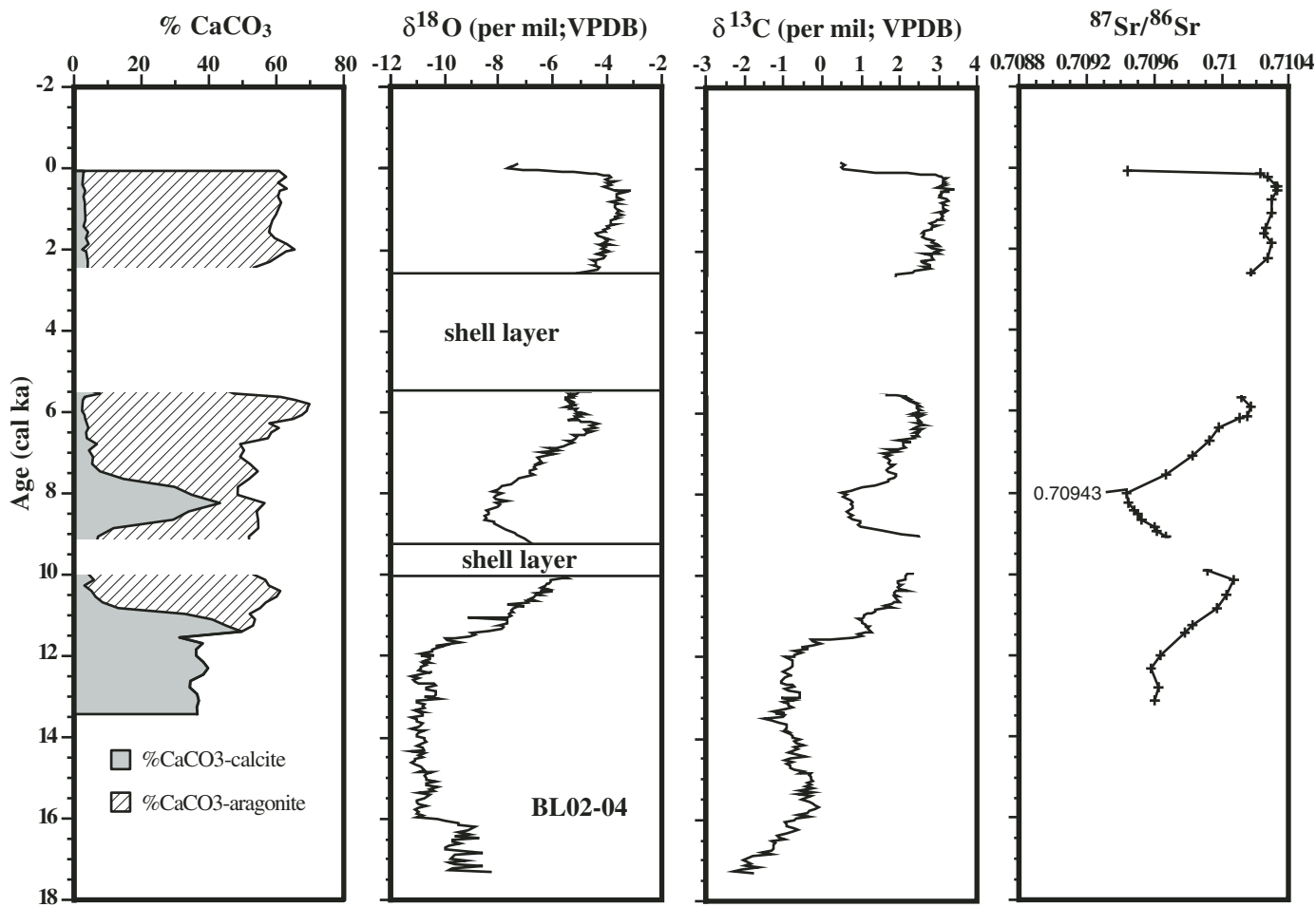


Figure 12. Profiles of percent CaCO₃ as aragonite and calcite (see text for method of calculation), and of $\delta^{18}O$, $\delta^{13}C$, and $^{87}Sr/^{86}Sr$ in bulk carbonate versus age (0–20 cal ka) in core BL02-4. The age model was developed by Colman et al. (this volume). VPDB—Vienna Pee Dee Belemnite.

environmental changes (Talbot, 1990), and this certainly applies to the last 15 k.y. in Bear Lake. Sediments deposited before 16 cal ka in BL02-4 and BL96-2, have their own covariant isotope trend, but with much lower values of $\delta^{13}C$ and much weaker isotope covariance than those in the trend for younger sediments (Fig. 11B). These sediments are considered to represent one end-member—sediments that contain glacial flour from the Uinta Mountains (Dean et al., 2006; Rosenbaum and Heil, this volume). However, present-day stream sediments in the Bear River in the Uinta Mountains do not contain carbonate (Table 3), so the carbonate in sediments deposited before 16 cal ka must have been derived from Bear River drainage north of the Uintas in Wyoming and Utah (Fig. 1B) and from the local catchment.

The sediments in BL96-2 and BL2002-4 deposited between 16 and 15 cal ka have a slight negative covariant trend (Fig. 11B). This trend goes from low values of $\delta^{13}C$ (~−2‰) in sediments deposited before 16 cal ka to higher values of $\delta^{13}C$ (~−0.5‰) in sediments deposited between 15.6 and 15.0 cal ka, values that are typical of the endogenic low-Mg calcite deposited between 15 and 12 cal ka (Fig. 11B). Magnetic properties suggest that between 20 and 16 cal ka there was a transition from predominantly glacial flour from the Uinta Mountains in the detrital fraction to predominantly detrital material derived first from Bear River north of the Uintas and then (ca. 16.5 cal ka) from the local Bear Lake catchment (Rosenbaum and Heil, this volume). However, this does not explain the negative isotope covariation between 16 and 15 cal ka, because Paleozoic carbonate rocks in the Bear Lake catchment are more enriched in both $^{18}O$ and $^{13}C$ (Fig. 11A), and replacing upper Bear River detrital carbonate (triangles in Fig. 11B) with debris from carbonate rocks would result in a positive covariation. The more likely explanation for the slight negative covariant trend between 16 cal ka and 15 cal ka is a mixing of detrital carbonate in sediments from the upper Bear River drainage north of the Uintas (triangles in Fig. 11B) with endogenic carbonate having higher values of $\delta^{13}C$, (diamonds in Fig. 11B). The endogenic calcite deposited between 15 and 12 cal ka begins to show a positive covariant trend but the variation is small (~1‰ for both $\delta^{18}O$ and $\delta^{13}C$; diamonds in Fig. 11B).

The covariant trend of enrichment in both $^{18}O$ and $^{13}C$ in endogenic carbonate younger than 15 cal ka, regardless of mineralogy (low-Mg calcite or aragonite), with a high degree of correlation, is the result of evaporation from a progressively more closed-basin lake with a long residence time as influx of the Bear River greatly diminished. Evaporation and warming of the lake water would lead to loss of $CO_2$ that is depleted in $^{13}C$. Resulting carbon-isotope fractionation between lake water and dissolved inorganic carbon (DIC) and $CO_2$ would result in $^{13}C$-enriched waters and carbonate minerals precipitated from the water (Li and Ku, 1997). In a lake with a long residence time, prolonged burial of $^{13}C$-depleted OC could also lead to elevated values of $\delta^{13}C$ in DIC (McKenzie, 1985; Hollander and McKenzie, 1991; Li and Ku, 1997). If this is true, then the slight positive covariant trend in carbonate deposited before 16 cal ka (triangles in Fig. 11B) suggests that the isotopic composition of that carbonate also was influenced by evaporation,

which implies that some of that carbonate is endogenic mixed with carbonate from the upper Bear River drainage.

At about 12 cal ka, values of both $\delta^{18}O$ and $\delta^{13}C$, but especially $\delta^{18}O$, began to increase rapidly (Figs. 10 and 12), most likely due to increased evaporation and/or reduced influx of Bear River water. Between 12 and 10 cal ka, values of $\delta^{18}O$ increased by 6‰ (−11‰ to −5‰). Values of $^{87}Sr/^{86}Sr$ also increased during this interval (Figs. 10 and 12). Evaporation and reduction in Bear River inflow increased the total dissolved solids and $Mg^{2+}$:$Ca^{2+}$ in the lake water, and by 11 cal ka, a critical value of $Mg^{2+}$:$Ca^{2+}$ was reached and calcite deposition was rapidly replaced by aragonite deposition (unit 4 in Fig. 10; Fig. 12). As discussed earlier, Müller et al. (1972) suggested that aragonite precipitates only in lakes having an $Mg^{2+}$:$Ca^{2+}$ of >12. I suggest that by 11 cal ka Bear River was entirely detached from Bear Lake. By 10 cal ka, the mineralogy and isotopic composition of bulk carbonate were approaching values that would be typical of the middle Holocene (6–4 cal ka; Fig. 10). This increase in isotopic values between 12 and 10 cal ka was accompanied by a drop in lake level of more than 25 m below modern lake level, judging from grain-size data and other sedimentological evidence (Smoot and Rosenbaum, this volume). The lower shell layer in core BL02-4 represents the culmination of this lowstand (Fig. 12).

At about 9 cal ka, values of $\delta^{18}O$, $\delta^{13}C$, and $^{87}Sr/^{86}Sr$ decreased considerably (Figs. 10 and 12), suggesting that Bear River water was again entering Bear Lake. Values of $\delta^{18}O$ were at a minimum between 9 and 8 cal ka. Lake level went back up by at least 25 m (Smoot and Rosenbaum, this volume), and the rise was accompanied by a decrease in salinity (decrease in values of $\delta^{18}O$; Figs. 10 and 12) and, presumably, a decrease in $Mg^{2+}$:$Ca^{2+}$ that resulted in cessation of aragonite precipitation and a return of calcite precipitation (unit 5 in Fig. 10; Fig. 12). The decrease in $^{87}Sr/^{86}Sr$ began a little earlier (9.5 cal ka in both cores) than the decrease in $\delta^{18}O$. The $^{87}Sr/^{86}Sr$ data from core BL02-4 suggest that the Sr isotope decrease may have begun as early as 10 cal ka, below the shell layer in core BL02-04 (Fig. 12), but it is hard to tell because of the shell layer. Values of $^{87}Sr/^{86}Sr$ reached a minimum ca. 9 cal ka in core BL96-2 and ca. 8.2 cal ka in core BL2002-4. The lead of the Sr isotope signal over the O isotope signal probably is because the Sr concentration in the lake today is much lower than that of Bear River, and likely was even lower prior to Bear River diversion. Therefore, changes in Sr isotopic composition of the low-Sr-concentration lake water were not as well buffered as changes in O isotopic composition, which had to affect a much larger oxygen reservoir.

Calcite precipitation lasted ~1000 yr (unit 5 in BL96-2, Fig. 10; upper calcite layer in BL02-4, Fig. 12). Unlike the rapid reversal in the Sr isotope excursion, the well-buffered C and O isotopic composition remained at low values for ~1000 yr before increasing (Figs. 10 and 12). The entire negative Sr and O isotope excursions lasted ~2000 yr. As the river again abandoned the lake, lake level rapidly declined, $Mg^{2+}$:$Ca^{2+}$ of the lake increased, and by 7.5 cal ka aragonite was again forming (unit 6 in Fig. 10; Fig. 12). The mineralogy of carbonate in Bear Lake sediments

has been relatively constant for the last 7 cal k.y. (Fig. 10), but values of δ¹⁸O and δ¹³C continued to increase slightly (Figs. 10 and 12), suggesting that the salinity continued to increase after Bear River was again disconnected from Bear Lake.

## Carbonate Deposition in Bear Lake during the Last Two Glacial-Interglacial Cycles

The lithology and mineralogy of the sediments collected over the last 26,000 yr, as recovered in piston cores, consists of red, calcareous silty clay deposited between 26 and 16 cal ka overlain by marl deposited during the last 16,000 yr, in which the mineralogy of the carbonate alternates between low-Mg calcite and aragonite with minor amounts of dolomite and quartz. As

discussed earlier, the Holocene environmental history of the Bear Lake is complex (Dean et al., 2006), and the alternation of calcite and aragonite is an important part of that history.

The visual core descriptions of cores recovered from holes BL00-1D and -1E indicate that the top of the recovered section contains the same massive, tan, bioturbated aragonitic marl and underlying red silty clay (Fig. 13, lithologic units 6 and 7; http://www.ngdc.noaa.gov/mgg/curator/laccore.html) that were recovered in piston cores. The overall CaCO₃ content of the marl (unit 7) is high, ranging from 60% to 80% with an average of ~70%, mostly in the form of aragonite (Fig. 13). The OC content also is high when compared with the underlying red clay, ranging mostly between 1% and 2%, but higher values commonly occur in older sediments (Fig. 13).
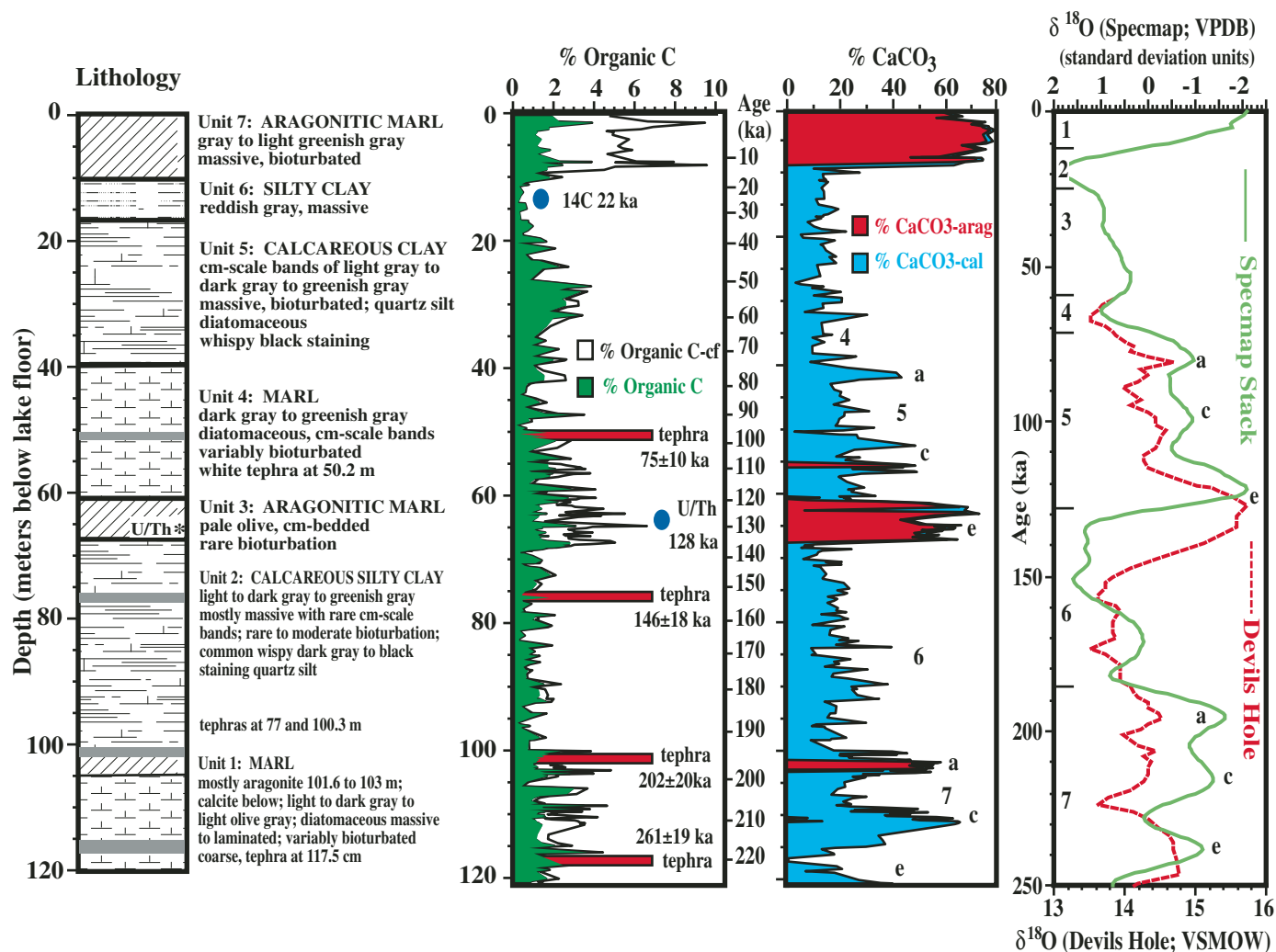


Figure 13. Lithology of sediment; profiles of percentages (calculated from XRD peak intensities) of dolomite and quartz; profiles of percent organic carbon (raw and carbonate-free, cf) and percent CaCO₃ as aragonite and calcite (see text for method of calculation) versus depth in samples from cores from holes BL00-1D and -1E. Radiocarbon age, U/Th age, and ages of tephras on the profile for organic carbon are from Colman et al. (2006). Profiles on the right are of values of δ¹⁸O versus age in the SPECMAP stacked marine benthic foraminifera record (Martinson et al., 1987), and in a calcite vein in Devils Hole, Nevada (Winograd et al., 1992). Numbers (1–7) next to the time scale on the δ¹⁸O profile indicate marine oxygen isotope stages 1–7. VPDB—Vienna Pee Dee Belemnite; VSMOW—Vienna Standard Mean Ocean Water. XRD—x-ray diffraction.

*Dean*

On the basis of visual core descriptions and smear-slide analyses, most of the sediments recovered below these top two units were initially described as calcareous clay or silty clay exhibiting some wispy laminations, and some centimeter-scale bedding cycles, distinguished mainly by the degree and style of bioturbation (Fig. 13). Subsequent carbon analyses showed that the clay has highly variable amounts of OC, but generally in the range of 1%–3% (Fig. 13). Variations in amounts of fine-grained sulfides and OC undoubtedly account for most of the variations in color (shades of gray) observed on fresh-cut surfaces. The non-carbonate fraction consists mainly of quartz, with minor amounts of dolomite and feldspar (see Kaufman et al., this volume, for further discussion of the non-carbonate mineralogy). Although several lithologic units contain as much quartz as in the red silty clay deposited during the LGI (~75%), there was no red coloration mentioned in the visual core descriptions, suggesting either that the lake did not receive large quantities of hematite-rich sediment from the Uinta Mountains or that the hematite pigment was destroyed by reducing conditions in the sediments.

The OC concentration in unit 6 is the lowest sustained OC concentration (<0.5%) in the entire record (Fig. 13); few other intervals have OC concentrations this low. In order to determine the effect of $CaCO_3$ dilution on the OC content, OC concentrations were calculated on a carbonate-free (cf) basis (Fig. 13). In the carbonate-rich intervals, the cf concentrations of OC are considerably higher than in bulk sediment. In unit 7, the average bulk-sediment OC concentration is 1.78%, and that on a cf-basis is 5.81%. In units 4 and 3, the average bulk-sediment OC concentration is 1.49%, and that on a cf-basis is 3.18%. In unit 1, the average bulk-sediment OC concentration is 1.36%, and that on a cf-basis is 2.38%. Using OC content as a semiquantitative measure of organic productivity, these results suggest that organic productivity was highest in Bear Lake during the high-carbonate, dry interglacial intervals that have been correlated with marine oxygen isotope stages (MIS) 1, 5, and 7 (Fig. 13; Bright et al., 2006; Colman et al., 2006; Kaufman et al., this volume). The sharpest increase in productivity occurred between units 6 and 7 (Fig. 13).

$CaCO_3$ contents of sediments below lithologic unit 7 are mostly in the range of 20%–40% (Fig. 13), but there are several carbonate-rich (>50% $CaCO_3$) intervals below unit 7. To distinguish high- and low-carbonate intervals, sediments with greater than 20% $CaCO_3$ are here called marls (Fig. 13, lithologic units 1, 3, and 4), and sediments with less than 20% $CaCO_3$ are called calcareous clay or silty clay. The interval from 60 to 67 meters below lake floor (mblf; lithologic unit 3) contains as much as 70% $CaCO_3$, and XRD indicates that most of this is in the form of aragonite (Fig. 13). This interval has been correlated with MIS 5e (Fig. 13; Bright et al., 2006; Colman et al., 2006; Kaufman et al., this volume). Several intervals containing >50% $CaCO_3$ occur between 100 and 115 mblf (Fig. 13, lithologic unit 1) with calcite as the dominant carbonate mineral. This high-carbonate interval has been correlated with MIS 7 (Fig. 13; Bright et al., 2006; Colman et al., 2006; Kaufman et al., this volume). In hole BL00-1E, the high-carbonate peak correlated with MIS 7a contains mostly aragonite (Fig. 13). The peak corre-

lated with MIS 7c contains very minor amounts of aragonite. Below 120 mblf, the $CaCO_3$ content increases rapidly downward (Fig. 13). Perhaps a few more meters of sediment at the base of hole BL00-1E would have obtained a well-defined carbonate peak, possibly aragonite, coincident with MIS 7e.

The striking resemblance of the smoothed $CaCO_3$ carbonate curve to the SPECMAP stacked oxygen isotope curve (Martinson et al., 1987), and to the oxygen isotope curve from a calcite vein in Devils Hole, Nevada (Fig. 13; Winograd et al., 1992) was pointed out by Colman et al. (2006) who used the Devils Hole chronology to "tune" the age model for the BL00-1 cores (see also Kaufman et al., this volume). The correlation of high-carbonate, aragonitic intervals with interglacials has been corroborated by high values of $\delta^{18}O$ and $Sr^{87}/Sr^{86}$ in these intervals (Bright et al., 2006; Kaufman et al., this volume).

Owens Lake, southeastern California, was hydrologically closed, saline, alkaline and highly productive during interglacials when it was disconnected from the Owens River (Bischoff et al., 1997; Menking et al., 1997). The low-carbonate intervals in Bear Lake correlate with global glacial conditions and represent periods when Bear River was directly connected to Bear Lake. Owens Lake was hydrologically open during glacials, when influx from the Owens River was augmented by meltwaters from Sierran glaciers. During the dry interglacials, Bear River was not connected to Bear Lake, but a considerable amount of freshwater influx must have occurred, presumably as groundwater base flow to other streams entering the lake; otherwise, the salinity of the lake would have gone more evaporative, beyond aragonite precipitation, precipitating evaporite minerals (e.g., gypsum and halite; Dean et al., 2007) and producing high values of $\delta^{18}O$. During deposition of aragonitic unit 3 in BL00-1 (Fig. 13), values of $\delta^{18}O$ in bulk carbonate were as high as in aragonite deposited during the late Holocene in BL96-2 and BL02-4 (~–4‰, Figs. 10 and 12; Bright et al., 2006; Kaufman et al., this volume). Values of $\delta^{18}O$ in unit 1 bulk carbonate (MIS-7) only got as high as –8‰ (Bright et al., 2006; Kaufman et al., this volume), suggesting that conditions during deposition of unit 1 were not as evaporative as during deposition of units 3 and 7. Similarly, $^{87}Sr/^{86}Sr$ values in bulk carbonate in unit 3 were as high as in unit 7 (~0.7102), but values in bulk sediment in unit 1 were much lower (~0.7100, Kaufman et al., this volume), indicating more influx of Bear River water with a low $^{87}Sr/^{86}Sr$ (~0.7085). What climatic conditions were ultimately driving wet glacial and dry interglacial conditions in Bear Lake and Owens Lake? The answer has to be in atmospheric circulation over the northeastern Pacific Ocean.

## The Bear Lake Record and Glacial/Interglacial Climates of Western United States and NE Pacific

The climate history of the last glacial/interglacial cycle in the western United States and adjacent northeastern Pacific Ocean is fairly well known from both continental and marine records. Most of these records have been obtained from conventional piston coring in lakes and the Pacific margin of the United States,

or from outcrops. Continuous marine records of older glacial/interglacial cycles only recently have been revealed by Ocean Drilling Program (ODP) drilling on the Pacific margin (ODP Leg 167; Lyle et al., 2000). The Bear Lake record is the only long, continuous, continental record from the western United States collected from an extant lake.

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 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 °C 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 was greatest under the axis of the polar jet, decreased abruptly south of the axis, and decreased less abruptly north of the axis. Consequently, the Great Basin was wetter than present during the LGI, and the Pacific Northwest was drier (Barnosky et al., 1987; Thompson et al., 1993). Sediments deposited during the LGI in Owens Lake, California, at a latitude of 36.5°N contain an ostracode assemblage dominated by *Cytherissa lacustrus*, which indicates very cold, stable limnoclimatic conditions (Bradbury and Forester, 2002). Today *C. lacustrus* lives in cold lakes in boreal forests of Canada and Alaska, and 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 by Hostetler et al. (1994) indicate that the presence of the lakes had a considerable influence on the precipitation over these large lake basins. They estimate that at 20 cal ka precipitation in the Bonneville Basin (including the lake) was 3% greater in January and 38% greater in July than it would have been without the lake, due to lake-atmosphere feedbacks.

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 ca. 16 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. By that time (ca. 16 cal ka) the flow in the Bear River had decreased, red clay deposition ceased (top of unit 6), and endogenic carbonate began to increase (increase in $CaCO_3$ in core BL96-2, Fig. 10).

We know from recent changes in levels of lakes that are the remnants of the larger glacial lakes in the Great Basin (e.g., Mono, Pyramid, Walker, and Great Salt Lakes) that lake levels go up when winter precipitation is greater than normal, and go down when winter precipitation is lower than normal (e.g., Benson and Thompson, 1987). The sediments of Pleistocene Lake Manix record major lacustrine phases, interpreted to represent high lake levels, that are coincident with MIS 4 and 6, and lowstands are coincident with MIS 5 and 7 (Jefferson, 2003).

Winter precipitation comes from the Pacific Ocean, and increased winter precipitation occurs by enhanced cyclonic flow around the Aleutian low-pressure system when the polar jet stream moves south of its normal position. Today, this condition exists only during the winter, but during the LGI it may have prevailed throughout the year (e.g., Kutzbach, 1987). When the polar jet stream was in this southerly position, and the Aleutian Low dominated atmospheric circulation over the eastern North Pacific, the southwestern United States was wet (e.g., Van Devender et al., 1987) and the Pacific Northwest was dry (e.g., Barnosky et al., 1987).

A major reconfiguration of Pacific atmospheric circulation and the dominance of cyclonic circulation driven by the Aleutian Low also would have had a major effect on surface oceanic circulation in the North Pacific. Today, surface circulation in the North Pacific is dominated by the North Pacific subtropical gyre. The California Current is the southward flowing, wind-driven, eastern limb of the North Pacific subtropical gyre formed by the divergence of the West Wind Drift on the western margin of North America. This oceanic circulation is controlled by atmospheric-pressure systems of the North Pacific and western North America (the North Pacific High, the Aleutian Low, and the North American Low; Fig. 14). The seasonal strengths and positions of these pressure systems not only generate the weather and climate of the western United States (Huyer and Kosro, 1987; Strub et al., 1987; Thomas and Strub, 1990; Abbott and Barksdale, 1991), but also are part of the atmospheric teleconnections that stretch across the Northern Hemisphere (e.g., Namias et al., 1988). Changes in the position and strength of these atmospheric systems that occurred as a result of the change from global glacial to interglacial conditions had large influences on the climate of the western United States.

Today, the climate of southern Oregon and California is characterized in the spring and summer by strong, persistent, northwesterly winds generated by the juxtaposition of the North Pacific High and North American Low. The average summer position of the subtropical North Pacific High is centered at ~45°N and the average summer position of the strong North American Low, which is maintained by intense heating of the Great Basin, is centered at ~38°N (Fig. 14A). The summer polar jet stream typically is located north of 50°N. Winters are influenced by a weakened North American Low, the migration of the North Pacific High south of 30°N, and the migration of the polar jet stream and associated Aleutian Low to an average position of ~45°N (Fig. 14B). Winters are typically mild, wet, and stormy, with zonal westerly winds.

Temperatures in the continental interior were higher in the Holocene than during the LGI, but what about sea-surface temperatures (SSTs) in the North Pacific? Recent studies suggest that SSTs may have been 3–5 °C cooler than present during the LGI off Oregon and California (e.g., Ortiz et al., 1997; Mix et al.,

1999; Herbert et al., 2001; Pisias et al., 2001; Trend-Staid and Prell, 2002; Barron et al., 2003). Results from analysis of recent ODP cores from the California margin suggest that this glacial-interglacial contrast in SST may have been even greater between MIS 6 and MIS 5 (Lyle et al., 2001).

As discussed earlier, the glacial-to-interglacial (Holocene) transition in Bear Lake in terms of carbonate mineralogy is represented in great detail with excellent radiocarbon dating in cores BL96-2 and BL02-04. The details of this transition are difficult



Figure 14. Generalized surface atmospheric pressure cells in the eastern North Pacific and adjacent North America during summer (A) and winter (B).

to see in data from site BL00-1 (Fig. 13) because of the coarser sampling interval. I assume that a similar scenario occurred at the transitions to earlier interglacial high-carbonate, aragonitic intervals equivalent to MIS-5 and -7 (Fig. 13), i.e., decreasing precipitation, increasing evaporation:precipitation ratio, increasing salinity leading to carbonate precipitation, first as calcite then, at highest salinity, as aragonite. If this interpretation is correct, it would appear that carbonate precipitation remained more in the calcite-stability environment during the MIS-5 carbonate "event" and even more so during the MIS-7 "event" (Fig. 13).

These high-carbonate, aragonitic interglacial intervals are interpreted to coincide with warm continental climates and warm SSTs recorded in recent ODP cores along the Pacific margin (Lyle et al., 2000; Lyle et al., 2001). This implies that the subtropical North Pacific High and North American Low that today bring warm-dry conditions during the summer (Fig. 14A) were more permanent features of North Pacific circulation during interglacials. During the last glacial interval and other stadials, the Aleutian low-pressure system, presently dominant during the winter (Fig. 14B), was a more permanent feature of North Pacific circulation, producing storms and precipitation that increased the levels of Lakes Bonneville and Lahontan and increased the flow of Bear River. These climatic patterns over the North Pacific are recorded in Bear Lake by river-borne detrital clastic sediments deposited during glacial periods and carbonate sediments, dominated by aragonite, during the brief, more arid interglacial periods.

## SUMMARY

1. Time-marking sediment traps placed 10 m below the surface of the lake and 2 m above the bottom of the lake show that high-Mg calcite precipitates from the surface waters of the lake, but aragonite is the dominant mineral that accumulates on the lake floor, along with lesser amounts of high-Mg calcite, low-Mg calcite, dolomite, and quartz.

2. The chemical and isotopic compositions of the high-Mg calcite that forms in the surface waters of the lake today are different from those of the aragonitic sediments deposited in Bear Lake prior to the diversion of Bear River into Bear Lake in the early twentieth century. The aragonite collected in the deep traps did not precipitate from present-day Bear Lake water, but is reworked aragonite from water depths shallower than ~30 m that is at least 50 yr old and possibly older. The diversion of Bear River into Bear Lake created an experiment in which the pre- and post-diversion chemical composition, isotopic composition, and mineralogy of bulk carbonate allows us to distinguish between modern and reworked endogenic carbonate.

3. Radiocarbon ages on samples from the top of an extensive, aragonite-cemented microbialite mound in the southwestern part of the lake suggest that the mound formed during the middle to late Holocene (6500–3000 yr ago). However, O and C isotope values in transects of samples from a thick carbonate rind on one microbialite have values that occur only in sediments
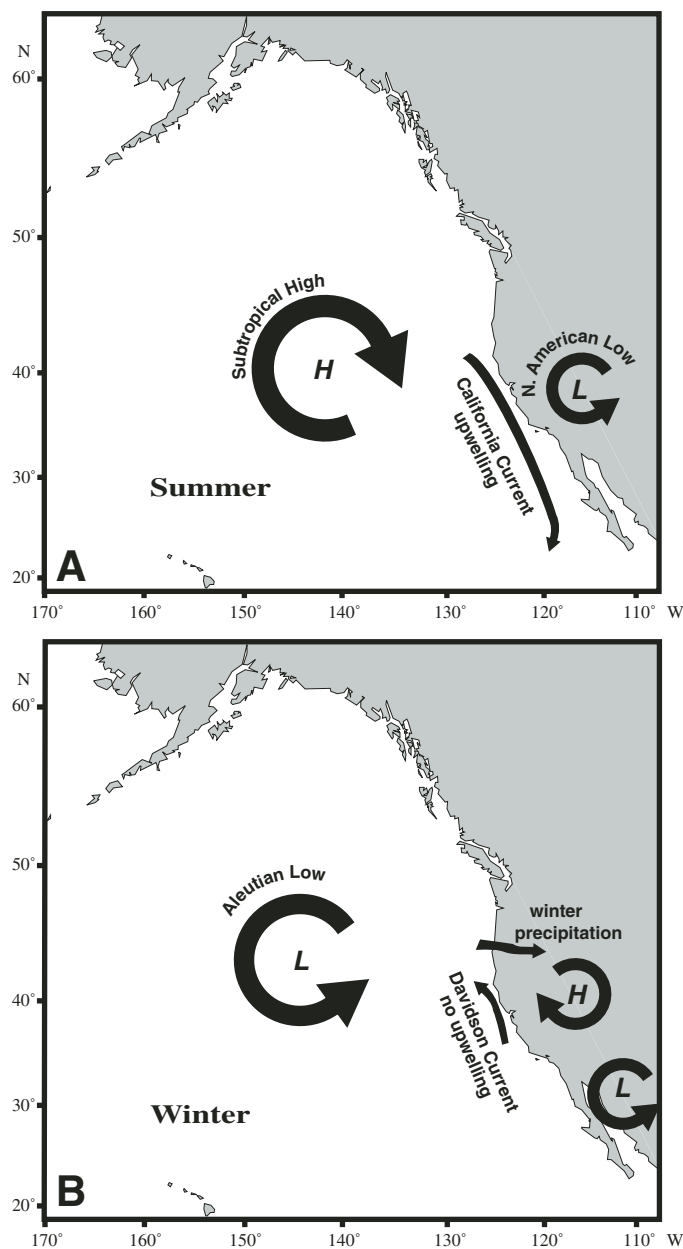
in cores that were deposited during the last several thousand years. This indicates that the top of the mound began forming several thousand years ago but stopped precipitating aragonite after Bear River diversion. The mound may have originated by $CO_2$-charged groundwater entering the lake, discharging $CO_2$, and precipitating $CaCO_3$, mediated by cyanobacteria and other microbes in much the same way that microbialite mounds or tufa mounds form in other lakes today.

4. During the last glacial interval, when increased Pacific moisture filled the large lakes of the Great Basin, the Bear River entered Bear Lake and deposited red, calcareous, silty clay. When Pacific moisture was greatly reduced ~16,000 yr ago, Bear River began to lose its connection to Bear Lake. A combination of warmer water temperatures, increased evaporation, and increased organic productivity triggered the precipitation of $CaCO_3$ as low-Mg calcite. The isotopic evidence shows that the salinity of the lake continued to increase, and by ~11,000 yr ago a threshold was reached when aragonite became the dominate carbonate precipitate. A return to calcite deposition between 8.5 and 7.5 cal ka, accompanied by lower O, C, and Sr isotope ratios, indicates a brief re-entry of Bear River into the lake. The sediments deposited in Bear Lake over the past 7000 years consist of ~75% $CaCO_3$, mostly as aragonite and only ~5% as low-Mg calcite.

5. Most of the time during the past 220,000 years, Bear Lake was connected to Bear River. During these times the Aleutian low-pressure atmospheric circulation system was the dominant feature of the eastern North Pacific, bringing abundant moisture to the western United States and enhancing the flow of Bear River. However, three high-carbonate, aragonitic intervals in sediments deposited over the past 220,000 years indicate times when Bear River was not connected to Bear Lake, and the salinity of the lake increased. During these three intervals the North Pacific subtropical high-pressure system dominated circulation in the eastern North Pacific, bringing dry conditions to the western United States. These three intervals correspond in time to warm, dry interglacials equivalent to marine oxygen isotope stages 1, 5, and 7.

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## ARCHIVED DATA

Archived data for this chapter can be obtained from the NOAA World Data Center for Paleoclimatology at http://www.ncdc.noaa.gov/paleo/pubs/gsa2009bearlake/.

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# Geological Society of America Special Papers

## Endogenic carbonate sedimentation in Bear Lake, Utah and Idaho, over the last two glacial-interglacial cycles

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