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Climatic and limnologic setting of Bear Lake, Utah and Idaho

Walter E. Dean
U.S. Geological Survey, MS 980 Federal Center, Denver, Colorado 80225, USA

Wayne A. Wurtsbaugh
Department of Watershed Sciences and the Ecology Center, Utah State University, Logan, Utah 84322, USA

Vincent A. Lamarra
Ecosystems Research Institute, Logan, Utah 84321, USA

ABSTRACT

Bear Lake is a large alkaline lake on a high plateau on the Utah-Idaho border. The Bear River was partly diverted into the lake in the early twentieth century so that Bear Lake could serve as a reservoir to supply water for hydropower and irrigation downstream, which continues today. The northern Rocky Mountain region is within the belt of the strongest of the westerly winds that transport moisture during the winter and spring over coastal mountain ranges and into the Great Basin and Rocky Mountains. As a result of this dominant winter precipitation pattern, most of the water entering the lake is from snowmelt, but with net evaporation. The dominant solutes in the lake water are Ca^{2+}, Mg^{2+}, and HCO_3^{2-}, derived from Paleozoic carbonate rocks in the Bear River Range west of the lake. The lake is saturated with calcite, aragonite, and dolomite at all depths, and produces vast amounts of carbonate minerals. The chemistry of the lake has changed considerably over the past 100 years as a result of the diversion of Bear River. The net effect of the diversion was to dilute the lake water, especially the Mg^{2+} concentration.

Bear Lake is oligotrophic and coprecipitation of phosphate with $CaCO_3$ helps to keep productivity low. However, algal growth is colimited by nitrogen availability. Phytoplankton densities are low, with a mean summer chlorophyll *a* concentration of 0.4 mg L^{-1}. Phytoplankton are dominated by diatoms, but they have not been studied extensively (but see Moser and Kimball, this volume). Zooplankton densities usually are low (<10 L^{-1}) and highly seasonal, dominated by calanoid copepods and cladocera. Benthic invertebrate densities are extremely low; chironomid larvae are dominant at depths <30 m, and are partially replaced with ostracodes and oligochaetes in deeper water. The ostracode species in water depths >10 m are all endemic. Bear Lake has 13 species of fish, four of which are endemic.

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INTRODUCTION

Bear Lake (42°N, 111°20′ W) is an alkaline lake that occupies the southern half of the Bear Lake Valley on the Utah-Idaho border (Fig. 1). The present elevation of the lake when full is 1805 m (5922 feet) above sea level, but this level has varied considerably over the past 100 years, mainly in response to drought conditions (Fig. 2). Flow volumes of the Green River at Green River, Utah, 125 km east of the lake are low when Bear Lake level is low (Fig. 2), indicating that fluctuations of the elevation of Bear Lake are due to regional and not local conditions. The low levels of Bear Lake during the drought years of the 1990s and 2000s approached the low levels during the dust-bowl drought of the 1930s (Fig. 2).

The natural watershed of the lake is relatively small, having a basin- to lake-area ratio of ~4.8 (Wurtsbaugh and Luecke, 1997). The Bear River is the largest river in the Great Basin, and is the principal source of surface water flowing into Great Salt Lake. The Bear River originates in the Uinta Mountains 125 km southeast of Bear Lake in bedrock composed mostly of quartzite. Tributaries of the Bear River, and the streams that flow directly into Bear Lake, originate in the northern Wasatch Range, which is composed primarily of Paleozoic carbonate rocks (Hintze, 1973; Reheis et al., this volume).

Within historic times, the Bear River bypassed Bear Lake. However, part of the river's flow was diverted into Bear Lake through a series of canals, beginning in 1909 with completion in 1918 (Birdsey, 1989), making Bear Lake a reservoir to supply water for hydropower and irrigation downstream, and this continues today. Apparently the first Bear River water entered Mud Lake (Fig. 1) through a canal at the north end of Mud Lake in May 1911. Presumably some of this water overflowed into Bear Lake, but there are no records of the timing or amount of this overflow (Mitch Poulsen, Bear Lake Regional Commission, 2004, personal commun.; Connely Baldwin, PacifiCorp, 2005, personal commun.). A control structure was subsequently added to control the flow of water from Mud Lake to Bear Lake, but a minimal volume of water was diverted into Bear Lake before 1913 and possibly even later (Connely Baldwin, PacifiCorp, 2005, personal commun.).

The diversion of Bear River into Bear Lake increased the basin-to-lake-area ratio considerably, to 29.5. Since the diversion, the mean annual surface hydrologic flux (including precipitation) to the lake is estimated to be 0.48×10^9 m^3 yr^{-1} (Lamarra et al., 1986). Outflow is estimated as 0.214×10^9 m^3 yr^{-1}, which is only ~3% of the lake volume (8.0×10^9 m^3), giving an average residence time of ~37 yr. The amount of groundwater influx may be considerable but is not known. Bright (this volume, Chapter 4) concluded that the hydrologic budget of the lake is near zero.

The main purpose of this paper is to document the physical, chemical, and biological limnologic conditions in Bear Lake based mainly on unpublished data, some of them going back as far as 1975. A second objective is to examine the climatic setting of Bear Lake, both in terms of how the Bear Lake region is related to the climate of the western United States, and in terms of local meteorological conditions.

METHODS

A vertical profile of water samples at 10 m intervals in the deepest part of Bear Lake was collected using a Kemmerer water sampler between six and 13 times yearly between 1981 and the present, allowing the documentation of water quality conditions during periods of spring and fall circulation and summer and winter stratification. The samples were analyzed for total and orthophosphorus, nitrate, nitrite, ammonia, total suspended solids, and chlorophyll *a* using standard methods (American Public Health Association, 1992) at the Ecosystems Research Institute (ERI), Logan, Utah, a state and EPA certified laboratory. Prior to 1996, samples for dissolved oxygen were collected in the field and analyzed at the ERI laboratory by the standard Winkler titration method. Field temperature, dissolved oxygen, pH, conductivity, and turbidity were measured using a Hydrolab H_2O Water Quality Multiprobe (1996–2002), or with an In Situ MP-Troll 9000 (2003-present). In the field, each sample was split immediately into a bottle with acid preservative for the ammonia, nitrate, and total phosphorus analyses, an unpreserved bottle for the total suspended solids, nitrite, and orthophosphorus analyses, and an unpreserved bottle for chlorophyll *a* analysis. The chlorophyll *a* sample was filtered in the laboratory, frozen, and subsequently extracted in 100% buffered methanol for 24 h at room temperature. The extracts were analyzed fluorometrically with a correction for phaeophytin in a fluorometer calibrated with standard chlorophyll *a* (Holm-Hansen and Riemann, 1978).

Secchi-disk transparency was measured with a 20 cm white disk from the shaded side of a boat or under the ice. Temperature profiles were made with a Yellow Springs Instruments Model 58 thermistor. Specific conductance was made with an SBE 25 Sealogger profiler.

CLIMATIC SETTING

Regional Climatic Setting

The climate of the western United States is dominated by atmospheric circulation over the North Pacific Ocean and adjacent land areas (the North Pacific High, the Aleutian Low, and the North American Low). The seasonal strengths and positions of these pressure systems not only generate the weather and climate of the western United States (e.g., Strub et al., 1987; Thompson et al., 1993), but are part of the atmospheric teleconnections that stretch across the Northern Hemisphere (e.g., Namias et al., 1988). Extreme differences in relief in the western United States create strong elevation gradients in climate. Today, the climate of the Pacific Northwest is characterized in the spring and summer by strong, persistent, northwesterly winds generated by the juxtaposition of the North Pacific High over the eastern North Pacific and North American Low over the Great Basin, which generally

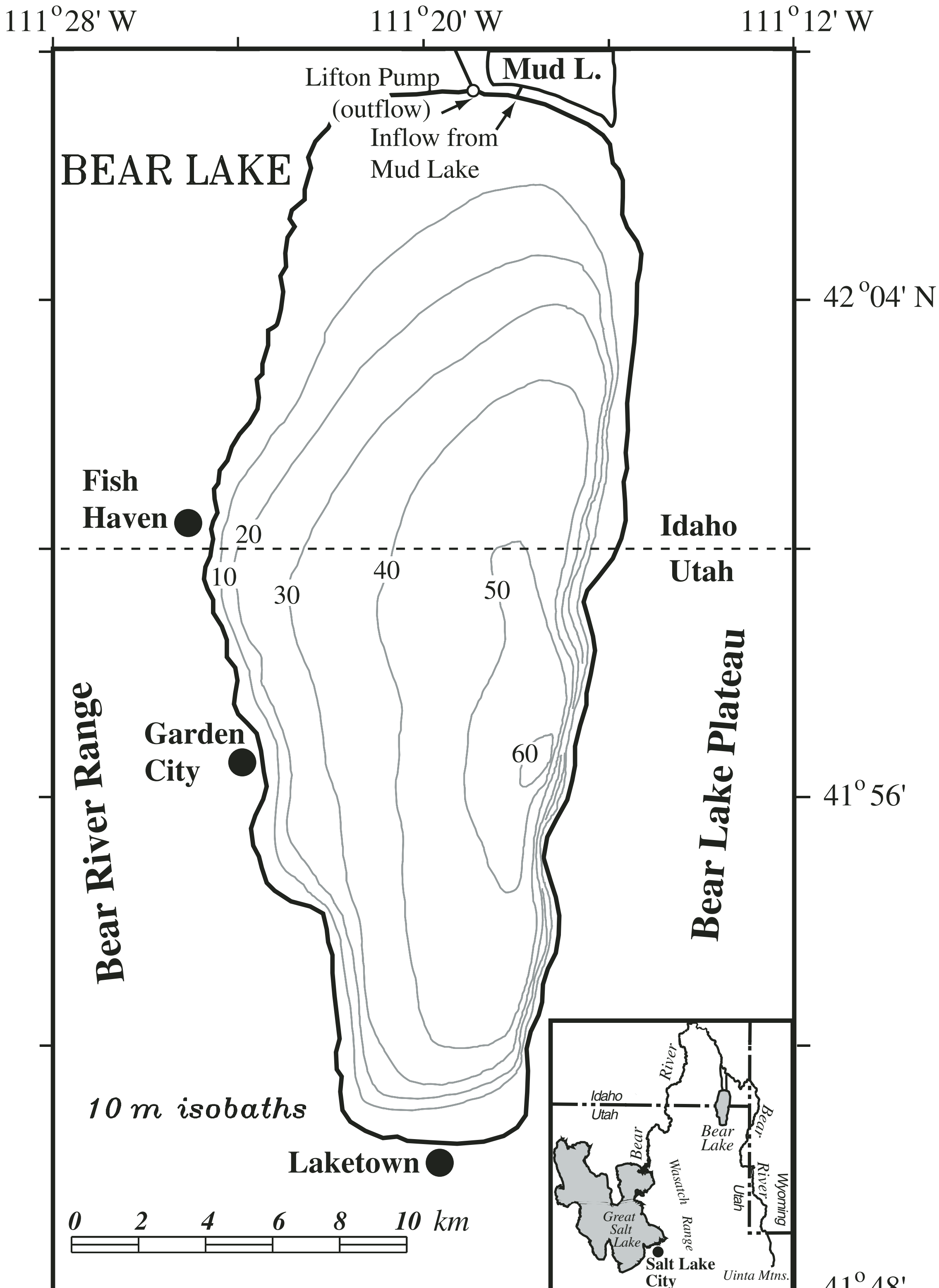

Figure 1. Bathymetric map of Bear Lake, Utah and Idaho. The inset shows the location of Bear Lake relative to Bear River and Great Salt Lake.

results in dry conditions (Thompson et al., 1993). The Aleutian Low that drives the jet stream is displaced far to the north at that time. 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. The winter dominance of the Aleutian Low produces wet and stormy weather with zonal westerly winds (Thompson et al., 1993). These winter Pacific storms lose most of their moisture as they rise over the Sierra Nevada and Cascade Ranges so that westerly air currents reaching Utah contain little moisture.

The Bear Lake region is within the belt of the strongest of these westerly winds, which transport most of the moisture into the region. As a result of marked seasonal changes in atmospheric conditions over the eastern North Pacific, the Bear Lake region experiences hot, dry summers and cold, wet winters. The limited summer moisture arrives primarily as thunderstorms that are associated with moisture-laden monsoonal air masses from the Gulf of Mexico and Gulf of California (wrcc.dri.edu/narratives/UTAH). Summer-wet/winter-dry conditions dominate the region from the southwestern United States to the southern Rocky Mountains and Great Plains, including the basins of Wyoming due to monsoonal moisture (Whitlock et al., 1993). Summer-dry/winter-wet conditions dominate higher elevations of the northern Rocky Mountains that are able to intercept winter storms that move inland from the Pacific. However, during years with increased summer monsoonal precipitation from the Pacific and Gulf of Mexico, late summer precipitation from the southwest may reach northern Utah. Data from 500 meteorological stations in the western United States for the period 1946–1994 show that most stations in Arizona, Utah, and western Colorado exceeded the average August precipitation more than 30% of the time as the result of monsoonal moisture (Mock and Brunelle-Daines, 1999). Such departures also have occurred in the past (e.g., Thompson et al., 1993; Mock and Bartlein, 1995; Mock and Brunelle-Daines, 1999). Therefore, Bear Lake has the potential to record changes in the strengths of monsoonal circulation and Aleutian Low circulation.

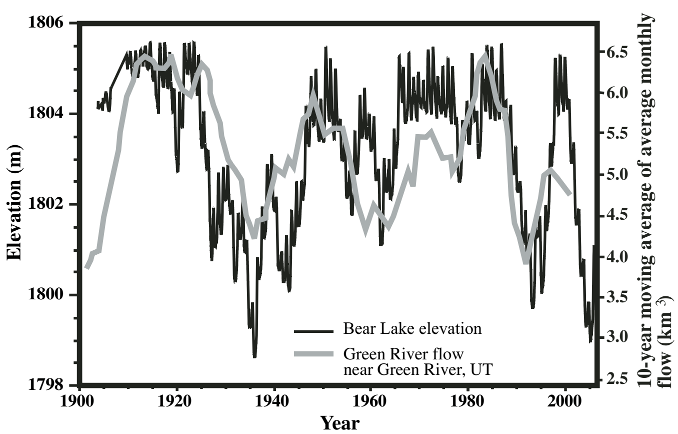

Figure 2. Elevation of Bear Lake (in meters above sea level) since 1905, and flow of the Green River near Green River, Utah. Elevation data are from PacifiCorp, Salt Lake City, Utah. Green River flow diagram is from Pieochota et al. (2004).

Local Temperature and Precipitation

The Bear Lake Valley is on a high plateau (1805 m) between the Bear River Range to the west and the Bear Lake Plateau to the east (Fig. 1; see Reheis et al., this volume). As described above, the valley has a continental climate with cold winters and warm to hot summers. Annual precipitation at Tony Grove Lake (elev 2415 m), in the Bear River Range west of the lake from 1979 to 2005 averages 124 cm, with the majority falling in the winter (90 cm) (wcc.nrcs.usda.gov/snow). Mean annual precipitation at Laketown is 30 cm (Fig. 3A), with the majority falling in the winter (wrcc.dri.edu/summary/Climsmut.html). The annual precipitation at Laketown over the last century has increased by ~9 cm (Fig. 3A) due mainly to an increase in winter (December–March)

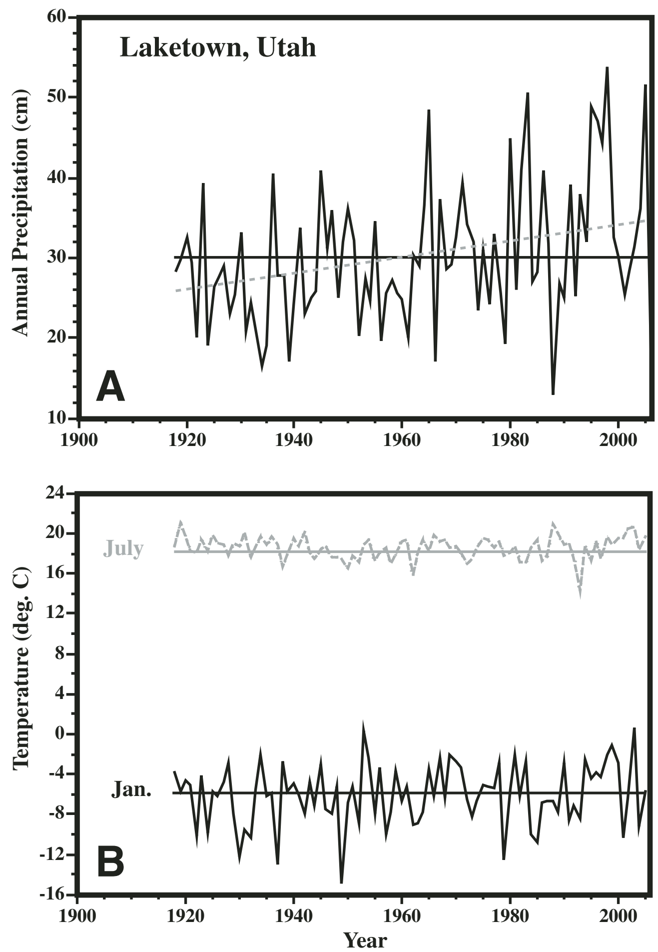

Figure 3. Annual precipitation (A) and July and January temperatures (B) since 1920 for Laketown, Utah, at the southern end of Bear Lake. Solid horizontal lines through each plot represent the mean of all data. Slanted dashed line in (A) is a linear regression through the data. Data are from wrcc.dri.edu/summary/Climsmut.html.

precipitation. Mean January temperature at Laketown, Utah, at the south end of the lake for the period 1918–2005 was –6.0 °C (Fig. 3B), and the mean July temperature for the same period was 18.2 °C (wrcc.dri.edu/summary/Climsmut.html).

Because of the dominance of winter precipitation, most of the water that enters Bear Lake, either from direct precipitation, or surface- and groundwater flow, is from snowmelt, and this is reflected in the isotopic composition of source waters to the lake (Dean et al. 2007; Bright, this volume, Chapter 4). Evaporation is poorly known, but pan estimations place it at ~100 cm yr^{-1} (Kaliser, 1972), and the average of annual pan evaporation measurements at Logan, Utah (1969–2005) and Bear River Refuge (1948–1984) are 130 cm (wrcc.dri.edu/htmfiles/westevap.final.html). The average of pan measurements from May to October for the period 1935–2002 at Lifton Pump Station, Idaho, is 107 cm (http://wrcc.dri.edu), but lake evaporation would be much lower.

PHYSICAL FEATURES

Morphometry

Bear Lake is 32 km long and has a maximum width of 12 km. At full capacity (an elevation of 1805 m), the lake has a surface area of 282 km^2, a maximum depth of 63 m, and a mean depth of 28 m (Birdsey, 1989). The volume of the lake at full capacity is 8.0×10^9 m^3. A chirp (4–24 kHz) acoustic profile (Colman, 2006) indicates that the principal structure of the basin is a half graben, with a steep, N-S oriented normal-fault margin on the east (east Bear Lake fault) and a ramp margin on the west. As a result of this structure, the lake deepens gradually from west to east, but precipitously from east to west (Fig. 1). Acoustic reflectors diverge toward the east Bear Lake fault, forming eastward-thickening sediment wedges, so that sedimentary units pinch out to the west.

Ice Cover, Thermal and Chemical Stratification

Although the Bear Lake Valley is cold in winter, the lake does not always freeze. The lake has been ice free for 25 of the past 80 years, and the frequency of freezing has decreased in the last few decades (Fig. 4). In the 11 years between 1995 and 2005, there were seven years when the lake did not freeze. The duration of ice cover is highly variable, and can range from >100 days one year to no ice cover the next (Fig. 4A). There is a general tendency for lower duration of ice cover to correspond to lower lake levels (compare Fig. 4B with Fig. 2), but there are many exceptions. For example, the lake froze over during most of the years of low lake levels in the early to mid-1990s (Fig. 4B). Ice-out usually is in April (Fig. 4A), and the timing depends on the seasonal progression of temperature and wind, generally associated with storms.

In years when Bear Lake freezes over, it behaves like a typical dimictic lake with spring and fall overturns. In years when it does not freeze over, it is monomictic with overturn in January (Wurtsbaugh and Luecke, 1997). During the annual cycle, a thermocline forms at ~10 m in May, and gradually deepens throughout summer and fall until complete mixing occurs in late December or January (Figs. 5 and 6B). During late-summer thermal stratification, the base of the epilimnion typically is between 10 and 15 m with a broad, diffuse metalimnion (Figs. 5 and 7A). The temperature of the epilimnion ranges from 2 to 3 °C in February to 18–21 °C in August and September (Figs. 5 and 6B). The temperature of the hypolimnion is relatively stable at ~5 ± 2 °C throughout the year (range 2–8 °C; Figs. 5 and 6B).

Internal waves (seiches) are common in Bear Lake, but have not been specifically studied. SCUBA-based observers report suspended sediment in the water column where the thermocline intersects the bottom, which suggests that sediments are resuspended. Because the thermocline deepens steadily throughout the summer, and seiches are common, there is ample opportunity for the resuspension of sediments into the water column. Model and empirical analyses have shown that turbulence where the thermocline intersects the bottom can entrain nutrients from the

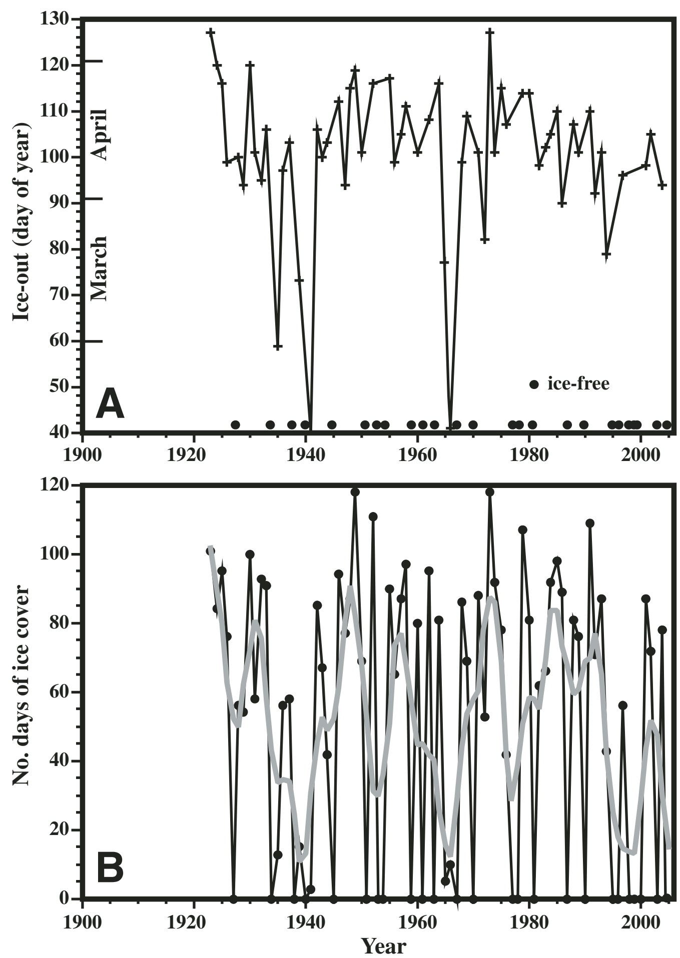

Figure 4. Day of year of ice-out (A) and number of days of ice cover (B) for Bear Lake since 1923. Heavy curved line through data in (B) is a weighted least-squares smoothing function. Data are from PacifiCorp, Salt Lake City, Utah.

sediments into the water column (Wüest and Lorke, 2003). Surface waves also entrain littoral sediments into the water column, which decreases visibility. The combination of internal and surface waves causes erosion of sediments.

Thermal stratification leads to marked chemical stratification of some parameters. Specific conductance at 25 °C in the epilimnion in late summer is ~685 µS cm^{-1}, increasing to ~700 µS cm^{-1} in the hypolimnion (Fig. 7A). The total dissolved solids (TDS in mg L^{-1}) content increases from 533 in the epilimnion to 582 in the hypolimnion (Table 1; Dean et al., 2007). Dissolved oxygen (DO) concentrations generally are high throughout the water column for most of the year, often with a maximum in the metalimnion that may be 1–2 mg L^{-1} higher than in the epilimnion due to a concentration of algal productivity. Concentrations of DO may decline below 4 mg L^{-1} in the deep hypolimnion (>50 m) by September or October (Fig. 6C) due to decomposition of produced organic matter. However, the average summer (July–September) DO at 50 m from 1981 to 2003 was 7.0 ± 1.05 mg L^{-1}, indicating that oxic conditions with high redox states predominate above the sediments.

Light Transmission

Light extinction coefficients in Bear Lake range from 0.19 to 0.28 (Neverman and Wurtsbaugh, 1992). Consequently, light intensities at the mean and maximum depths typically are 0.2% and 0.0001%, respectively, of those at the surface. Wurtsbaugh and Luecke (1997) found that Secchi depths only partly reflect changes in chlorophyll concentration (primary productivity) because of suspended carbonate particles. Secchi depth generally varies between 4 and 6 m (Fig. 6A). The average (±1 standard deviation) Secchi depth between 1975 and 2000 was 5.0 ± 1.7 m (range 1.4–12.0 m). Unusually deep Secchi depths occurred in 1996 and again in 1998 (Fig. 4), coincident with relatively high densities of the zooplankton grazer *Daphnia pulex* (Wurtsbaugh and Luecke, 1998).

There are abundant suspended $CaCO_3$ particles in the water column, which means that water transparencies are not as great as they should be for an oligotrophic lake. This is a common phenomenon in many hard-water lakes that precipitate $CaCO_3$ during the warm summer months. In these lakes, suspended and colloidal $CaCO_3$ scatters light in the blue and green wavelengths, giving these lakes a very characteristic blue color (e.g., Wetzel, 2001). Sediment-trap studies show that in Bear Lake most $CaCO_3$ precipitation as high-Mg calcite does indeed occur in the epilimnion from April through September, but below ~10 m the water column always contains particles of carbonate due to resuspension from the bottom in water <30 m (Dean et al., 2005, 2007; Dean, this volume).

LAKE CHEMISTRY

Major Ions and Carbonate Precipitation

The dominant cations in Bear Lake water today are calcium (Ca^{2+}), magnesium (Mg^{2+},) and sodium (Na^{+}) (Table 1). The dominant anion is bicarbonate (HCO_3^-), but there are also

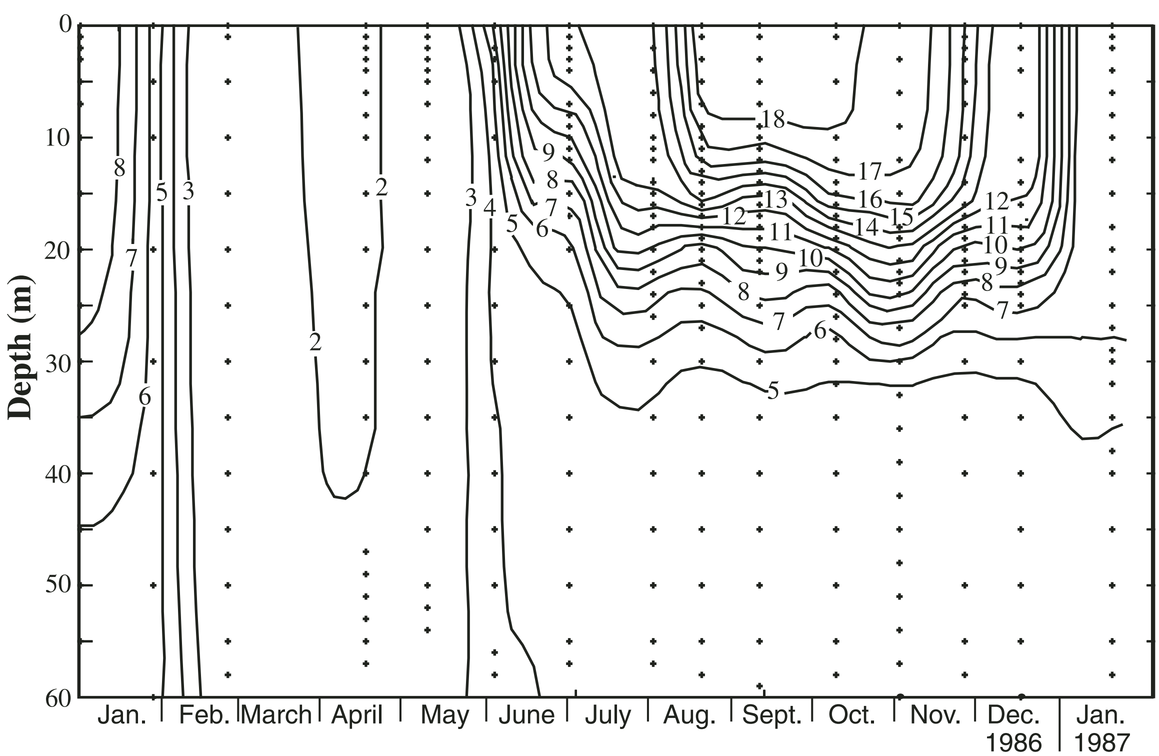

Figure 5. Temperature isopleths of Bear Lake from January 1986 through January 1987. Note that wind mixing in December and January allows the lake to cool to near 2 °C. Small crosses indicate the depths and dates of measurements.

relatively high concentrations of sulfate (SO_4^{2-}) and chloride (Cl^-). Present-day Bear Lake has two natural solute sources (west-side and east-side streams and springs) and one human-controlled solute source (Bear River). The solutes in the west-side waters (springs and streams) are Ca^{2+}-HCO_3^- dominated (Dean et al., 2007). They have high HCO_3^-:Ca^{2+} ratios (average of 4.6) due to low Ca^{2+} concentrations. They also have low concentrations of most other ions. The west-side creeks originate as springs on the east side of the Bear River Range, fed by groundwaters flowing through cavernous Paleozoic carbonate rocks in the Bear River Range (Dean et al., 2007). The east-side waters (springs and streams) have a wide variety of compositions dominated by some

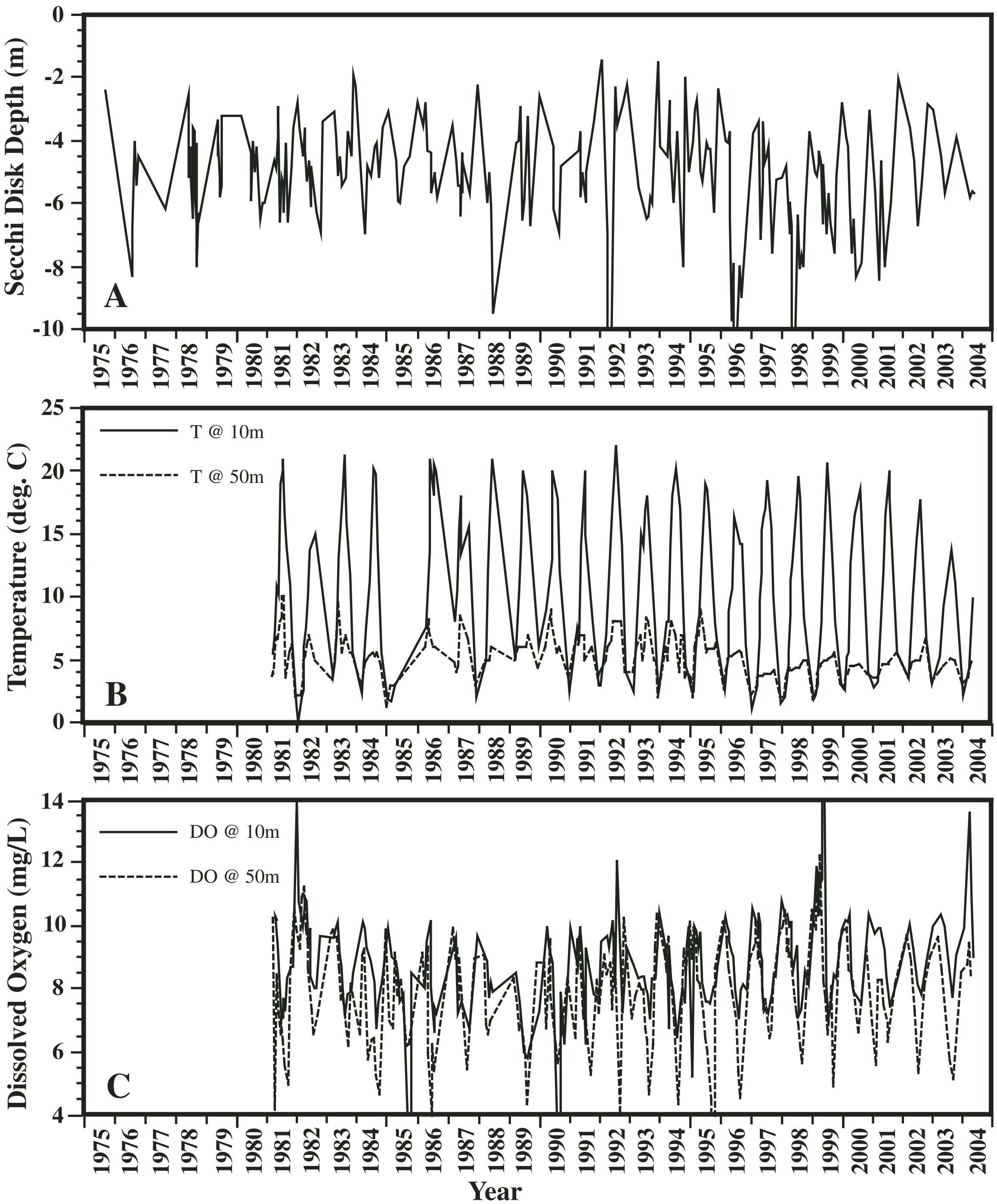

Figure 6. Secchi-disk depths (A), temperatures at 10 m and 50 m (B), and dissolved oxygen concentrations at 10 m and 50 m (C) in Bear Lake from 1975 to 2004.

combination of bicarbonate, Ca^{2+}, Mg^{2+}, Na^{+}, SO_4^{2-}, or Cl^{-}. They have low HCO_3^{-}:Ca^{2+} ratios (average of 2.9) due to high Ca^{2+} concentrations (Dean et al., 2007). The composition of Bear River water more closely resembles that of east-side waters (Figs. 8A and 8B), which may reflect base flow from the same groundwaters that discharge to the east-side springs and streams.

The oldest chemical analysis from Bear Lake is of a sample collected in 1912 (Table 1; Kemmerer et al., 1923). We use 1912 as the nominal time of Bear River diversion, and assume that the 1912 analysis is close to the composition of the lake at the time of diversion (Dean et al., 2007). The chemistry of the lake has changed considerably over the past 100 years. Now the lake water is highly enriched in Mg^{2+} relative to Bear River, and was even more so prior to Bear River diversion (Fig. 8B, Table 1). The present lake water also is more enriched in Mg^{2+} relative to the surface streams entering the lake (Fig. 8B; Dean et al., 2007). The decrease in Ca^{2+} and increase in Mg^{2+} in the lake relative to inflowing surface streams is mostly due to precipitation of $CaCO_3$. Most likely there is, and has always been, a large Mg-rich source of groundwater entering Bear Lake. This groundwater source was even more important prior to diversion as evidenced by the extremely high Mg^{2+}:Ca^{2+} in the 1912 water (Table 1). Prior to diversion, this groundwater source prevented the lake from becoming saline or even drying up (Dean et al., 2007). This groundwater source is probably from a deep aquifer because shallow groundwaters, as sampled in springs and wells around the lake, all have compositions similar to surface waters (Dean et al., 2007). A thermal inversion was observed within 2 m of the lake bottom on one occasion in summer. A warm layer was overlain by cooler water, suggesting that a denser, possibly saltier, water was flowing into the lake from a sublacustrine spring. Although we were unable to find the source of the spring, this observation supports the hypothesis that some aspects of Bear Lake's chemistry are due to the chemical composition of sublacustrine spring inflow (Dean et al., 2007).

The net effect of the diversion of Bear River into Bear Lake was to dilute the lake water. The Mg^{2+}:Ca^{2+} and TDS in the Bear Lake water sample collected in 1912 were 38 (62.5 molar) and 1280 mg L^{-1}; today they are 1.7 (3.0 molar) and 530 mg L^{-1} (Table 1). In other words, following diversion of the Bear River into Bear Lake, the Mg^{2+}:Ca^{2+} was reduced 22-fold, whereas the TDS content decreased only 2.4-fold. There was also a significant reduction in HCO_3^{-} and a large increase in Ca^{2+}. The chemistry of the lake in 1912 was dominated by Mg^{2+}-HCO_3^{-}, which is highly unusual. Among hard-water lakes in glaciated temperate regions, Ca^{2+}-HCO_3^{-} lakes predominate and Mg^{2+}-HCO_3^{-} lakes are uncommon (Wetzel, 2001). The low Ca^{2+} concentration in the lake in1912 can be explained by massive precipitation of $CaCO_3$ over

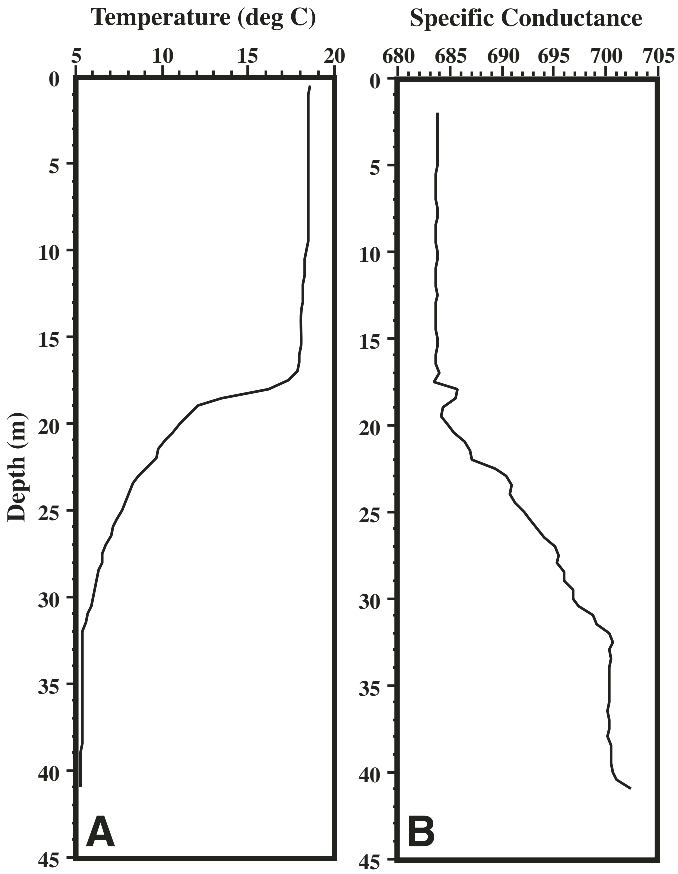

Figure 7. Temperature (A) and specific conductance (B) (in µS cm^{-1} at 25 °C) versus depth in September 2000, at a location where the water depth was 41 m.

TABLE 1. MAJOR DISOLVED IONS IN BEAR LAKE IN mg/L, FROM DEAN ET AL. (2007)

Water	Ca	Mg	Na	K	HCO_3*	SO_4	Cl	TDS†
Lake depth 4 m	31.8	51.8	40.7	4.6	293	67.7	43.4	533
Lake depth 10 m	30.9	52.8	41.0	4.7	293	68.1	43.4	534
Lake depth 15 m	32.7	52.9	40.7	4.6	287	68.5	43.6	530
Lake depth 43 m	29.7	52.2	39.1	3.8	347	66.7	43.6	582
East Shore	26.3	47.8	35.5	4.7	298	68.0	45.1	525
Lake 1912§	4.1	152	66.3	10.5	715	96.8	78.5	1123
Lake 1952#	17.0	78	23	6	313	78	57	572
Bear River at gauging station, Idaho	69.1	23.5	31.5	2.2	256	78	40	501

*HCO_3 is calculated from total alkalinity; HCO_3 for 1912 in Dean et al. (2007) is total alkalinity
†TDS—total dissolved solids (major ions)
§Kemmerer et al. (1923)
#Birdsey (1989)

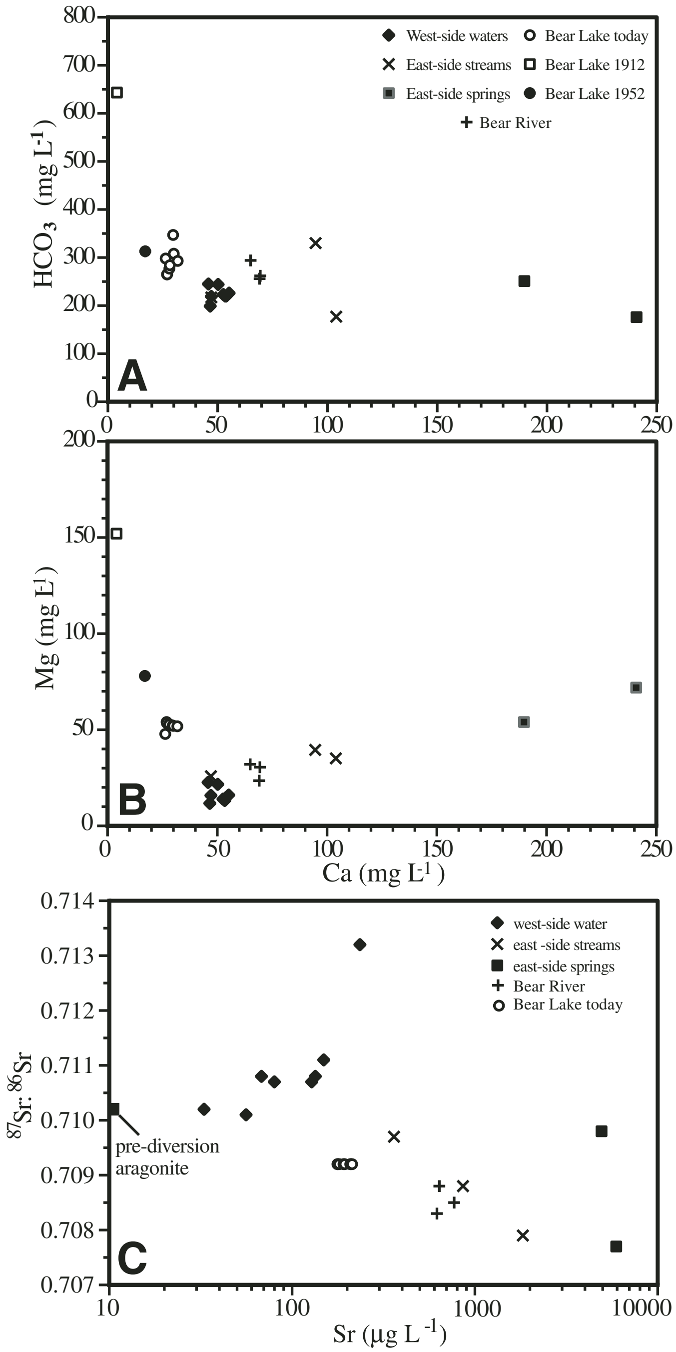

Figure 8. Crossplots of samples from Bear Lake and associated waters. (A) Total dissolved calcium (Ca) versus bicarbonate (HCO_3). (B) Total dissolved Ca versus total dissolved magnesium (Mg). (C) Dissolved strontium (Sr) concentration versus $^{87}Sr/^{86}Sr$. Note that dissolved Sr concentration is shown on a $\log_{10}$ scale. The $^{87}Sr/^{86}Sr$ value for the 1912 Bear Lake comes from samples of aragonite taken from cores at a stratigraphic horizon believed to just precede the diversion of Bear River into Bear Lake (Dean et al., 2007; noted as pre-diversion aragonite, Fig. 8C).

thousands of years. However, the high HCO_3^- is more difficult to explain because precipitation of $CaCO_3$ should remove equal molar proportions of Ca^{2+} and HCO_3^-. Dean et al. (2007) concluded that excess HCO_3^- probably was produced by evaporative concentration. Although the TDS of the 1912 lake was considerably higher than that of the present lake, the TDS of the lake was never very high, because if it had been, endemic fish and ostracode populations would have died out and saline minerals would have precipitated. In a region of warm summers, net evaporation, and little surface-water inflow (prior to Bear River diversion), lake level in the past varied considerably (e.g., Laabs and Kaufman, 2003; Reheis et al., this volume; Smoot and Rosenbaum, this volume) but the regional hydrology must have supplied the lake with a constant input of snowmelt-derived groundwater.

Qualitatively, the water chemistry is consistent with the abundance of Paleozoic carbonate rocks (predominantly dolomite) in the drainage basin, especially on the west side of the lake. Quantitatively, however, Bear Lake water does not reflect the chemistry of surface-water inputs (Dean et al., 2007). The present-day composition of Bear Lake more closely resembles the composition expected when old Bear Lake water (pre-diversion) is mixed with Bear River water introduced through the canals in the early twentieth century. In terms of the major ions (bicarbonate, Ca^{2+}, and Mg^{2+}), present Bear Lake is intermediate in composition between Bear River and 1912 lake water (Figs. 8A and B; Table 1).

Bicarbonate, Ca^{2+}, and Mg^{2+} are reactive ions in the lake due to precipitation of carbonate minerals, whereas the other major dissolved ions show conservative or near-conservative behavior (Dean et al., 2007). Bicarbonate and Ca^{2+} are lost to precipitation of aragonite (pre-diversion) or calcite (post-diversion; see discussion below). Magnesium is lost to solid solution in the carbonate minerals.

The relative importance of the three solute sources to Bear Lake hydrochemistry is shown with a plot of the $^{87}Sr/^{86}Sr$ values versus Sr concentrations (Fig. 8C). West side sources have high $^{87}Sr/^{86}Sr$ values, but low Sr concentrations, whereas the Bear River and east side sources have low $^{87}Sr/^{86}Sr$ values and high Sr concentrations. Bear Lake's $^{87}Sr/^{86}Sr$ values fall in between those of its sources (Fig. 8C), but indicate that there must be a significant input of west-side waters in order to compensate for their low Sr concentrations.

The lake is saturated at all depths with respect to calcite, aragonite, and dolomite (Dean et al., 2007). The high TDS and $Mg^{2+}:Ca^{2+}$ of the lake in 1912 should have favored the precipitation of $CaCO_3$ as aragonite (Morse and Mackenzie, 1990). Sediment-core studies show that the pre-1912 sediments deposited over the past 7000 years consist of ~80% aragonite and minor low-Mg calcite, quartz, and dolomite (Dean et al., 2006; Dean, this volume). Sediment-trap studies indicate that precipitation of $CaCO_3$ occurs in the epilimnion during the late spring and summer (April through September) as high-Mg calcite. However, sediment traps placed 2 m above the bottom in 40 m water depth show that the sediment that is accumulating on the bottom of the lake today consists predominantly of aragonite, in addition to high-Mg calcite, low-Mg calcite, quartz, and minor dolomite (Dean et al., 2007; Dean, this volume). Because so little high-Mg calcite is being incorporated in sediment on the lake floor today, the dominant $CaCO_3$ mineral (aragonite) must be aragonite that was precipitated at least 50 years ago. This depositional pattern could be explained by erosion, reworking, and "focusing" of sediment into deeper water (Dean et al., 2006; Dean, this volume).

Nutrients

Bioassay experiments have shown that algal growth in Bear Lake is usually limited by nitrogen (Wurtsbaugh, 1988) as it is in many western lakes that have not suffered from excessive agricultural or atmospheric deposition of this nutrient (Stoddard, 1994). An important dissolved inorganic form of nitrogen utilized by phytoplankton is nitrate (NO_3^-). Prior to 1997, the average epilimnetic nitrate concentration in Bear Lake was 16 µg L^{-1}, but often was below the level of detection (1 µg L^{-1}). Beginning in 1997, the nitrate concentration increased by more than an order of magnitude (Fig. 9B) when high spring runoff flushed large quantities of total inorganic nitrogen into Bear Lake from the marsh and pasture lands north of the lake.

The nutrient loading in Bear Lake is mainly from Bear River. For example, Birdsey (1989) estimated that 60%–80% of phosphorus delivered to Bear Lake is from Bear River. Phosphorus may at times limit phytoplankton growth in the lake. At the high pH levels in Bear Lake (average surface pH from 1989 to 2004 was 8.43, which changed little with depth), phosphorus can precipitate as calcium phosphate (hydroxyapatite), and, more commonly, it can coprecipitate with $CaCO_3$ and/or adsorb onto $CaCO_3$ crystals (Otsuki and Wetzel, 1972; Wetzel, 2001). Coprecipitation with $CaCO_3$ can markedly decrease the amount of phosphorus available for phytoplankton, and this may limit algal productivity in Bear Lake, helping to keep it oligotrophic (Birdsey, 1985, 1989).

Prior to 1983, the total phosphorus concentration in Bear Lake was low (usually <10 µg L^{-1}; Fig. 9C) with little buildup in the hypolimnion during summer stratification, which is characteristic of oligotrophic lakes. However, the phosphorus concentration has changed considerably over the past 25 years (Fig. 9C). The total phosphorus (TP) concentration began to increase in 1983, peaked in the early 1990s at >20 µg L^{-1}, and then began to decline. However, the TP concentration increased again in the late 1990s. The most significant form of phosphorus for plant growth is soluble inorganic phosphorus (orthophosphate, PO_4^{3-}), and it is often the limiting nutrient in lakes. Prior to 1987, the concentration of orthophosphate (OP) was low (average of ~2 µg L^{-1}), but since 1987 the average OP concentration has doubled (4 µg L^{-1}), peaking in the early 1990s along with TP (Fig. 9C).

BIOLOGICAL PROPERTIES

Phytoplankton

Bear Lake is oligotrophic, with an average (±1 standard deviation) chlorophyll *a* concentration at the surface between 1980 and 1998 of 0.53 (±0.39) µg L^{-1}. Peaks in the concentration of chlorophyll *a* >1.0 µg L^{-1} (and as high as 5.5 µg L^{-1}) occurred throughout the water column in Bear Lake in 1999, 2000, and 2004 (Fig. 9A). This suggests that algal blooms are becoming more common in Bear Lake. In April 1999 there was an unusually large algal bloom in Bear Lake, marked by unusually high chlorophyll *a* concentrations (Fig. 9A). Sedimentary evidence for this bloom was captured in a sediment trap in a water depth of 10 m (Dean et al., 2007; Dean, his volume). Chlorophyll *a* concentrations during the summer growing season generally are higher in the metalimnion than in the epilimnion (Fig. 10). The nominal base of the photic zone (1% light intensity) ranges from 15 to 25 m, so that there is sufficient light in the metalimnion for photosynthesis, which is responsible for the metalimnetic O_2 maximum.

The phytoplankton in Bear Lake have not been studied extensively. Diatoms are the most abundant taxa, and some information about their abundance is presented by Moser and Kimball (this volume). Birdsey (1989) reported that diatoms constituted ~80% of the algal abundance.

Zooplankton

Total macrozooplankton densities are low, with seasonal peaks usually of <10 individual crustaceans L^{-1} (Wurtsbaugh and Luecke, 1997). Numerically, the community is usually dominated by the cladoceran *Bosmina longirostris*, the copepod *Epischura nevadensis,* and the colonial rotifer *Conochilus unicornis*, and occasionally by *Daphnia* spp. However, *Daphnia* often dominates the community when biomass rather than numerical densities are considered (Wurtsbaugh and Luecke, 1997). There were large blooms of *Daphnia pulex* and *Daphnia galeata* in 1995 and 1996. *Daphnia* in Bear Lake may reside on the benthic sediments during the day but move into the water column at night to feed. Therefore the dominance of *Daphnia* in the zooplankton community in those years could possibly reflect a change in the magnitude of daily vertical migration (Wurtsbaugh and Luecke, 1997).

Macrozooplankton abundance is highly seasonal, with biomass minimums in winter less than 20% of those in summer. This undoubtedly decreases grazing rates on the phytoplankton during the winter, and may explain the higher chlorophyll concentrations during that period. The seasonality and compositional changes of the zooplankton will also contribute to temporal variation in the sedimentation rate. High zooplankton grazing can effectively remove phytoplankton from the epilimnion (Lampitt et al., 1990; Pilati and Wurtsbaugh, 2003), and also increase the flux of carbonate particles because these are also ingested by some zooplankton and excreted as fecal pelets (Vanderploeg, 1981; Honjo, 1996).

Benthic Invertebrates

Little research has been done on the benthic invertebrates in Bear Lake. Erman and Helm (1971) studied community composition of the invertebrates and Wurtsbaugh and Hawkins (1990) described spatial and temporal variations of the macrobenthos that are important for fish feeding. Chironomid larvae are dominant at depths <30m and are partially replaced in deeper strata by ostracodes and oligochaetes. The ostracodes in depths >10 m are all endemic species (R. Forester, 2005, personal commun.; Bright et al., 2006; Bright, this volume, Chapter 8). Biomass is lowest in winter, and more than doubles by late summer or fall. In the one year it was studied, the mean annual biomass of macroinvertebrates decreased from 0.8 g m^{-2} in the littoral zone to less than 0.15 g m^{-2} at 50 m. Overall, the mean benthic invertebrate biomass of 0.34 g m^{-2} is among the lowest recorded for any lake (Wurtsbaugh and Hawkins, 1990). This is in part because primary production in the lake is low, but also because the soft marl

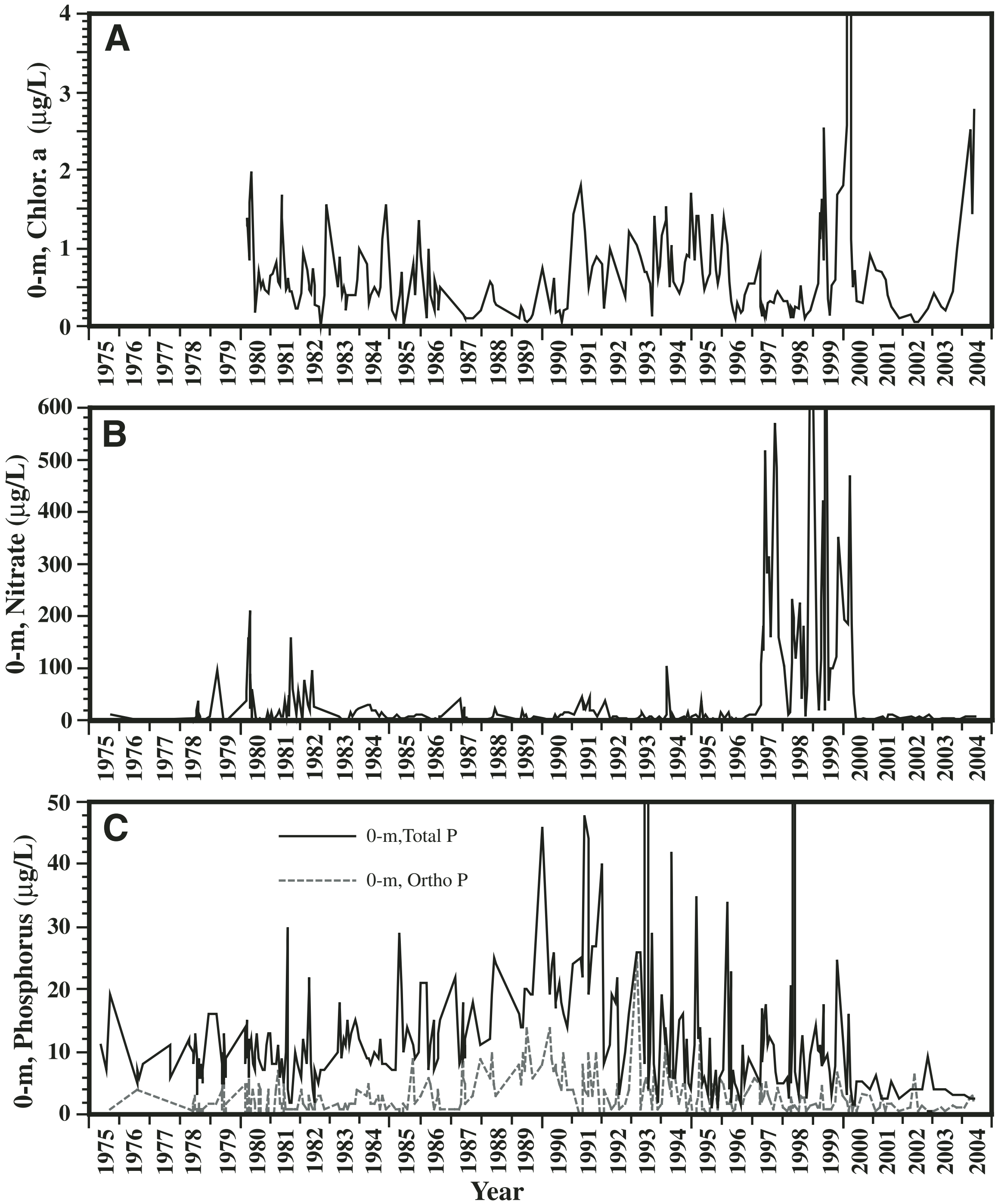

Figure 9. Concentrations of chlorophyll *a* (A), nitrate (B), and total phosphorus and orthophosphate (C) in Bear Lake at a depth of 0 m from 1975 to 2004.

sediments that dominate most of the bottom do not provide good habitats for invertebrates.

Bioturbation of the sediments by benthic invertebrates likely occurs (e.g., Martin et al., 2005), but the magnitude of disturbance may be limited because the low-organic contents of the substrate may force the invertebrates to stay close to the surface.

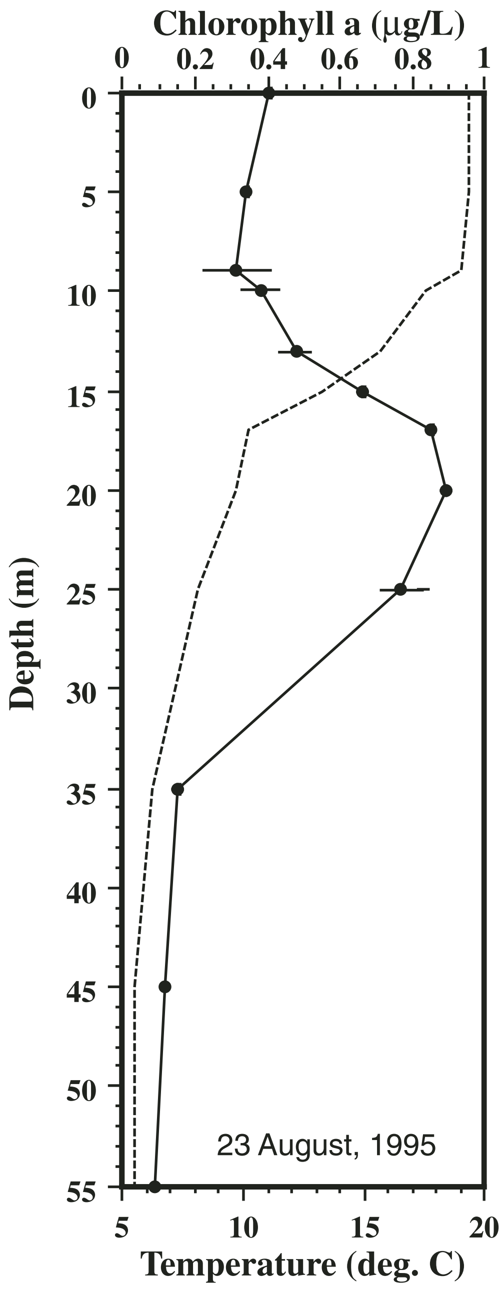

Figure 10. Depth profiles of chlorophyll *a* concentrations (solid line) and temperature (dashed line) in Bear Lake on 23 August 1995, showing the prominent deep chlorophyll layer in the metalimnion. Ranges of chlorophyll *a* concentrations are shown as bars when larger than the data points (solid circles).

Fish

Bear Lake has 13 species of fish, with four endemic species (Bonneville cisco, *Prosopium gemmifer*; Bear Lake whitefish, *Prosopium abyssicola;* Bonneville whitefish, *Prosopium spilonotus;* and Bear Lake sculpin, *Cottus extensus*). Utah suckers (*Catastromus ardens*) along with the two whitefish and sculpin are abundant and likely provide significant bioturbation of the sediments in their feeding activities at all depths in the lake. With the exception of the lake trout (*Salvelinus namaychus*), introduced species are rare. Only 1% of the fish captured in an intensive study of the lake were introduced (Wurtsbaugh and Hawkins, 1990).

The greatest biomass of fish in the lake is associated with the benthic zone, and during the summer many species are located in the littoral zone or where the metalimnion intersects the lake bottom (Wurtsbaugh and Hawkins, 1990). The metalimnion intersect provides a range of optimal temperatures for different species and often has the highest benthic invertebrate abundances. Larval sculpin living in the profundal zone of the lake undergo daily vertical migrations to the warmer metalimnion to increase digestion and growth rates (Wurtsbaugh and Neverman, 1988). The abundant sculpin also undergo an ontogenetic shift in distribution from the profundal zone when they are small, to the warmer littoral zone where growth rates are higher (Ruzycki and Wurtsbaugh, 1999).

Most fish in the lake feed primarily on benthic chironomid larvae, ostracodes, and other benthic invertebrates (Wurtsbaugh and Hawkins, 1990). Subadult cutthroat trout and the less abundant redside shiners (*Richardsonius balteatus*) rely extensively on terrestrial insects for food. Only the highly specialized zooplanktivore, the Bonneville cisco, feeds on the sparse zooplankton population (Wurtsbaugh and Hawkins, 1990). The top of the food web is dominated by the two important sport fish, adult Bonneville cutthroat trout (*Oncorhynchus clarki Utah*) and the introduced lake trout that feed initially on sculpin, but shift to preying on cisco and whitefish when they attain lengths greater than 400 mm (Ruzycki et al., 2001). The low yield of sport fish in the lake (~0.5 kg $ha^{-1}y^{-1}$; Nielson and Birdsey, 1989) is consistent with the low primary productivity and availability of invertebrate prey (Wurtsbaugh and Hawkins, 1990), but a lack of rock substrates may also limit habitat for fish spawning and rearing (Bouwes and Luecke, 1997; Ruzycki et al., 1998).

SUMMARY

Bear Lake is an old, alkaline, oligotrophic lake with endemic fish and ostracodes. The lake has no natural direct outlet, but for the last century the level of the lake has been artificially controlled by a series of canals connected at the north end to the Bear River. Most of the precipitation in the drainage basin falls in the winter and spring from Pacific storms so that most of the water entering the lake is from snowmelt. As a result of marked seasonal changes in atmospheric circulation over the eastern North Pacific, the Bear Lake region experiences hot, dry summers and cold, wet winters. This winter-wet, summer-dry moisture regime results in net evaporation, which keeps the lake saturated with carbonate minerals. The precipitation of large quantities of $CaCO_3$ coprecipitates phosphate, helping to keep the lake oligotrophic. Today the precipitated $CaCO_3$ is in the form of high-Mg calcite, but before the introduction of lower-salinity Bear River water, the lake precipitated aragonite. Suspended and colloidal particles of $CaCO_3$ scatter light in the blue and green wavelengths, giving the lake its characteristic blue color. Most of the supply of surface water to the lake is from west-side streams that contain relatively high concentrations of calcium, magnesium, and bicarbonate derived mainly from Paleozoic carbonate rocks in the Bear River Range west of the lake, and qualitatively the present water chemistry is consistent with this supply of solutes. However, strontium isotope studies show that present lake water more closely resembles the composition expected when old Bear Lake water (pre-diversion) is mixed with Bear River water introduced through the canals in the early twentieth century. Primary productivity is colimited by phosphate and nitrate. Secondary productivity by zooplankton also is low. Four of 13 fish species in the lake are endemic, and all ostracodes in water deeper than 10 m are endemic, which indicates that the lake is old. The greatest fish biomass is associated with the benthic zone where most fish feed on benthic invertebrates.

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Climatic and limnologic setting of Bear Lake, Utah and Idaho

Walter E Dean, Wayne A Wurtsbaugh and Vincent A Lamarra

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