

Isotope and major-ion chemistry of groundwater in Bear Lake Valley, Utah and Idaho, with emphasis on the Bear River Range

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ABSTRACT

Major-ion chemistry, strontium isotope ratios ($^{87}\text{Sr}/^{86}\text{Sr}$), stable isotope ratios ($\delta^{18}\text{O}$, $\delta^2\text{H}$), and tritium were analyzed for water samples from the southern Bear Lake Valley, Utah and Idaho, to characterize the types and distribution of groundwater sources and their relation to Bear Lake's pre-diversion chemistry. Four groundwater types were identified: (1) Ca-Mg- HCO_3 water with $^{87}\text{Sr}/^{86}\text{Sr}$ values of ~ 0.71050 and modern tritium concentrations was found in the mountainous carbonate terrain of the Bear River Range. Magnesium (Mg) and bicarbonate (HCO_3) concentrations at Swan Creek Spring are discharge dependent and result from differential carbonate bedrock dissolution within the Bear River Range. (2) Cl-rich groundwater with elevated barium and strontium concentrations and $^{87}\text{Sr}/^{86}\text{Sr}$ values between 0.71021 and 0.71322 was found in the southwestern part of the valley. This groundwater discharges at several small, fault-controlled springs along the margin of the lake and contains solutes derived from the Wasatch Formation. (3) SO_4 -rich groundwater with $^{87}\text{Sr}/^{86}\text{Sr}$ values of ~ 0.70865 , and lacking detectable tritium, discharges from two springs in the northeast quadrant of the study area and along the East Bear Lake fault. (4) Ca-Mg- HCO_3 - SO_4 -Cl water with $^{87}\text{Sr}/^{86}\text{Sr}$ values of ~ 0.71060 and sub-modern tritium concentrations discharges from several small springs emanating from the Wasatch Formation on the Bear Lake Plateau.

The $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values from springs and streams discharging in the Bear River Range fall along the Global Meteoric Water Line (GMWL), but are more negative at the southern end of the valley and at lower elevations. The $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values from springs discharging on the Bear Lake Plateau plot on an evaporation line slightly below the GMWL. Stable isotope data suggest that precipitation falling in Bear Lake Valley is affected by orographic effects as storms pass over the Bear River Range, and by evaporation prior to recharging the Bear Lake Plateau aquifers.

Approximately 99% of the solutes constituting Bear Lake's pre-diversion chemistry were derived from stream discharge and shallow groundwater sources located within the Bear River Range. Lake-marginal springs exposed during the recent low lake levels and springs and streams draining the Bear Lake Plateau did not contribute significantly to the pre-diversion chemistry of Bear Lake.

INTRODUCTION

Bear Lake is a large (>280 km²), deep (>60 m), turquoise-blue lake straddling the border of north-central Utah and south-eastern Idaho (Fig. 1). Bear Lake is situated in the rain shadow of the Bear River Range and had a small local watershed prior to the 1912 diversion of the Bear River (pre-diversion watershed:lake

area = 4.5:1; Lamarra et al., 1986). Only a handful of small streams drain the surrounding highlands, yet the hydrologic budget of the lake is balanced, or nearly so (Lamarra et al., 1986; Bright et al., 2006). Long sediment cores (100 and 120 m) from the lake extend back over 250,000 years (Bright et al., 2006; Kaufman et al., this volume) and seismic evidence reveals that the valley, and likely the lake, has been in existence much longer (Colman,

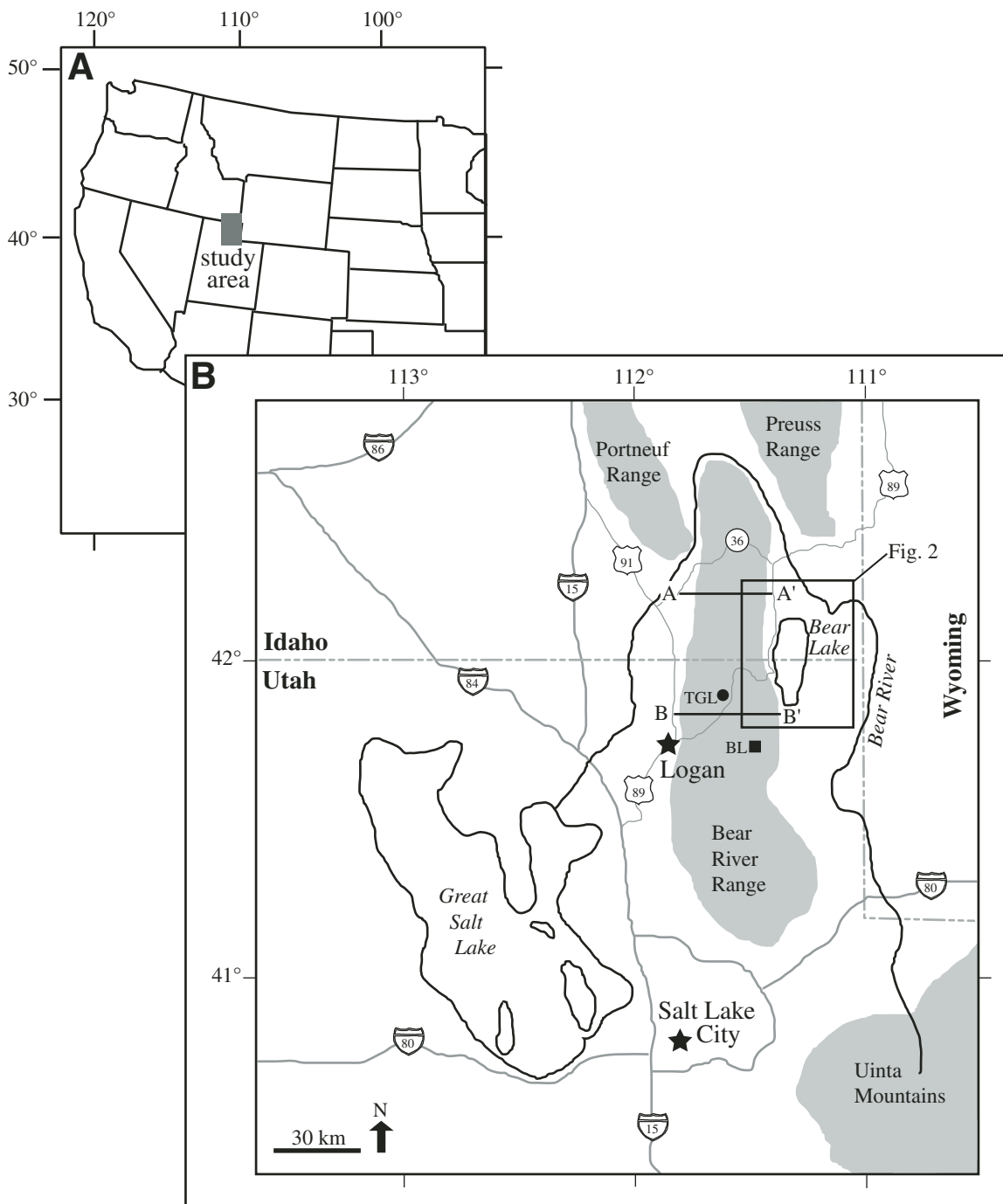


Figure 1. Bear Lake study area relative to (A) the western United States, and (B) the Great Salt Lake. TGL—Tony Grove Lake; BL—Bug Lake; cross sections A–A' and B–B' are shown in Figure 9.

2006). There are no evaporite minerals (gypsum or halite) in the long cores, indicating that the lake has survived major changes in climate without becoming saline, like the Great Salt Lake, or drying out. The lake's existence and survival are thought to be strongly dependent on groundwater, but only two small studies of groundwater in the area have been published (Kaliser, 1972; Wylie et al., 2005).

Previously published papers and other papers in this volume discuss Bear Lake's carbonate-rich sedimentary sequence and the unusual geochemistry of Bear Lake, both before and after the 1912 diversion that connected Bear River to the lake via a series of canals (e.g., Dean et al., 2006, 2007; Bright et al., 2006; Fig. 2). These previously published papers focused on Bear Lake itself, and on a subset of the available geochemical and isotopic data from the surrounding watershed as they pertained to the lake. This paper describes groundwater chemistry and distribution surrounding Bear Lake in a more spatially comprehensive manner using the entire chemical and isotopic data set, and incorporates data from studies by Kaliser (1972) and Wylie et al. (2005). The impact of groundwater on Bear Lake's pre-diversion chemistry is revisited in context of this more comprehensive assessment.

The Bear River currently bypasses the Bear Lake watershed (Figs. 1 and 2) but it has played a major role in the history of Bear Lake (e.g., Bright et al., 2006; Dean et al., 2006, 2007; Dean, this volume; Kaufman et al., this volume). Although not discussed in this paper, pertinent Bear River major-ion and strontium data are included in this paper for completeness, and for comparison to other water sources within the Bear Lake watershed.

Throughout this paper, locations with formal names are referenced by their name followed by a site number in parentheses. Many sites do not have formal names and are referenced by a general description followed by a site number in parentheses. Site numbers correspond to the numbers on Figure 2, for example, "Swan Creek Spring (22)," or "lake-marginal spring (26)."

Geologic Setting

Bear Lake Valley (Figs. 1 and 2) is situated within the Laramide overthrust belt, along the eastern margin of the Basin and Range geologic province. Traces of the Paris, Willard, Meade and Laketown thrust faults crop out sequentially in a west-to-east fashion across the valley (Fig. 2; Oriel and Platt, 1980; Dover, 1995; Willis, 1999; Liu et al., 2005). Bear Lake Valley is a north-south-trending, southeast-dipping, half-graben bounded on the west and east by the West Bear Lake Fault and East Bear Lake Fault, respectively (Fig. 2; McCalpin, 1993; Reheis et al., this volume). The majority of offset is on the East Bear Lake Fault, as illustrated by eastward-thickening lacustrine deposits within Bear Lake (Colman, 2006) and truncated spurs and prominent fault scarps along the eastern margin of the lake (Kaliser, 1972; McCalpin, 1993; Reheis et al., this volume). The East Bear Lake Fault and possibly the West Bear Lake Fault are thought to sole into the deeper thrust faults and may provide conduits for groundwater movement (Reheis et al., this volume).

The western margin of Bear Lake Valley is bounded by the Bear River Range. The Bear River Range has a maximum elevation 3042 m and in the study area is composed primarily of westward-dipping Paleozoic (Cambrian–Permian) marine limestone and dolomite, with lesser amounts of quartzite and shale (Wilson, 1979; Oriel and Platt, 1980; Dover, 1995; Spangler, 2001). Both the Eocene Wasatch Formation and Mio-Pliocene Salt Lake Formation crop out at lower elevations along the eastern flank of the Bear River Range, north of Bear Lake. The Wasatch Formation is prevalent at the southern end the valley, where it overlies the local Paleozoic sequence (Oriel and Platt, 1980; Dover, 1995; Reheis et al., this volume).

The eastern margin of Bear Lake Valley incorporates the western portion of the Bear Lake Plateau. The Bear Lake Plateau has a maximum elevation of 2349 m and is composed of early Mesozoic (Triassic and Jurassic) limestone and sandstone that is exposed in the three major drainages (Indian, North Eden, and South Eden Creeks; Fig. 2) and along the main ridgeline immediately east of Bear Lake. Several of the Paleozoic marine carbonate units in the Bear River Range crop out at the northern and southern ends of the Bear Lake Plateau (Oriel and Platt, 1980; Dover, 1995). The Bear Lake Plateau is mantled by the Eocene Wasatch Formation (Dover, 1995), although Oriel and Platt (1980) map the exposures on the Idaho portion of the Bear Lake Plateau as the Salt Lake Formation (see Coogan, 1992; Reheis et al., this volume). Coogan (1992) extensively mapped the Bear Lake Plateau and assigned four informal names to the Wasatch Formation sediments. The majority of the sediments were classified as the Diamictite (gravel and massive mudstone) and Main Body Members (sandstone and mudstone fluvial sequence). Less extensive exposures of the Quartzite Conglomerate (gravel and sandstone) and Limestone (lacustrine limestone with abundant coarse clasts) members are also present on the eastern and western margins, respectively, of the plateau. For additional discussion of the local geology, see Reheis et al. (this volume).

Precipitation, and Spring and Stream Discharge

Regional precipitation is dominated by winter storms that originate in the central and northern Pacific Ocean and move west to east across the study area. The average maximum accumulated precipitation (1979–2005) in the central Bear River Range (Tony Grove Lake, 2583 m) is ~ 125 cm yr⁻¹, and decreases southward to ~ 77 cm yr⁻¹ at Bug Lake (2423 m; rcc.nrcs.usda.gov/snotel; Fig. 1). No precipitation data exist for the Bear Lake Plateau, but three stations (15, 24, Laketown) bordering Bear Lake report mean annual precipitation values of ~ 30 cm yr⁻¹ (wrcc.sage.dri.edu/summary/climsmut and [/climsmid](http://climsmid)). The majority ($\sim 60\%$) of precipitation at Bear Lake falls as snow during the months of October through April.

Large areas of sinkholes and solution basins are common in the Bear River Range (Fig. 2; Wilson, 1979), facilitating infiltration to the aquifer systems within the mountain range. Groundwater moves across topographic drainage divides and has

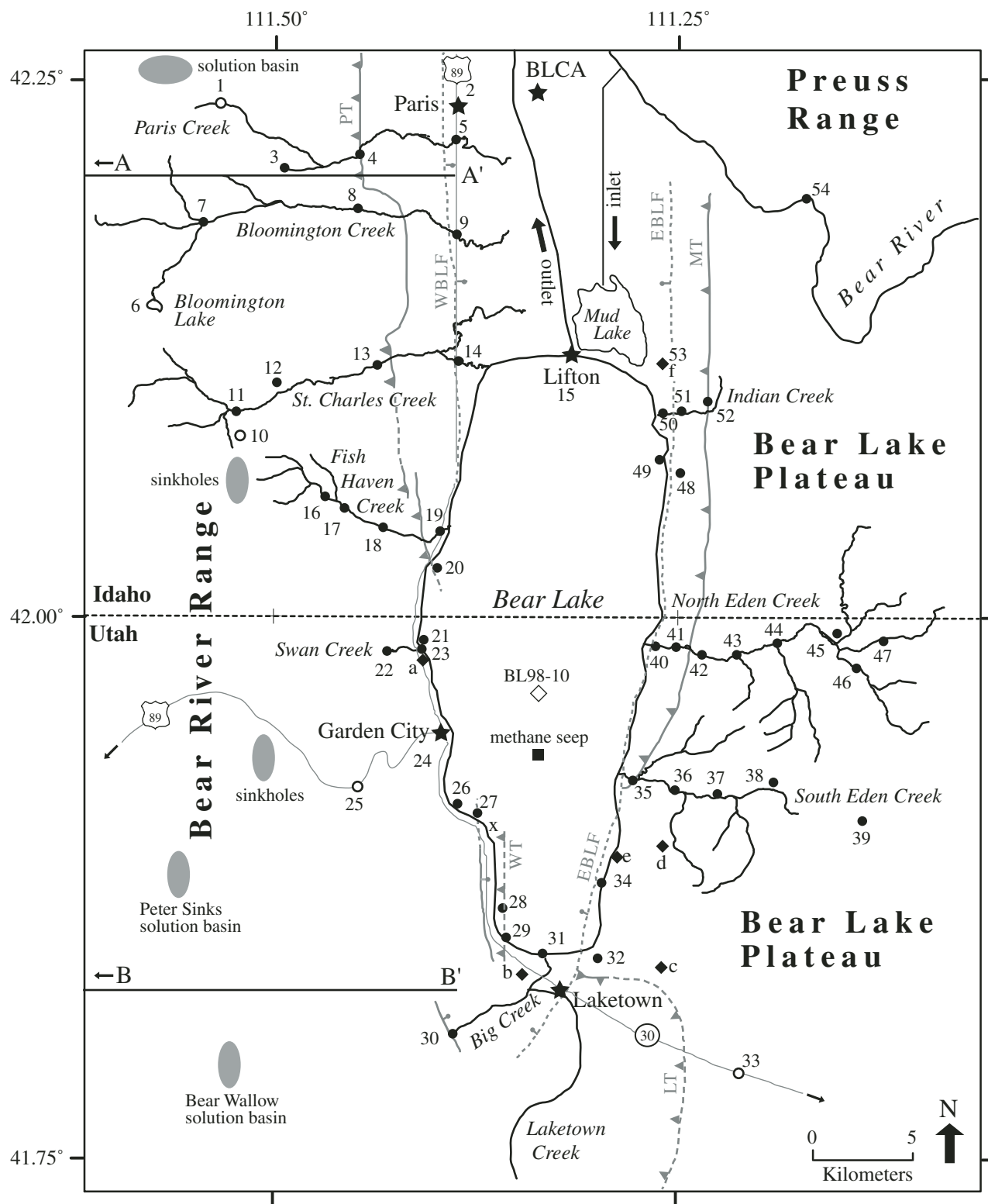


Figure 2. Locations of water samples from Bear Lake Valley. Numbers correspond to chemical and isotopic data in Tables 2–6. Solid circles are isotope and chemistry sampling sites. Open circles are snow-pit sampling sites. Solid diamonds labeled a–f represent approximate locations of sites from Kaliser (1972). Solid square in Bear Lake is location of a large methane seep (see Fig. 14 and text for discussion). Open diamond in Bear Lake is location of sediment core BL98-10. “x” is location of a dry tufa mound, along southwest shore of Bear Lake. PT—Paris Thrust; WT—Willard Thrust; MT—Meade Thrust; LT—Laketown Thrust; WBLF—West Bear Lake Fault; EBLF—East Bear Lake Fault. Approximate locations of faults shown in gray; solid where known, dashed where inferred. Teeth are on upper thrust plate, ball and pillar are on downthrown blocks. BLCA—Bear Lake County Airport; cross sections A–A’ and B–B’ are shown in Figure 9.

transmission times of less than a month in the central Bear River Range (Spangler, 2001), immediately west of the study area. Groundwater transmission times in the eastern Bear River Range area are currently unknown, but are likely similar to those reported by Spangler (2001).

The Paleozoic carbonate units that make up most of the lithologies in the Bear River Range are fractured, faulted, and karsted, which facilitates groundwater movement. For example, Logan Cave Spring and Ricks Spring, both located west of our study area, discharge from a prominent bedrock joint and along a fault, respectively (Spangler, 2001). Solution caverns in the Bear River Range may be well developed, such as Minnetonka Cave (St. Charles Canyon; not shown on Fig. 2), which is more than 600 m long with individual rooms up to 100 m long and 30 m high. Hydrologic studies by Rice and Spangler (1999) in the northern Wasatch Range, an area with a geologic and lithologic setting similar to the study area, suggest a duality in spring discharge. Their study showed that rapidly moving snowmelt pulses passed through that groundwater system within days and were superimposed on older (3–13 yr) base flow discharge. A similar situation likely occurs within the eastern Bear River Range study area. Locally, the Paleozoic Brigham (Geertzen Canyon) Quartzite is fractured and produces water (Kaliser, 1972; Wylie et al., 2005), but much less than the carbonate bedrock. Quartzites and shales within the Bear River Range are likely barriers to local groundwater movement (Wylie et al., 2005).

Spring Discharge

Springs emanating from carbonate terrain in the eastern Bear River Range are numerous and their discharges differ by an order of magnitude (Table 1). Swan Creek Spring (22) has an average discharge of $\sim 1.69 \text{ m}^3 \text{ s}^{-1}$ (Mundorff, 1971) and a maximum discharge of $\sim 9.1 \text{ m}^3 \text{ s}^{-1}$ (epa.gov/storet/dw_home.html; station 4907200). In comparison, Bloomington Spring (not shown on Fig. 2) has an average discharge of $\sim 0.03 \text{ m}^3 \text{ s}^{-1}$ (Idaho Department

of Environmental Quality, 2002), and four other large springs in the Bear River Range west of the study area have discharges on the order of $0.1\text{--}2.1 \text{ m}^3 \text{ s}^{-1}$ (Table 1; Spangler, 2001).

Springs discharging from the Wasatch Formation on the Bear Lake Plateau are small, generally 1–2 orders of magnitude less than spring discharges in the Bear River Range. All of the Bear Lake Plateau Wasatch Formation springs sampled in this study (39, 45–47) discharge from the Main Body Member. The sandstone lenses of this member may be potential aquifers but they are discontinuous and confined by surrounding mudstone. None of the local Bear Lake Plateau springs are gauged, but several springs emanating from the Wasatch Formation east and south of the study area have discharges of $<3 \times 10^{-3} \text{ m}^3 \text{ s}^{-1}$, but most estimates are in the 3×10^{-4} to $5 \times 10^{-4} \text{ m}^3 \text{ s}^{-1}$ range (waterdata.usgs.gov/nwis/gwsi). Big Spring (30), situated on a fault in the southwest corner of the study area, is the only significant spring discharging from Wasatch Formation terrain. Coogan (1992) does not describe the Wasatch Formation in the Big Spring area other than to generalize it as being fine-grained strata (and therefore probably the Main Body or possibly the Limestone Member). Big Spring is not gauged, but its discharge is larger (see stream discussion below) than that of most other springs in the Bear River Range (Table 1).

The Mesozoic Twin Creek Limestone, which is exposed primarily in the North Eden Creek drainage, is considered a confining unit within the local geologic sequence (capp.water.usgs.gov/gwa/gwa.html). Only one small spring (37) in the South Eden Creek drainage emanates from the Twin Creek Limestone within the study area. Springs emanating from the Twin Creek Limestone east of the study area are small and have discharges of $<1.5 \times 10^{-3} \text{ m}^3 \text{ s}^{-1}$ (waterdata.usgs.gov/nwis/gwsi).

The Mesozoic Nugget Sandstone crops out conspicuously along the ridgeline immediately east of Bear Lake (Dover, 1995) and is the only unit in the Bear Lake Plateau classified as an aquifer (capp.water.usgs.gov/gwa/gwa.html). No springs

TABLE 1. SPRING AND STREAM DISCHARGE IN THE BEAR RIVER RANGE AND BEAR LAKE PLATEAU

Name	Discharge ($\text{m}^3 \text{ s}^{-1}$)	Source	Maximum discharge ($\text{m}^3 \text{ s}^{-1}$)	Source
Swan Creek Spring	1.69 [†]	epa.gov/storet/dw_home.html; 4907200	9.08	epa.gov/storet/dw_home.html; 4907200
Bloomington Spring	0.03 [†]	Idaho Dept. Env. Quality (2002)	N.D.	N.D.
Bloomington Creek	0.89 [†]	waterdata.usgs.gov; station 10058600	6.43	waterdata.usgs.gov; station 10058600
Paris Creek	0.30 [†]	waterdata.usgs.gov; station 10060500	5.01	waterdata.usgs.gov; station 10060500
St. Charles Creek	1.64 [†]	waterdata.usgs.gov; station 10054600	11.13	waterdata.usgs.gov; station 10054600
North Eden Creek	0.11 [†]	epa.gov/storet/dw_home.html: station 4907120	0.34	epa.gov/storet/dw_home.html: 4907120
Big Creek	0.71 [†]	epa.gov/storet/dw_home.html: station 4907100	1.90	epa.gov/storet/dw_home.html: station 4907100
Ricks Spring*	0.06–2.10	Mundorff (1971)	4.25	Wilson (1979)
Dewitt Spring*	0.28–0.99	Spangler (2001)	0.99	Spangler (2001)
Logan Cave Spring*	0.03–0.28	Spangler (2001)	0.71	Wilson (1979)
Wood Camp Hollow Spring*	0.08–>1.13	Spangler (2001)	1.84	Wilson (1979)

Note: N.D.—no data.

*Springs located in central Bear River Range, outside of study area. “Discharge” values for these springs are ranges as provided in references.

[†]Mean discharge values for the time of record. Bloomington Spring interval not reported. Swan Creek Spring, periodically from 1979 to 2004; Bloomington Creek, 1960–1986; Paris Creek, 1942–1946; St. Charles Creek, 1961–1966; Big Creek, periodically from 1979 to 2004.

discharge directly from the Nugget Sandstone within the study area, although Falula Spring (32) may discharge at the contact between the Nugget Sandstone and valley alluvium (Kalisser, 1972). Springs discharging from the Nugget Sandstone to the east of the study area are typically small ($<1 \times 10^{-3} \text{ m}^3 \text{ s}^{-1}$; waterdata.usgs.gov/nwis/gwsi).

During 2000–2004, the Bear Lake watershed experienced its most intense drought since the 1930s. The negative moisture balance and the release of stored water caused lake level to fall nearly 6 m. As a result, several lake-marginal springs emerged, especially around the southwest margin of the lake (26–28; Fig. 2). These springs appear to discharge along several faults that define the southern margin of the lake (Fig. 2). Discharge from these springs was localized at well-defined orifices and water flowed into the lake. Wet areas several meters above lake level also developed (29, 34, 49), but these areas lacked distinct orifices and water did not reach the lake. The persistence of these beach seeps over several years suggests that they are areas of diffuse groundwater discharge. Stable isotope values from sites 29 and 34 are more positive than those from the beach springs with distinct orifices, indicating a higher degree of evaporative enrichment. These seeps are not actively precipitating tufa, but a crystalline precipitate with a salty taste was present at site 34. The chemistry of these diffuse beach seeps is reported but not discussed further because evaporation and mineral precipitation potentially alter their composition such that they may no longer be representative of the local geohydrology. A small tufa mound on an otherwise sandy expanse of beach was exposed at site 34 during the recent low lake levels as well. The tufa suggests that groundwater discharge does (or did) occur at this site. A similar tufa mound that would be flooded at full lake level is located at site X (Fig. 2) on the southwest margin of the lake. No water was associated with this mound when visited in April 2004.

Stream Discharge

Streams in the study area receive their water from two principal sources, and potentially a third source: (1) Essentially instantaneous overland flow and spring discharge from snowmelt during the spring, and to a lesser extent from infrequent summer rainstorms. Consequently, runoff to Bear Lake is strongly correlated with precipitation in the Bear River Range (Fig. 3). (2) Delayed discharge of infiltrated snowmelt (and rain) that sustains local springs and streams throughout most of the year. Stream sediment from Bloomington Creek, Swan Creek, and North Eden Creek contains grains of “popcorn tufa” (Bright et al., 2006; Dean et al., 2006), which indicate that calcite-saturated groundwater is discharged and degasses along the streambeds. (3) Extrabasinal groundwater sourced outside of the study area may discharge along faults, fractures, or bedding planes within the study area.

Stream discharge in the Bear River Range is not currently gauged, but available data for Paris Creek, Bloomington Creek, and St. Charles Creek indicate average discharges are on the order of $0.3\text{--}1.6 \text{ m}^3 \text{ s}^{-1}$, with peak discharges in excess of $6.3 \text{ m}^3 \text{ s}^{-1}$ (Table 1; waterdata.usgs.gov). Big Creek is sourced at Big Spring and flows several kilometers before reaching the lake. Laketown Creek (Fig. 2) is often ephemeral in its lower reaches and does not contribute significantly to Big Creek discharge. The discharge values at the mouth of Big Creek (31) range from $<0.03 \text{ m}^3 \text{ s}^{-1}$ to $2.1 \text{ m}^3 \text{ s}^{-1}$, and average $0.7 \text{ m}^3 \text{ s}^{-1}$ (Table 1). These values must represent minimum values for the actual discharge at Big Spring because of agricultural diversions upstream of the Big Creek gauging station.

Stream discharge on the Bear Lake Plateau is not gauged, but recent (1999, 2004) monthly estimates of instantaneous discharge for North Eden Creek were 0.15 and $0.06 \text{ m}^3 \text{ s}^{-1}$, respectively, and peak estimated discharges were 0.33 and $0.17 \text{ m}^3 \text{ s}^{-1}$, respectively

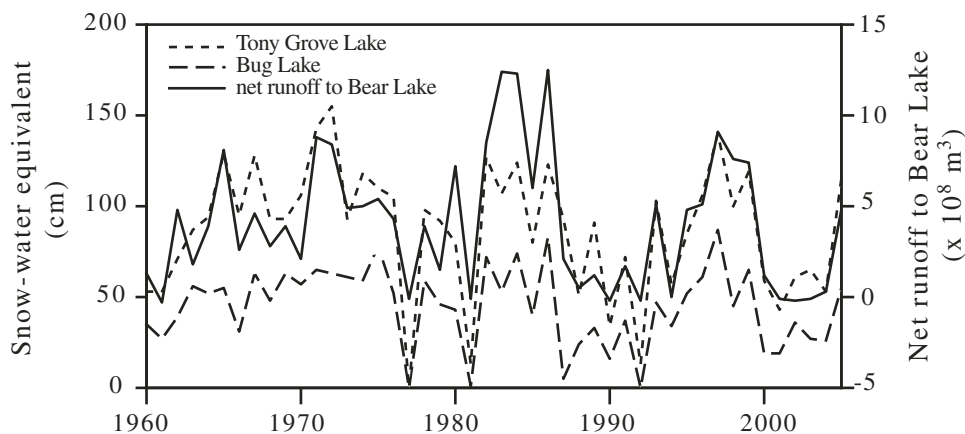


Figure 3. Snow-water-equivalent data for May at Tony Grove Lake (short-dashed line) and Bug Lake (long-dashed line) plotted against net runoff to Bear Lake (solid black line) for the years 1960–2005. Net runoff to Bear Lake = $\Delta S - (I + O + E)$; where ΔS = annual change in Bear Lake storage from elevation capacity curves ($\text{m}^3 \text{ yr}^{-1}$); I = annual inlet canal inflow ($\text{m}^3 \text{ yr}^{-1}$); O = annual outlet canal release ($\text{m}^3 \text{ yr}^{-1}$); E = evaporation ($\text{m}^3 \text{ yr}^{-1}$). Net runoff data provided by Connely Baldwin (PacifiCorp).

(Table 1; epa.gov/storet/dw_home.html; station 4907120). The 50% reduction in discharge from 1998–99 to 2003–04 illustrates the severity of the recent drought. In 2004, South Eden Creek was dry and Laketown Creek (Fig. 2) stopped flowing several kilometers upstream of the lake.

METHODS

Oxygen and Hydrogen Isotopes

Water samples for oxygen and hydrogen isotope analyses were collected between May 2003 and August 2004 ($n = 110$). Samples were collected in 25 ml heavy-gauge PVC bottles or 125 ml amber glass bottles and chilled until analysis. Snow samples (May 2003 and April 2004) were collected from the Bear River Range ($n = 4$) and Bear Lake Plateau ($n = 1$). Pits were dug into large drifts down to the ground surface and snow from the entire pit face was scraped into heavy-duty Ziplock® freezer bags and melted. Upon melting, samples were transferred to 25 ml heavy-gauge PVC bottles. Rain samples ($n = 5$) were collected during two frontal storms during 8–9 September 2003. Rain water was collected in 25 ml heavy-gauge PVC bottles as runoff from roofs at the Limber Pine trailhead (25; summit of Hwy 89 in the Bear River Range) and in the towns of Garden City (24), Utah, and Paris (2) and Lifton (15), Idaho (Fig. 2). Oxygen and hydrogen isotopes were measured on a gas-source isotope ratio mass spectrometer (Finnigan Delta S) at the University of Arizona. For oxygen, 3 ml water samples were equilibrated for 9 h with CO₂ gas at ~15 °C in an automated equilibration device coupled to the mass spectrometer. For hydrogen, samples were reacted at 750 °C with chromium metal using a Finnigan H/Device coupled to the mass spectrometer. Standardization was based on international reference materials Vienna Standard Mean Ocean Water (VSMOW) and Standard Light Arctic Precipitation (SLAP). Isotopic ratios are reported using standard del (δ) notation as per mil (‰) differences between the sample and a standard, where $\delta\text{‰} = [(R_{\text{sample}}/R_{\text{std}}) - 1] \times 10^3$ and $R =$ ratio of ¹⁸O:¹⁶O or ²H:¹H and R_{std} refers to the standards VSMOW or SLAP. Precision is 0.08‰ or better for $\delta^{18}\text{O}$ and 0.9‰ or better for $\delta^2\text{H}$ on the basis of repeated internal standards.

Major-Ion Chemistry

Twenty-six water samples for major-ion chemistry were collected from springs and streams in the Bear River Range and Bear Lake Plateau during 1999, 2000, and 2004. Water samples were collected as follows: 0.2 μm filtered, HNO₃ acidified water samples for cation analyses; 0.2 μm filtered, unacidified samples for anion analyses; and raw, unfiltered samples for total carbonate alkalinity. Samples were kept chilled on ice or refrigerated until analyzed. The 1999 samples were analyzed at the University of Minnesota. The 2000 and 2004 samples were analyzed at the U.S. Geological Survey (USGS) in Denver, Colorado; the methods of this analysis are described in Fishman and Friedman

(1985). Previously published major-ion data included in this study from Kaliser (1972) and Wylie et al. (2005), and data for the Bear River published in Dean et al. (2007), are reported in Appendix 1¹ for completeness.

Strontium Isotopes

Water samples for strontium isotope (⁸⁷Sr/⁸⁶Sr) analysis were collected in 1996, 1999, 2000, and 2004 ($n = 46$). All analyses were done on filtered, unacidified water samples that were collected in acid-washed 125 ml PVC bottles. Twenty-eight outcrop samples from 13 bedrock units were collected from the Bear River Range and Bear Lake Plateau. Rock samples were leached in 5 M acetic acid. The water samples and the rock sample leachates were centrifuged, loaded onto a cation-exchange column, and extracted with hydrochloric acid. Samples were loaded on a single tantalum filament with phosphoric acid. All isotope ratios were measured with an automated VG54 sector multi-collector, thermal-ionization mass spectrometer in dynamic mode in the USGS isotope geology laboratory, Denver, Colorado. Mass-dependent fractionation was corrected assuming a ⁸⁷Sr/⁸⁶Sr value of 0.1194. Strontium isotope ratios are reported relative to the SRM-987 standard value of 0.71025. Precision is usually ± 0.00001 . Although not discussed here, numerous ⁸⁷Sr/⁸⁶Sr values and Sr concentrations for samples from the Bear River are reported in Appendix DR2 (see footnote 1) for completeness.

Tritium

Five unfiltered, unacidified tritium samples were collected in 1 L heavy-gauge PVC bottles with no headspace. Tritium concentration was measured by liquid scintillation spectrophotometry on samples that were first distilled to remove nonvolatile solutes, and then enriched by electrolysis by a factor of about nine. Enriched samples were mixed 1:1 with Ultimagold Low Level Tritium (R) cocktail, and counted for 1500 min in a Quantulus 1220 Spectrophotometer in an underground counting laboratory at the University of Arizona. The detection limit under these conditions is 0.5 tritium units (TU).

RESULTS AND DISCUSSION

Oxygen and Hydrogen Isotopes

The stable isotopic compositions of precipitation and groundwater collected for this study and other previously published data from the surrounding area (Friedman et al., 2002) are plotted together against the Global Meteoric Water Line (GMWL; Craig,

¹GSA Data Repository item 2009047, Appendix DR1, including previously published major-ion data for springs in the Bear River Range and Bear Lake Plateau, and from the Bear River, and Appendix 2, including ⁸⁷Sr/⁸⁶Sr values and strontium concentrations for water samples taken from the Bear River, is available at <http://www.geosociety.org/pubs/ft2009.htm> or by request to editing@geosociety.org.

1961) in Figure 4A. The majority of the data plot very near the GMWL. The high isotope ratios measured in snow collected from the Bear River Range in May 2003 (Table 2) were probably the result of melting and refreezing of the snow, which preferentially removed the lighter isotopes (Cooper, 1998). Although collected in April 2004, and also subjected to melting and refreezing, the snow sample (33) from the Bear Lake Plateau has the most negative isotope values in the precipitation data (Table 2).

The isotopic composition of rain collected within the study area during a two-day frontal storm event (8–9 September 2003) shows no consistent differences between sample elevations or days of collection (Table 2). The data for rain plot to the right of the GMWL, indicating evaporation prior to reaching the ground surface (Fig. 4A). The isotopic composition of the rain is considerably heavier (more enriched) than the isotopic composition of the local groundwater discharge. Whereas rain may make up nearly 30% of the annual precipitation it does not contribute significantly to the isotopic composition of the local groundwater (e.g., Winograd et al., 1998).

The results from the spring and stream samples collected during this study fall into two distinctive groups (Fig. 4B). The Bear River Range data plot very near the GMWL ($\delta^2\text{H} = 8.2(\delta^{18}\text{O}) + 13.2$; $r^2 = 0.80$, $n = 66$) and the Bear Lake Plateau data plot slightly below the GMWL ($\delta^2\text{H} = 6.2(\delta^{18}\text{O}) - 27.6$; $r^2 = 0.89$, $n = 40$). The average $\delta^{18}\text{O}$ values overlap for the Bear River Range and Bear Lake Plateau data ($-17.4 \pm 0.5\text{‰}$ and $-17.6 \pm 0.7\text{‰}$, respectively), but water from the Bear Lake Plateau has a lower (more negative) average $\delta^2\text{H}$ value ($-136.2 \pm 4.1\text{‰}$; Table 2) than

water from the Bear River Range ($-129.7 \pm 4.2\text{‰}$; Table 2). The lower average $\delta^2\text{H}$ value from the Bear Lake Plateau spring and stream samples may result from the more isotopically negative precipitation (site 33; Table 2) that falls on the Bear Lake Plateau. Additional precipitation isotope samples from the Bear Lake Plateau are needed to verify this relationship, however.

Hydrogen isotope values of water from springs and streams in the Bear River Range become more negative from north to south (Fig. 5A) and, with the exception of Paris Creek, $\delta^2\text{H}$ values are more negative at lower elevations within individual drainage basins (Fig. 5B). In contrast to the Bear River Range, there are no apparent elevational or latitudinal trends in $\delta^{18}\text{O}$ or $\delta^2\text{H}$ values on the Bear Lake Plateau. North Eden Creek shows a slight decrease in $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values along its path, but the difference between samples is small (Table 2). Indian Creek (51) and springs 48 and 53 show decreased stable isotope values in the vicinity of the East Bear Lake Fault (Table 2, Fig. 2), suggesting that the fault may be a conduit for an isotopically depleted groundwater source.

Major-Ion Chemistry

Most reported major-ion analyses for water samples collected in Bear Lake Valley have charge balances $<5\%$, but in an effort to report a comprehensive data set, data from five stations with charge balances $>5\%$ are included in Table 3. Four of the five stations represent unique locations that were sampled only once, and for that reason they are included. The higher charge balances were most likely due to errors in the alkalinity measurements.

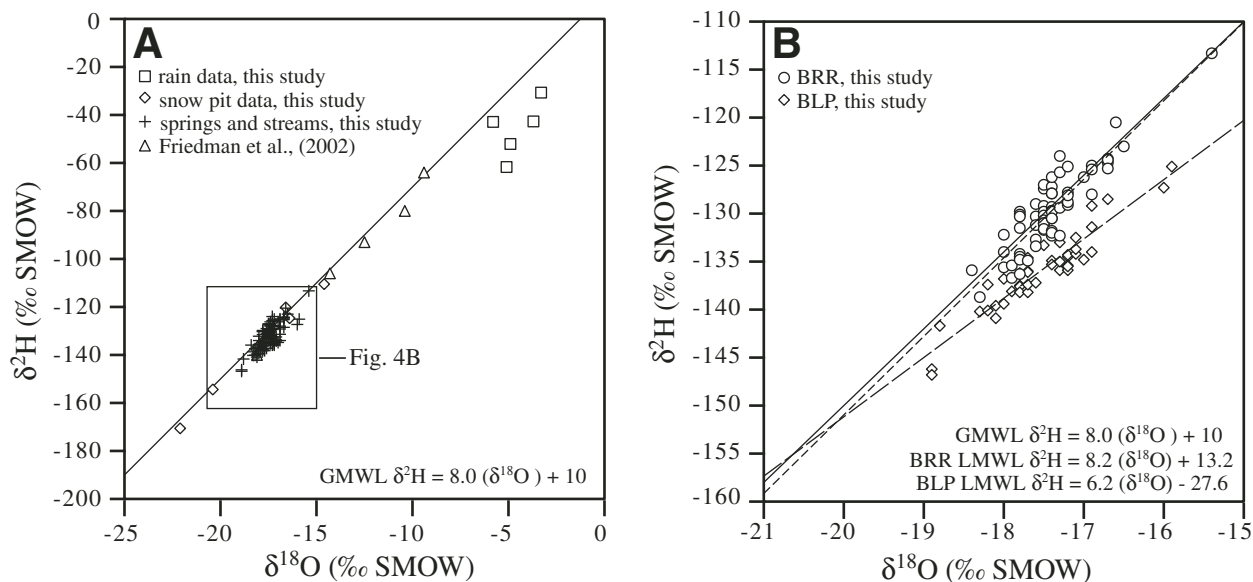


Figure 4. Stable isotope composition of water samples within and around Bear Lake Valley. (A) $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values in precipitation, springs, and streams in Bear Lake Valley. (B) Increased detail of the $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values from spring and stream samples in Bear Lake Valley (this study). Solid line represents Global Meteoric Water Line (GMWL); short-dashed line represents Local Meteoric Water Line (LMWL) for Bear River Range (BRR) samples; long-dashed line represents LMWL for Bear Lake Plateau (BLP) water samples. Data from this study are presented in Table 2. SMOW—Standard Mean Ocean Water.

TABLE 2. OXYGEN ($\delta^{18}\text{O}$) AND HYDROGEN ($\delta^2\text{H}$) ISOTOPE VALUES FROM SPRINGS AND STREAMS IN BEAR LAKE VALLEY

Site (Fig. 2)	Lat. (°N)	Long. (°W)	Elev. (m)	Site name	May 2003 $\delta^{18}\text{O}$ (‰)	May 2003 $\delta^2\text{H}$ (‰)	Sep. 2003 $\delta^{18}\text{O}$ (‰)	Sep. 2003 $\delta^2\text{H}$ (‰)	Apr. 2004 $\delta^{18}\text{O}$ (‰)	Apr. 2004 $\delta^2\text{H}$ (‰)	Aug. 2004 $\delta^{18}\text{O}$ (‰)	Aug. 2004 $\delta^2\text{H}$ (‰)	Average $\delta^{18}\text{O}$ (‰)	Average $\delta^2\text{H}$ (‰)
Precipitation														
Snow samples														
1	42.236	111.529	2315	Paris Canyon	-14.6	-110.5	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
10	42.088	111.531	2214	St. Charles Canyon	-16.4	-124.7	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
25	41.925	111.471	2288	Limber Pine Trailhead	-16.6	-120.2	N.D.	N.D.	-20.4	-154.4	N.D.	N.D.	N.D.	N.D.
33	41.805	111.248	2200	Hwy. 30	N.D.	N.D.	N.D.	N.D.	-22.1	-170.6	N.D.	N.D.	N.D.	N.D.
Rain samples														
2	42.236	111.407	1820	Paris, Idaho	N.D.	N.D.	-5.1	-61.7	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
15	42.123	111.313	1806	Lifton, Idaho	N.D.	N.D.	-3.7	-42.7	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
24	41.948	111.395	1819	Garden City, Utah	N.D.	N.D.	-4.9	-52.0	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
25	41.925	111.471	2288	Limber Pine Trailhead	N.D.	N.D.	-3.3	-30.7	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
25	41.925	111.471	2288	Limber Pine Trailhead	N.D.	N.D.	-5.2	-44.5	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
Bear River Range														
Paris Creek														
3	42.206	111.498	2001	Paris Spring	-16.9	-125.0	-17.4	-129.5	N.D.	N.D.	-17.4	-127.2	-17.2	-127.2
4	42.212	111.449	1946	Paris Creek	-16.9	-125.4	-16.9	-128.0	N.D.	N.D.	-15.4	-113.3	-16.4	-122.2
5	42.219	111.400	1813	Paris Creek	-16.7	-124.7	-17.2	-129.1	N.D.	N.D.	-17.2	-125.1	-17.0	-126.3
Bloomington Creek														
6	42.146	111.575	2499	Bloomington Lake*	N.D.	N.D.	-13.7	-110.0	N.D.	N.D.	-14.3	-110.5	-14.0	-110.3
7	42.182	111.544	2202	Bloomington Creek	-16.5	-123.0	-16.7	-125.3	N.D.	N.D.	-16.6	-120.5	-16.6	-122.9
8	42.188	111.447	1879	Bloomington Creek	-16.7	-124.3	-17.4	-129.3	N.D.	N.D.	N.D.	N.D.	-17.1	-126.8
9	42.183	111.401	1815	Bloomington Creek	-16.7	-124.4	-17.2	-128.8	-17.5	-129.2	-17.3	-124.0	-17.2	-126.6
St. Charles Creek														
11	42.096	111.530	2092	St. Charles Creek	-17.0	-126.2	-17.5	-129.8	N.D.	N.D.	-17.3	-125.7	-17.3	-127.2
12	42.105	111.495	1975	Blue Pond Spring	-17.3	-129.4	-17.5	-130.2	-17.6	-132.7	-17.4	-126.2	-17.5	-129.6
13	42.113	111.446	1940	St. Charles Creek	-17.2	-127.8	-17.5	-131.1	N.D.	N.D.	-17.5	-127.4	-17.4	-128.8
14	42.124	111.391	1817	St. Charles Creek	-17.2	-128.1	-17.4	-131.8	N.D.	N.D.	-17.5	-127.0	-17.4	-129.0
Fish Haven Creek														
16	42.054	111.467	2123	Fish Haven Creek	-17.5	-130.7	-17.3	-132.3	N.D.	N.D.	-17.4	-127.9	-17.4	-130.3
17	42.052	111.459	2086	Sadducee Spring	-17.4	-130.5	-17.6	-133.4	N.D.	N.D.	-17.6	-129.0	-17.5	-131.0
18	42.043	111.437	1960	Fish Haven Creek	-17.5	-131.0	-17.8	-134.7	N.D.	N.D.	-17.8	-129.8	-17.7	-131.8
19	42.037	111.410	1815	Fish Haven Creek	-17.5	-131.0	-17.7	-134.9	N.D.	N.D.	-17.8	-130.1	-17.7	-132.0
20	42.026	111.402	1806	spring	N.D.	N.D.	N.D.	N.D.	-17.8	-134.8	-17.8	-131.5	-17.8	-133.2
Swan Creek														
21	41.985	111.406	1807	spring	N.D.	N.D.	N.D.	N.D.	-17.8	-136.3	-18.0	-134.0	-17.9	-135.2
22	41.985	111.427	1891	Swan Creek Spring	-17.4	-129.7	-17.8	-134.2	-17.5	-131.7	-17.6	-130.3	-17.6	-131.5
23	41.985	111.410	1813	Swan Creek	-17.4	-129.8	-17.8	-134.5	N.D.	N.D.	-17.8	-130.3	-17.7	-131.5
Big Creek														
30	41.809	111.389	1824	Big Spring	-18.0	-135.6	-17.9	-136.7	-17.9	-135.4	-18.0	-132.2	-18.0	-135.0
31	41.846	111.337	1806	Big Creek*	N.D.	N.D.	N.D.	N.D.	-16.9	-131.4	-14.5	-114.2	-15.7	-122.8

(Continued)

TABLE 2. OXYGEN ($\delta^{18}\text{O}$) AND HYDROGEN ($\delta^2\text{H}$) ISOTOPE VALUES FROM SPRING AND STREAMS IN BEAR LAKE VALLEY (Continued)

Site (Fig. 2)	Lat. (°N)	Long. (°W)	Elev. (m)	Site name	May 2003 $\delta^{18}\text{O}$ (‰)	May 2003 $\delta^2\text{H}$ (‰)	Sep. 2003 $\delta^{18}\text{O}$ (‰)	Sep. 2003 $\delta^2\text{H}$ (‰)	Apr. 2004 $\delta^{18}\text{O}$ (‰)	Apr. 2004 $\delta^2\text{H}$ (‰)	Aug. 2004 $\delta^{18}\text{O}$ (‰)	Aug. 2004 $\delta^2\text{H}$ (‰)	Average $\delta^{18}\text{O}$ (‰)	Average $\delta^2\text{H}$ (‰)
Miscellaneous Bear River Range springs														
26	41.915	111.389	1806	spring	-17.5	-131.6	-17.4	-132.3	-17.4	-130.5	-17.4	-127.9	-17.4	-130.6
27	41.909	111.372	1806	spring	N.D.	N.D.	N.D.	N.D.	-17.4	-132.0	-17.6	-131.2	-17.5	-131.6
28	41.865	111.360	1801	spring	N.D.	N.D.	N.D.	N.D.	-18.3	-138.7	-18.4	-135.9	-18.4	-137.3
29	41.851	111.356	1805	beach seep [†]	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	-6.8	-74.0	N.D.	N.D.
Average (n = 66)													-17.4 ± 0.5	-129.7 ± 4.2
Bear Lake Plateau														
South Eden Creek														
36	41.920	111.254	1891	South Eden Creek	-15.9	-125.1	-16.0	-127.3	N.D.	N.D.	N.D.	N.D.	-15.9	-126.2
37	41.918	111.228	1937	spring (Twin Creek Ls)	-17.7	-136.0	-17.6	-137.2	N.D.	N.D.	N.D.	N.D.	-17.6	-136.6
38	41.921	111.189	2001	spring (Wasatch Fm.)	-17.2	-134.5	-17.3	-135.9	N.D.	N.D.	-17.3	-133.0	-17.3	-134.5
North Eden Creek														
41	41.986	111.255	1842	North Eden Creek	-17.1	-133.7	-17.2	-135.5	N.D.	N.D.	-16.7	-128.5	-17.0	-132.6
42	41.983	111.233	1873	North Eden Creek	-17.0	-134.8	-17.2	-135.4	N.D.	N.D.	N.D.	N.D.	-17.1	-135.1
43	41.984	111.212	1885	North Eden Creek	-17.1	-134.2	-17.4	-135.3	N.D.	N.D.	N.D.	N.D.	-17.3	-134.8
44	41.988	111.189	1896	North Eden Creek	-17.4	-134.9	-17.2	-135.9	N.D.	N.D.	N.D.	N.D.	-17.3	-135.4
45	41.993	111.148	2001	spring (Wasatch Fm.)	-17.8	-137.6	-17.7	-138.2	N.D.	N.D.	N.D.	N.D.	-17.8	-137.9
46	41.997	111.140	2001	spring (Wasatch Fm.)	N.D.	N.D.	N.D.	N.D.	-17.7	-136.1	-17.7	-134.6	-17.7	-135.4
47	41.991	111.119	1983	spring (Wasatch Fm.)	-17.9	-138.1	-18.0	-139.4	-17.8	-137.5	-17.8	-136.2	-17.9	-137.8
Indian Creek														
50	42.094	111.256	1830	Indian Creek	-17.7	-137.4	-17.8	-138.2	N.D.	N.D.	-18.0	-136.8	-17.8	-137.5
51	42.095	111.247	1879	Indian Creek	-18.1	-139.6	-18.2	-140.1	N.D.	N.D.	N.D.	N.D.	-18.2	-139.8
52	42.096	111.232	1964	spring	-17.2	-134.3	-17.3	-135.0	N.D.	N.D.	N.D.	N.D.	-17.3	-134.7
Miscellaneous Bear Lake Plateau springs														
32	41.842	111.302	1812	Falula Spring	-16.9	-134.0	-17.1	-132.5	N.D.	N.D.	N.D.	N.D.	-17.0	-133.3
34	41.878	111.294	1803	beach seep [†]	N.D.	N.D.	-12.4	-111.4	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
39	41.910	111.139	2054	Rabbit Spring	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	-17.5	-133.3	-17.5	-133.3
48	42.075	111.250	1835	spring	-18.3	-140.2	-18.1	-140.9	N.D.	N.D.	-18.2	-137.4	-18.2	-139.5
49	N.D.	N.D.	1804	beach seep [†]	N.D.	N.D.	-17.6	-135.5	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
53	42.115	111.264	1824	Mud Lake Hot Spring	-18.9	-146.2	-18.9	-146.8	N.D.	N.D.	-18.8	-141.7	-18.9	-144.9
Average (n = 40)													-17.6 ± 0.7	-136.2 ± 4.1

Note: Lat.—latitude; Long.—longitude; N.D.—no data.

^{*}Stable isotope values not used in Bear River Range average due to lacustrine setting and evaporative enrichment.

[†]Stable isotope values not used in averages due to diffuse character and full exposure on beach.

The remaining ions in the five questionable samples have concentrations similar to those of neighboring samples that have good charge balances (<5%), further suggesting that the alkalinity measurements are the cause of the imbalances. The data from samples with high charge balances do not appreciably affect the conclusions of this study, however.

Bear River Range

Water temperatures measured at spring orifices and streams discharging from the Bear River Range ranged from ~5.5 to ~16.0 °C. Mountain springs (3, 12, 17, 22) are consistently colder than lake-marginal and low-elevation springs (20, 21, 26–28, 30; Table 4, Fig. 6). A warm lake-marginal spring (a; Figs. 2 and 6) was also encountered by Kaliser (1972). Groundwater sourced directly from snowmelt infiltration and passing quickly through the Bear River Range along shallow flow paths should have temperatures near or slightly below the local mean annual temperature (MAT). Groundwater following deeper flow paths may be heated at depth and have temperatures above the MAT. The MAT in Minnetonka Cave (St. Charles Canyon; not shown on Fig. 2) is 4 °C, and the MAT at Bear Lake is ~6 °C. These temperatures are similar to the high elevation Bear River Range spring and stream temperatures, suggesting shallow flow paths (see section on tritium results below). Geothermal data for the Bear River Range are not available, but geothermal gradients measured in wells on the Bear Lake Plateau range from 19 to 37 °C km⁻¹ (Blackett, 2004; www.smu.edu/geothermal), and the majority of wells in Cache Valley, west of Bear Lake Valley, have geothermal gradients of ~30 °C km⁻¹ (Blackett, 2004). Assuming a geothermal gradient of ~30 °C km⁻¹ for the Bear Lake region, the lake-marginal spring discharge temperatures that are ~10 °C above MAT suggest flow depths on the order of a few hundred meters.

Total dissolved-solids concentrations (TDS) of Bear River Range springs and streams typically range from ~250 to 350 mg L⁻¹, with the higher TDS values typically occurring at lake-marginal springs (Table 3). Bear River Range water samples are dominated by calcium (Ca), magnesium (Mg), and bicarbonate (HCO₃⁻; Fig. 7). These three ions constitute >90% of the TDS (Table 5), as expected from the dissolution of carbonate bedrock that dominates the Bear River Range. Sodium (Na), chloride (Cl), and sulfate (SO₄²⁻) typically constitute <2% of the TDS in water samples north of Fish Haven Creek, but their proportions increase in Swan Creek Spring (22), Big Spring (30), and the lake-marginal springs (20, 21, 26–28) where they constitute anywhere from 3% to 15% of the TDS (Table 5). Consequently, water samples collected in the carbonate terrain north of Fish Haven Creek have high Ca:(SO₄+Cl) and Mg:(SO₄+Cl) values, but water from Swan Creek Spring (22), Big Spring (30), and the lake-marginal springs (20, 21, 26–28, a) cluster with noticeably lower Ca:(SO₄+Cl) and Mg:(SO₄+Cl) values (Fig. 8A). In addition, the lower Ca:(SO₄+Cl) values occur with higher ⁸⁷Sr/⁸⁶Sr values and lower average δ¹⁸O values (Figs. 8B and 8C).

Barium (Ba) and strontium (Sr) concentrations are also low in water samples taken north of Garden City (Table 6), indicating that these ions are minor constituents in the Bear River Range carbonate bedrock. Water samples north of Garden City (3, 9, 12–14, 19, 22) cluster with low Ba and Sr concentrations and the lake-marginal springs (20, 21, 26–28) and Big Spring (30) cluster with significantly higher values (Fig. 8D). The increased Cl, SO₄, Ba, and Sr in the low-elevation springs indicate different water-rock interactions (e.g., different source area). The concentrations of these ions increase toward the south, where the Wasatch Formation is more prevalent. Water discharging from springs emanating from the Wasatch Formation on the Bear Lake

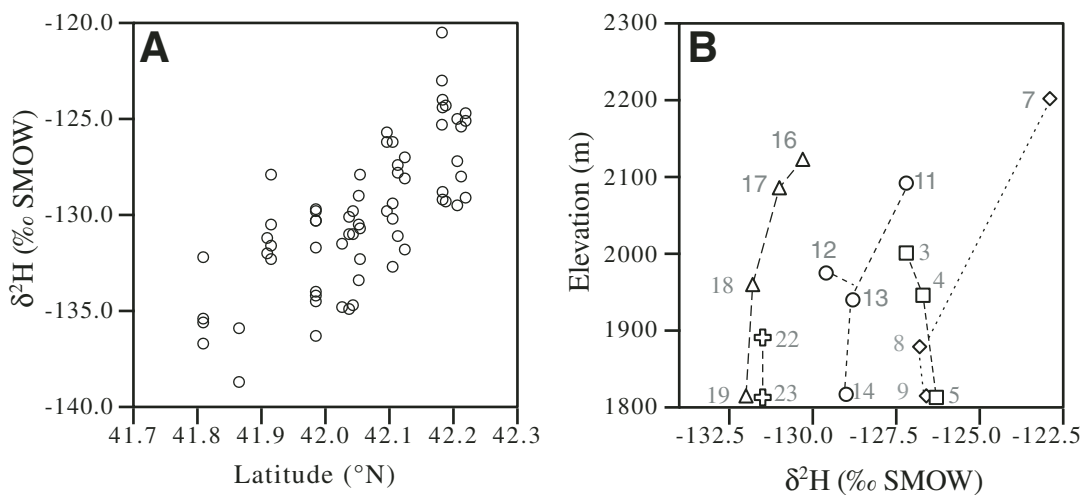


Figure 5. Hydrogen isotope ($\delta^2\text{H}$) values in spring and stream discharge for the Bear River Range. (A) Individual $\delta^2\text{H}$ values plotted by latitude. (B) Averaged $\delta^2\text{H}$ values for sample sites north of and including Swan Creek Spring plotted by elevation within individual drainages. Numbers refer to sampling locations on Figure 2 and in Table 2. Dashed lines connect samples within a common drainage. Note the progressive decrease in average $\delta^2\text{H}$ values southward along the Bear River Range. SMOW—Standard Mean Ocean Water.

TABLE 3. LOCATION, ELEVATION, AND MAJOR-ELEMENT CHEMISTRY FOR SPRINGS AND STREAMS IN BEAR LAKE VALLEY

Site (Fig. 2)	Lat. (°N)	Long. (°W)	Elev. (m)	Site name	Date (m/d/yr)	TDS (mg L ⁻¹)	Ca (mg L ⁻¹)	Mg (mg L ⁻¹)	Na (mg L ⁻¹)	K (mg L ⁻¹)	HCO ₃ ⁻ (mg L ⁻¹)	SO ₄ ⁻² (mg L ⁻¹)	Cl (mg L ⁻¹)	SiO ₂ (mg L ⁻¹)	NO ₃ ⁻ (mg L ⁻¹)	Balance* (% error)
Bear River Range																
3	42.206	111.498	2001	Paris Spring	8/1/00	292.2	46.6	11.7	2.1	0.3	228.2	2.0	1.5	6.1	N.D.	-6.0
9	42.183	111.401	1815	Bloomington Creek	9/22/99	283.9	49.7	19.3	2.5	0.4	207.4	2.9	2.1	3.9	N.D.	8.7
12	42.105	111.495	1975	Blue Pond Spring	4/5/04	256.6	41.1	17.5	1.5	0.4	192.1	2.4	1.7	2.7	1.3	4.7
13	42.113	111.446	1940	St. Charles Creek	8/1/00	329.7	45.8	22.6	1.5	0.4	256.2	2.0	1.4	4.4	N.D.	-0.6
14	42.124	111.391	1817	St. Charles Creek	9/22/99	317.3	53.2	24.9	2.1	0.3	231.8	3.0	1.8	4.6	N.D.	10.2
19	42.037	111.410	1815	Fish Haven Creek	9/22/99	272.5	35.5	19.5	2.7	0.5	207.4	5.0	2.3	4.4	N.D.	-0.8
20	42.026	111.402	1806	spring	4/5/04	368.6	54.4	21.2	7.6	1.3	268.1	9.1	7.0	4.8	1.6	0.3
21	41.985	111.406	1806	spring	4/5/04	359.7	52.4	23.2	6.9	1.4	258.1	10.0	7.8	5.5	0.5	2.1
22	41.985	111.427	1891	Swan Creek Spring	9/22/99	275.6	50.5	18.0	3.1	0.4	195.2	4.8	3.8	4.6	N.D.	9.7
22	41.985	111.427	1891	Swan Creek Spring	8/1/00	297.7	47.1	15.8	2.8	0.5	223.8	3.6	3.9	6.1	N.D.	-0.9
22	41.985	111.427	1891	Swan Creek Spring	4/5/04	296.6	53.4	13.2	4.3	0.5	214.1	3.0	8.2	2.8	1.0	1.9
26	41.915	111.389	1806	spring	8/1/04	443.1	74.5	20.4	9.8	0.8	324.6	5.0	7.6	18.7	N.D.	1.9
27	41.909	111.372	1805	spring	4/5/04	524.2	60.4	27.0	38.1	2.7	304.1	46.0	46.0	6.8	2.2	-2.0
28	41.864	111.360	1801	spring	4/5/04	369.1	50.2	21.6	19.3	2.0	234.1	19.0	23.0	4.9	1.6	2.8
29	41.851	111.356	1805	beach seep	8/23/04	1931.2	137.0	154.0	255.0	8.2	262.6	778.0	336.0	33.4	N.D.	1.3
30	41.809	111.389	1824	Big Spring	8/1/00	307.7	52.6	13.9	4.9	0.6	224.2	5.1	6.6	7.9	N.D.	0.3
30	41.809	111.389	1824	Big Spring	4/5/04	298.6	53.9	14.6	5.1	0.7	212.1	5.7	6.6	4.0	1.2	4.3
31	41.846	111.337	1806	Big Creek	9/22/99	320.4	55.2	16.0	6.4	1.0	225.7	6.7	9.1	4.3	N.D.	3.3
Bear Lake Plateau																
34	41.878	111.294	1803	beach seep	9/12/03	1694.0	120.0	54.0	340.0	20.0	190.0	660.0	310.0	2.0	N.D.	0.2
35	N.D.	N.D.	N.D.	South Eden Creek	9/22/99	659.5	94.6	39.5	32.5	2.0	329.4	105.5	55.4	9.0	N.D.	1.5
39	42.910	111.139	2054	Rabbit Spring	8/23/04	439.5	66.1	25.9	28.2	1.3	214.9	36.0	67.0	15.8	26.0	4.1
40	N.D.	N.D.	N.D.	North Eden Creek	9/22/99	462.3	66.9	32.2	32.6	2.1	207.4	69.6	51.9	7.7	N.D.	8.4
41	41.986	111.255	1842	North Eden Creek	4/5/04	404.3	47.1	25.8	27.0	2.2	217.2	40.8	44.4	8.6	2.6	0.4
46	42.977	111.140	2001	spring	4/5/04	455.9	63.8	26.1	31.9	1.1	228.1	31.0	74.0	7.1	21.0	2.0
47	42.991	111.119	1983	spring	4/5/04	402.0	31.7	19.8	17.1	1.4	248.1	21.0	33.0	4.7	4.5	0.5
48	42.075	111.250	1835	spring	8/1/00	1371.6	240.8	71.9	26.0	2.4	191.7	827.0	11.5	17.6	N.D.	-3.9
49	N.D.	N.D.	1804	beach seep	4/5/04	3172.3	555.0	198.0	80.5	1.8	396.2	1900.0	41.0	15.4	N.D.	0.3
50	42.094	111.256	1830	Indian Creek	8/23/04	601.4	104.0	35.1	21.1	1.8	176.7	240.0	22.4	15.3	N.D.	3.0
53	42.115	111.264	1824	Mud Lake Hot Spring	9/22/99	1558.9	173.2	56.5	163.7	40.8	268.4	779.4	77.3	17.6	N.D.	-3.1
53	42.115	111.264	1824	Mud Lake Hot Spring	8/1/00	1539.6	189.7	54.0	150.3	41.8	261.4	772.0	70.8	32.1	N.D.	-1.9

Note: Lat.—latitude; Long.—longitude; N.D.—no data.

*Charge balance calculated as [(cation sum - anion sum)/(cation sum + anion sum)] x 100, where units of measurement are in milliequivalents.

TABLE 4. WATER TEMPERATURES FROM SPRINGS IN THE BEAR RIVER RANGE AND BEAR LAKE PLATEAU

Site (Fig. 2)	Lat. (°N)	Long. (°W)	Elev. (m)	Site name	Sep. 2003 (°C)	Apr. 2004 (°C)	Aug. 2004 (°C)
Bear River Range							
3	42.206	111.498	2001	Paris Spring	5.2	N.D.	5.5
12	42.105	111.495	1975	Blue Pond Spring	6.3	6.2	8.5
17	42.052	111.459	2086	Sadduccee Spring	5.9	N.D.	6.0
20	42.026	111.402	1806	spring	N.D.	9.8	11.0
21	41.985	111.406	1807	spring	N.D.	15.0	16.0
22	41.985	111.427	1891	Swan Creek Spring	7.1	7.2	7.0
26	41.915	111.389	1806	spring	N.D.	9.4	10.0
27	41.909	111.372	1806	spring	N.D.	11.0	11.0
28	41.865	111.360	1801	spring	N.D.	13.0	12.0
30	41.809	111.389	1824	Big Spring	12.0	10.2	10.0
Bear Lake Plateau							
39	41.910	111.139	2054	Rabbit Spring	N.D.	N.D.	9.5
45	41.993	111.148	2001	spring	10.1	N.D.	N.D.
46	41.997	111.140	2001	spring	7.9	7.4	8.0
47	41.991	111.119	1983	spring	7.0	7.8	8.0
48	42.075	111.250	1835	spring	12.3	N.D.	13.0
53	42.115	111.264	1824	Mud Lake Hot Spring	42.1	N.D.	44.0

Note: Lat.—latitude; Long.—longitude; N.D.—no data.

Plateau (b, c, d, 39, 46, 47) is characterized by high Cl, SO₄, Ba, and Sr concentrations (Tables 3 and 6, Fig. 7, Appendix 1 [see footnote 1]). Consequently, within the limits of this study, the Wasatch Formation at the south end of Bear Lake Valley is the best candidate for the source of the elevated SO₄, Cl, Ba, and Sr concentrations found in Big Spring and the lake-marginal springs located along the western margin of Bear Lake.

Bear Lake Plateau

Spring discharge on the Bear Lake Plateau is typically cold, ~8–13 °C, except for Mud Lake Hot Spring where water temperatures are ~42 °C (Table 4). These temperatures are higher

than the local MAT by only a few degrees (except Mud Lake Hot Spring) and suggest that flow paths are not especially deep. Total dissolved-solids concentrations on the Bear Lake Plateau typically range between 400 and 600 mg L⁻¹, except for Mud Lake Hot Spring (53) and spring 48 where TDS concentrations approach 1600 mg L⁻¹ and 3200 mg L⁻¹, respectively (Table 3).

Water samples from springs emanating from the Wasatch Formation (b, c, d, 39, 46, 47) are Ca-Mg-HCO₃ waters but have moderate Na, Cl, and SO₄ concentrations (Tables 3 and 5; Appendix 1: Fig. 7 [see footnote 1]). Strontium concentrations are slightly higher than barium concentrations in three Wasatch Formation springs (39, 46, 47; Table 6). Springs that discharge near the East Bear Lake Fault (48, 53) contain very high concentrations of strontium and very little barium.

One of the more unusual springs on the Bear Lake Plateau is Mud Lake Hot Spring (f, 53), with a high Li concentration (Table 6). The high Li concentration probably indicates a hydrothermal source (White, 1957). Applying Na/Li, Li (Fouillac and Michard, 1981), and Na-K-Ca (Fournier and Truesdell, 1973) chemical geothermometry calculations to the most complete hot spring data (August 2000) suggests relatively similar water-rock interaction temperatures of 110°, 112°, and 107 °C, respectively. On the basis of these temperatures a fourth, chalcedony-based, silica geothermometry calculation was performed (Fournier, 1981). The silica thermometry result indicates a water-rock reaction temperature of only 52 °C. The lower silica result may suggest dilution or mixing of the thermal waters with another cooler water prior to reaching the surface (Fournier, 1981). Mixing with dilute surface runoff should have a negligible effect on the Na/Li, Li, and Na-K-Ca geothermometers (e.g., Fournier, 1981), but the proximity of Mud Lake Hot Spring to the East Bear Lake Fault makes the mixing of thermal water with a groundwater chemistry similar to site 48, or possibly site 50, a distinct possibility.

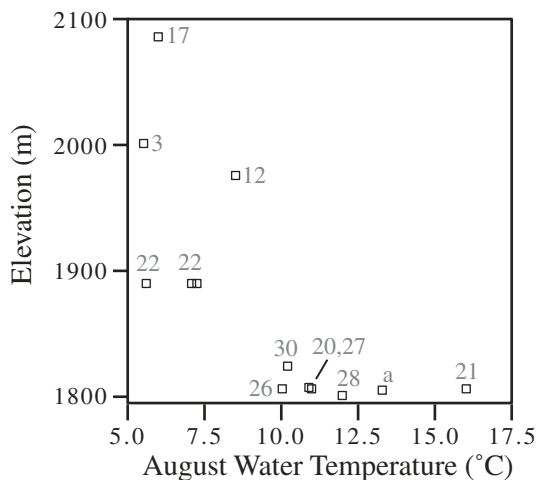


Figure 6. Elevation versus August 2004 water temperatures for spring sites along the eastern Bear River Range. Numbers and letter "a" refer to sample sites on Figure 2 and in Table 3.

Calculating what the unaltered thermal water composition may be by assuming that the Mud Lake Hot Spring chemistry is a mixture of 95% initial thermal water chemistry and 5% water chemistry from site 48 or 50 produces Na/Li, Li, and Na-K-Ca geothermometry values of $\sim 110^\circ$, 114° , and 110°C , respectively. These values are similar to the original values, and still internally consistent. Subtracting progressively larger proportions of site 48 water chemistry from the Mud Lake Hot Spring chemistry produces progressively higher Na/Li, Li, and Na-K-Ca water-rock interaction temperatures, but the results lose their internal

consistency and the silica values always remain lower than the other geothermometry values for all correction calculations. The loss of consistency among the Na/Li, Li, and Na-K-Ca results at higher mixture ratios suggests that the Mud Lake Hot Spring water is probably not composed of a high percentage of another water chemistry, East Bear Lake Fault-related or otherwise. The consistently lower silica-derived temperature is best explained by a loss of silica as the thermal waters rise toward the surface. Fournier (1981) states that solutions below $\sim 100^\circ\text{C}$ can remain supersaturated with respect to silica for an unspecified number

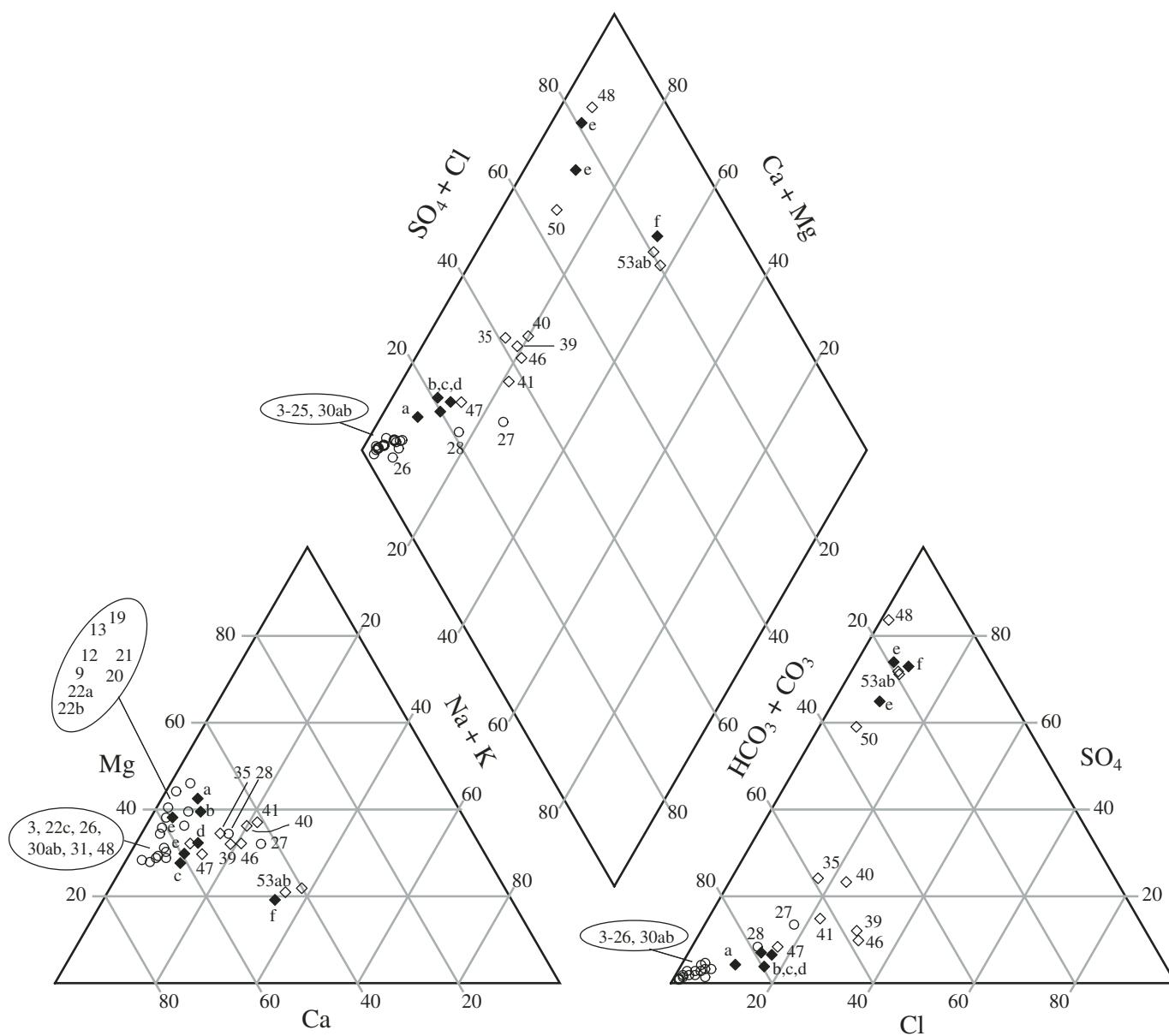


Figure 7. Piper trilinear plot of Bear Lake Valley water chemistry samples. Numbers and letters refer to sampling sites on Figure 2 and Table 3. Circles represent Bear River Range samples; open diamonds represent Bear Lake Plateau samples; solid diamonds with letters represent data from Kaliser (1972); a—lake-marginal spring; b–d—Wasatch Formation springs; e—Ca-SO₄-rich well water along East Bear Lake Fault; f—Mud Lake Hot Spring.

of years. The lack of detectable tritium in Mud Lake Hot Spring water (see below) suggests that the spring discharge is at least several decades old, and perhaps much older, and silica may have precipitated out of solution prior to reaching the surface.

Another unique spring on the Bear Lake Plateau is site 48, with high Ca and SO₄ concentrations (Table 3). This spring is situated near the East Bear Lake Fault. A well situated south of South Eden Creek and near the East Bear Lake Fault (site e in Fig. 2) was sampled by Kaliser (1972) and has a similar major-ion chemistry to spring 48, suggesting that Ca-SO₄-rich water extends southward along the eastern margin of the lake, possibly in association with the East Bear Lake Fault.

Of the three streams that discharge from the Bear Lake Plateau, North and South Eden Creeks (35, 40, 41) have relatively similar chemistry. The solutes in these streams are predominantly Ca-Mg-HCO₃-SO₄ with moderate Cl concentrations (Tables 3 and 5, Fig. 7). Strontium concentrations in both streams are higher in the vicinity of the East Bear Lake Fault (35, 40; Table 6, Fig. 2) than they are at the headwater springs east of the East Bear Lake Fault (39, 46, 47). The chemistry of Indian Creek (50) is distinct from that of North and South Eden Creeks by having high SO₄

and low Cl concentrations, and by having Ba and Sr concentrations more similar to Mud Lake Hot Spring (53) and the small sulfate-rich spring at site 48 than to either North or South Eden Creeks (Table 6, Fig. 7). These data indicate that an SO₄- and Sr-rich water discharges along the East Bear Lake Fault.

Strontium Isotopes

Strontium isotope ratios (⁸⁷Sr/⁸⁶Sr) are useful for determining water-rock interactions and serve as a groundwater tracer. Strontium isotopes do not fractionate between the solid and aqueous phase during weathering. Consequently, water-rock interactions result in a water with the same ⁸⁷Sr/⁸⁶Sr value as the rock (Bullen and Kendall, 1998).

Bear River Range

With the exception of one sample, the Paleozoic carbonate bedrock units of the Bear River Range have ⁸⁷Sr/⁸⁶Sr values ranging from 0.70811 to 0.71038, and average ~0.70928 (*n* = 12; Table 7). Most Bear River Range bedrock samples have ⁸⁷Sr/⁸⁶Sr values that are consistent with established Paleozoic seawater

TABLE 5. IONIC COMPOSITION AND TRITIUM CONCENTRATION FOR SPRINGS AND STREAMS IN BEAR LAKE VALLEY

Site (Fig. 2)	Site name	(Ca+Mg+HCO ₃)* (%)	Na* (%)	Cl* (%)	SO ₄ * (%)	Tritium† (TU)
Bear River Range						
3	Paris Spring	97.1	1.6	0.7	0.4	N.D.
9	Bloomington Creek	96.3	2.0	1.1	0.5	N.D.
12	Blue Pond Spring	97.0	1.4	1.0	0.4	N.D.
13	St. Charles Creek	97.8	1.1	0.6	0.3	N.D.
14	St. Charles Creek	97.2	1.4	0.8	0.5	N.D.
19	Fish Haven Creek	95.5	2.3	1.1	0.9	N.D.
20	spring	91.1	4.5	2.8	1.2	N.D.
21	spring	90.8	4.2	3.1	1.4	N.D.
22	Swan Creek Spring (1999)	94.5	2.4	2.0	0.9	N.D.
22	Swan Creek Spring (2000)	95.2	2.1	1.9	0.7	N.D.
22	Swan Creek Spring (2004)	92.1	3.3	3.9	0.5	11.5 ± 0.43
26	spring	91.9	4.9	2.4	0.6	N.D.
27	spring	68.4	14.9	11.7	4.3	N.D.
28	spring	77.5	10.9	8.4	2.6	N.D.
29	beach seep	32.7	25.8	22.1	18.9	N.D.
30	Big Spring (2000)	92.2	3.5	3.2	0.8	N.D.
30	Big Spring (2004)	91.7	3.7	3.2	1.0	5.0 ± 0.32
31	Big Creek–Laketown Creek	90.0	4.4	4.1	1.1	N.D.
Bear Lake Plateau						
34	beach seep	21.2	37.7	22.3	17.5	N.D.
35	South Eden Creek	69.5	10.4	11.5	8.1	N.D.
39	Rabbit Spring	63.9	12.6	19.4	3.8	N.D.
40	North Eden Creek (1999)	63.6	14.1	14.5	7.2	N.D.
41	North Eden Creek (2004)	66.7	13.4	14.4	4.8	N.D.
46	spring (Wasatch Fm.)	62.6	13.6	20.4	3.1	N.D.
47	spring (Wasatch Fm.)	76.9	8.9	11.1	2.6	2.6 ± 0.30
48	spring	54.5	5.1	1.4	38.7	<0.6
49	beach seep	53.8	6.6	2.2	37.3	N.D.
50	Indian Creek	62.8	8.3	5.7	22.7	N.D.
53	Mud Lake Hot Spring (1999)	37.4	24.1	7.4	27.5	N.D.
53	Mud Lake Hot Spring (2004)	38.9	22.6	6.9	27.8	<0.6

Note: TU—tritium units; N.D.—no data.

%X calculated from Table 4, where units of measurement are in millimoles.

†Tritium samples collected in September 2003.

$^{87}\text{Sr}/^{86}\text{Sr}$ reconstructions, although five samples yielded $^{87}\text{Sr}/^{86}\text{Sr}$ values that were slightly higher than the highest Paleozoic seawater values (Burke et al., 1982; Denison et al., 1998; Veizer et al., 1999). These higher $^{87}\text{Sr}/^{86}\text{Sr}$ values are likely due to post-depositional alteration (Burke et al., 1982; Clauer et al., 1989; Denison et al., 1994). In comparison, the $^{87}\text{Sr}/^{86}\text{Sr}$ values from the Bear River Range shale and quartzite bedrock samples were more radiogenic, averaging 0.71543 ($n = 9$; Table 7), which is reflective of their continental sources (Palmer and Edmond, 1992).

Water samples from the Bear River Range typically have low Sr concentrations (Fig. 8D) and $^{87}\text{Sr}/^{86}\text{Sr} \geq 0.71000$ (range from 0.71005 to 0.71322; Table 6, Fig. 8E). Although carbonate dissolution is nearly the sole source of solutes for the Bear River Range springs and streams, spring and stream $^{87}\text{Sr}/^{86}\text{Sr}$ values are higher than for the local carbonate bedrock. This suggests hydrologic interaction with the shale and quartzite units within the Bear River Range. Leaching of more radiogenic strontium from these units as they force groundwater to the surface leads to spring discharge with $^{87}\text{Sr}/^{86}\text{Sr}$ values higher than for the carbonate bedrock average. The $^{87}\text{Sr}/^{86}\text{Sr}$ value of the Wasatch Formation (see below) is higher than the local Paleozoic carbonates and may be responsible for the general southward increase in the Sr concentrations and the higher $^{87}\text{Sr}/^{86}\text{Sr}$ values of the small, lake-marginal springs located along the southwestern margin of the lake (20, 21, 27, 28; Table 6; Figs. 8D and 8E).

Bear Lake Plateau

One sample of the Wasatch Formation (reddish sandstone) from the South Eden Creek drainage has a $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.71367 (Table 7). Several samples from the Twin Creek Limestone from both North and South Eden Creek drainages produced relatively similar $^{87}\text{Sr}/^{86}\text{Sr}$ values from 0.70712 to 0.70790 (Table 7), which match well with reconstructed Jurassic seawater $^{87}\text{Sr}/^{86}\text{Sr}$ values (Burke et al., 1982; Denison et al., 1998; Veizer et al., 1999). Two samples of the Nugget Sandstone from the North Eden Creek drainage yielded slightly higher $^{87}\text{Sr}/^{86}\text{Sr}$ values of 0.70986 and 0.71066 (Table 7).

Strontium isotope ratios in aqueous samples from the Bear Lake Plateau fall into two distinct groups separated by a $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.71000. Spatially, the 0.71000 boundary appears to coincide with the East Bear Lake Fault. All water samples east of the fault have $^{87}\text{Sr}/^{86}\text{Sr}$ values >0.71000 , and water samples near, or west of the fault have $^{87}\text{Sr}/^{86}\text{Sr}$ values <0.71000 . For example, the $^{87}\text{Sr}/^{86}\text{Sr}$ values in water from North Eden Creek is uniformly ~ 0.71000 from its headwaters (45, 46, 47) to near its mouth, then decreases to <0.71000 at sample sites near the East Bear Lake Fault (41, 40; Table 6). In addition, the water at the mouth of North Eden Creek (40) contains nearly three times more Sr than the springs feeding the stream (46, 47; Table 6). A similar trend in $^{87}\text{Sr}/^{86}\text{Sr}$ values and Sr concentrations occurs in the South Eden Creek data (35, 39; Table 6).

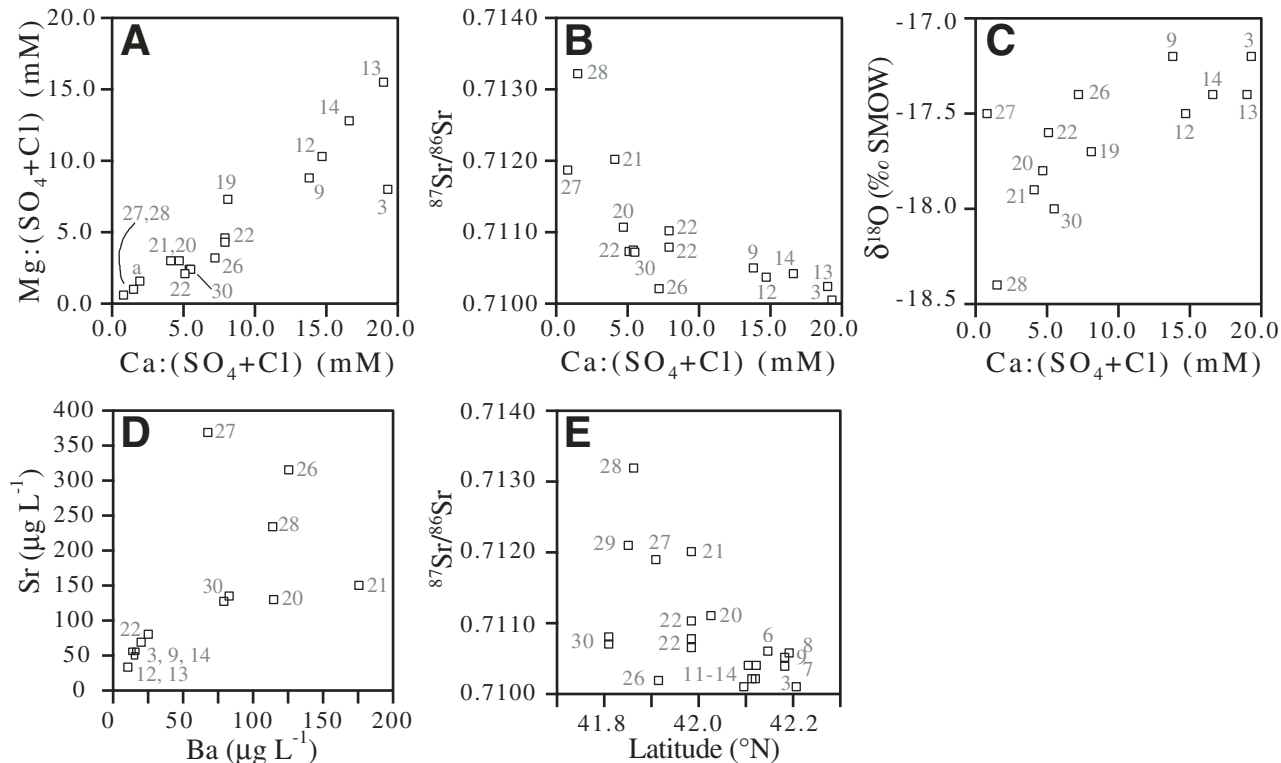


Figure 8. Solute and isotopic data for springs and streams along the eastern Bear River Range. (A) Mg:(SO₄+Cl) versus Ca:(SO₄+Cl). (B) $^{87}\text{Sr}/^{86}\text{Sr}$ versus Ca:(SO₄+Cl). (C) $\delta^{18}\text{O}$ versus Ca:(SO₄+Cl). (D) Sr versus Ba. (E) $^{87}\text{Sr}/^{86}\text{Sr}$ versus latitude. Numbers refer to sample sites on Figure 2. Data are presented in Tables 2, 3, and 6. SMOW—Standard Mean Ocean Water.

Streamflow at the mouth of Indian Creek (50) and the sulfate-rich spring at site 48 have low $^{87}\text{Sr}/^{86}\text{Sr}$ values of ~ 0.70770 and very high Sr concentrations (Table 6). A spring at the head of Indian Creek (52) has an equally low $^{87}\text{Sr}/^{86}\text{Sr}$ value (0.70783), but Sr concentration was not measured. These low $^{87}\text{Sr}/^{86}\text{Sr}$ values are similar to those in the Twin Creek Limestone that is prevalent in the area. In contrast, just to the north of Indian Creek, Mud Lake Hot Spring (53) has a $^{87}\text{Sr}/^{86}\text{Sr}$ value of ~ 0.70977 and a Sr concentration in excess of $4500 \mu\text{g L}^{-1}$ (Table 6). Falula Spring (32), located at the

southeast corner of Bear Lake, has a $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.71072 (Table 6), comparable to one of the Nugget Sandstone bedrock analyses. This value supports the proposal by Kaliser (1972) that Falula Spring may be fed by the Nugget Sandstone aquifer.

Tritium

Tritium concentrations in the atmosphere peaked in 1963–1964, at the end of atomic bomb testing, and have since decreased

TABLE 6. LITHIUM (Li), BARIUM (Ba), AND STRONTIUM (Sr) CONCENTRATIONS AND $^{87}\text{Sr}/^{86}\text{Sr}$ VALUES FOR SPRINGS AND STREAMS IN BEAR LAKE VALLEY

Site (Fig. 2)	Lat. ($^{\circ}\text{N}$)	Long. ($^{\circ}\text{W}$)	Elev. (m)	Site name	Date (m/d/yr)	Li ($\mu\text{g L}^{-1}$)	Ba ($\mu\text{g L}^{-1}$)	Sr ($\mu\text{g L}^{-1}$)	$^{87}\text{Sr}/^{86}\text{Sr}$
Bear River Range									
3	42.206	111.498	2001	Paris Spring	8/1/00	4	14	56	0.71005
6	42.146	111.575	2501	Bloomington Lake	8/23/04	2	5	13	0.71056
7	42.182	111.544	2202	Bloomington Creek	8/23/04	N.D.	N.D.	N.D.	0.71037
8	42.188	111.447	1879	Bloomington Creek	4/5/04	N.D.	N.D.	N.D.	0.71056
9	42.183	111.401	1815	Bloomington Creek	9/22/99	N.D.	16	57	0.71046
9	42.183	111.401	1815	Bloomington Creek	8/23/04	N.D.	N.D.	N.D.	0.71050
11	42.096	111.530	2092	St. Charles Creek	8/23/04	N.D.	N.D.	N.D.	0.71009
12	42.105	111.495	1975	Blue Pond Spring	4/5/04	0	11	33	0.71037
13	42.113	111.446	1940	St. Charles Creek	8/23/04	N.D.	N.D.	N.D.	0.71024
14	42.124	111.391	1817	St. Charles Creek	1996	N.D.	N.D.	N.D.	0.71042
14	42.124	111.391	1817	St. Charles Creek	9/22/99	N.D.	15	50	0.71036
14	42.124	111.391	1817	St. Charles Creek	8/1/00	0	11	33	0.71015
19	42.037	111.410	1815	Fish Haven Creek	9/22/99	N.D.	16	59	N.D.
20	42.026	111.402	1806	spring	4/5/04	6	115	129	0.71107
21	41.985	111.406	1806	spring	4/5/04	6	176	150	0.71202
22	41.985	111.427	1891	Swan Creek Spring	1996	N.D.	N.D.	N.D.	0.71104
22	41.985	111.427	1891	Swan Creek Spring	9/22/99	N.D.	25	81	0.71102
22	41.985	111.427	1891	Swan Creek Spring	8/1/00	0	20	68	0.71079
22	41.985	111.427	1891	Swan Creek Spring	4/5/04	2	25	80	0.71073
26	41.915	111.389	1806	spring	8/1/00	10	126	315	0.71021
27	41.909	111.372	1806	spring	4/5/04	30	68	369	0.71187
28	41.865	111.360	1801	spring	4/5/04	15	114	235	0.71322
29	41.851	111.356	1805	beach seep	8/23/04	138	72	1060	0.71210
30	41.809	111.389	1824	Big Spring	8/1/00	6	79	128	0.71072
30	41.809	111.389	1824	Big Spring	4/5/04	4	83	134	0.71075
31	41.846	111.337	1806	Big Creek	9/22/99	N.D.	88	149	0.71106
31	41.846	111.337	1806	Big Creek	4/5/04	N.D.	N.D.	N.D.	0.71104
Bear Lake Plateau									
32	41.842	111.302	1812	Falula Spring	5/28/03	N.D.	N.D.	N.D.	0.71072
34	41.878	111.294	1803	beach seep	5/28/03	130	90	3400	0.70950
35	N.D.	N.D.	N.D.	South Eden Creek	9/22/99	N.D.	117	861	0.70880
37	41.918	111.228	1937	spring (Twin Creek Ls)	4/5/04	N.D.	N.D.	N.D.	0.71114
38	41.921	111.189	2001	spring (Wasatch Fm.)	4/5/04	N.D.	N.D.	N.D.	0.71132
39	42.910	111.139	2054	Rabbit Spring	8/23/04	18	268	318	0.71060
40	N.D.	N.D.	N.D.	North Eden Creek	9/22/99	N.D.	156	645	0.70901
41	41.986	111.255	1842	North Eden Creek	8/23/04	22	119	369	0.70974
42	41.983	111.233	1873	North Eden Creek	5/28/03	N.D.	N.D.	N.D.	0.71013
43	41.984	111.212	1885	North Eden Creek	5/28/03	N.D.	N.D.	N.D.	0.71044
44	41.988	111.189	1896	North Eden Creek	5/28/03	N.D.	N.D.	N.D.	0.71044
45	41.993	111.148	2001	spring (Wasatch Fm.)	5/28/03	N.D.	N.D.	N.D.	0.71074
46	42.977	111.140	2001	spring (Wasatch Fm.)	4/5/04	16	239	292	0.71057
47	42.991	111.119	1983	spring (Wasatch Fm.)	4/5/04	16	186	250	0.71057
48	42.075	111.250	1835	spring	8/1/00	19	12	5933	0.70767
49	N.D.	N.D.	1804	beach seep	4/5/04	82	18	1850	0.70844
50	42.094	111.256	1830	Indian Creek	8/23/04	16	41	1830	0.70785
52	42.096	111.232	1964	spring	5/28/03	N.D.	N.D.	N.D.	0.70783
53	42.115	111.264	1824	Mud Lake Hot Spring	9/22/99	N.D.	25	4652	0.70976
53	42.115	111.264	1824	Mud Lake Hot Spring	8/1/00	265	28	4930	0.70978

Note: Lat.—latitude; Long.—longitude; N.D.—no data.

to pre-bomb era levels (Clark and Fritz, 1997). Clark and Fritz (1997) defined continental tritium values as follows: tritium values <0.8 tritium units (TU) are considered pre-bomb recharge, values of 5–15 TU are considered modern (<5–10 yr) recharge, and values >30 TU are considered to be recharge from the 1960s and 1970s. No long-term tritium data are available for the immediate study area, although the averaged annual tritium concentration in precipitation falling in Albuquerque, New Mexico, during 1990–2002 ranged from ~5–20 TU (IAEA, 2004) with the majority of the concentrations (82%) ranging between 6 and 13 TU. Rice and Spangler (1999) reported a single value of ~9 TU from winter precipitation collected in 1986 in Mantua Valley, ~100 km southwest of Bear Lake Valley. Sixteen precipitation samples (rain and snow) collected between September 2002 and June 2005 in Utah County, ~200 km southwest of Bear Lake, ranged from 2.1 to 11.7 TU, and averaged 7.2 TU (A. Mayo, 2008, personal commun.). A reasonable estimate for tritium concentration in modern precipitation falling in Bear Lake Valley is ~6–13 TU.

Of the five samples analyzed for tritium from Bear Lake Valley, two were from the largest springs in the Bear River Range: Swan Creek Spring (22) and Big Spring (30). Swan Creek Spring yielded a modern value of 11.5 TU (Table 5). Big Spring, which discharges along a fault in the Wasatch Formation at the south-

western end of the valley, had a lower value of 5.0 TU (Table 5). Three tritium samples were collected from the Bear Lake Plateau: one from a spring emanating from the Wasatch Formation at the head of North Eden Creek (47), and two from low-elevation springs near or along the East Bear Lake Fault (Mud Lake Hot Spring (53) and site 48). The spring at the head of North Eden Creek had a value of 2.6 TU (Table 5). Neither Mud Lake Hot Spring (53) nor the spring at site 48 contained detectable amounts of tritium (<0.6 TU; Table 5), suggesting that no modern recharge is present at these springs.

DISCUSSION

Stable Isotope Distribution in Springs and Streams of Bear Lake Valley

The $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of spring discharge, especially base flow, in the Bear River Range are likely homogenized values approaching the weighted average $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of winter precipitation over some interval of time (e.g., Winograd et al., 1998). Linking stable isotope values from spring discharge to recent spot-collections of winter precipitation is difficult because aquifers can store water for months to years (e.g., Rice and Spangler, 1999), resulting in stable isotope values in spring discharge

TABLE 7. STRONTIUM ISOTOPE RATIOS ($^{86}\text{Sr}/^{86}\text{Sr}$) OF BEDROCK UNITS IN THE BEAR RIVER RANGE AND BEAR LAKE PLATEAU

Unit	Location	Age	$^{87}\text{Sr}/^{86}\text{Sr}$	St. dev. (\pm)
Carbonate				
Twin Creek Limestone	South Eden Canyon, Bear Lake Plateau	Jurassic	0.70790	0.00001
Twin Creek Limestone	North Eden Canyon, Bear Lake Plateau	Jurassic	0.70712	0.00001
Twin Creek Limestone	North Eden Canyon, Bear Lake Plateau	Jurassic	0.70743	0.00001
Twin Creek Limestone	North Eden Canyon, Bear Lake Plateau	Jurassic	0.70720	0.00001
Lodgepole Limestone	Logan Canyon, Bear River Range	Mississippian	0.70811	0.00001
Hyrum Dolomite	Logan Canyon, Bear River Range	Devonian	0.70896	0.00001
Laketown Dolomite	Logan Canyon (float), Bear River Range	Silurian	0.70879	0.00001
Laketown Dolomite	Logan Canyon (float), Bear River Range	Silurian	0.70854	0.00001
Fish Haven Dolomite	Logan Canyon (float), Bear River Range	Ordovician	0.71011	0.00001
Garden City Limestone	Logan Canyon, Bear River Range	Ordovician	0.70965	0.00001
Garden City Limestone	Logan Canyon, Bear River Range	Ordovician	0.70901	0.00001
Garden City Limestone	Logan Canyon, Bear River Range	Ordovician	0.70918	0.00001
Bloomington Formation (oolitic limestone)	Logan Canyon, Bear River Range	Cambrian	0.71038	0.00001
Blacksmith Dolomite	Logan Canyon, Bear River Range	Cambrian	0.70975	0.00002
limestone facies of Langston Dolomite?	Bloomington Canyon, Bear River Range	Cambrian	0.70978	0.00001
Blacksmith Dolomite	Bloomington Canyon, Bear River Range	Cambrian	0.70906	0.00001
	average, Logan Canyon, Bear River Range (n = 12)		0.70928	0.00067
Non-carbonate				
Wasatch Formation	South Eden Canyon, Bear Lake Plateau	Tertiary (Eocene)	0.71367	0.00001
Nugget Sandstone	North Eden Canyon, Bear Lake Plateau	Triassic/Jurassic	0.70986	0.00001
Nugget Sandstone	North Eden Canyon, Bear Lake Plateau	Triassic/Jurassic	0.71066	0.00001
Swan Peak Quartzite	Bloomington Lake, Bear River Range	Ordovician	0.71262	0.00001
Swan Peak Quartzite	Logan Canyon, Bear River Range	Ordovician	0.71211	0.00001
Swan Peak Quartzite	St. Charles Canyon, Bear River Range	Ordovician	0.71153	0.00003
Swan Peak Quartzite, above shale	Logan Canyon, Bear River Range	Ordovician	0.71280	0.00001
Swan Peak Quartzite, interbedded with shale	Logan Canyon, Bear River Range	Ordovician	0.72206	0.00004
shale, base of Swan Peak Quartzite	Logan Canyon, Bear River Range	Ordovician	0.71544	0.00001
Bloomington Formation (shale)	Logan Canyon, Bear River Range	Cambrian	0.71743	0.00001
Geertzen Canyon Quartzite	Garden City, Bear River Range	Cambrian	0.71654	0.00003
Geertzen Canyon Quartzite	St. Charles Canyon, Bear River Range	Cambrian	0.71831	0.00013
	average, Logan Canyon, Bear River Range (n = 9)		0.71543	0.00351

Note: Unit names and ages from Dover (1995).

that are not representative of recent precipitation. The lack of continuous, long-term precipitation monitoring in the eastern Bear River Range hinders the discussion of the Bear River Range groundwater and stream stable isotope values, but two systematic trends are apparent. One is the progressive southward decrease in $\delta^{18}\text{O}$ and $\delta^2\text{H}$ from Bear River Range springs and streams given the geographically limited study area (Fig. 5A). And the other is the more negative stable isotope values in lower-elevation spring and stream samples within a particular watershed (Fig. 5B).

Maximum elevations in the eastern Bear River Range decrease southward by ~ 170 m, from Paris Peak (2918 m) at the head of Paris Canyon to Temple Peak (2751 m) located roughly 11 km due west of Big Spring. Intuitively, higher elevations should accumulate more isotopically negative precipitation (primarily snow) due to altitude-dependent fractionation effects (e.g., Gat, 1980; Rózanski et al., 1993; Poage and Chamberlain, 2001). Therefore, the higher elevations in the northern portion of the study area and within any individual watershed should gener-

ate more isotopically negative runoff than the lower elevations or more southerly locations.

The most likely explanations for the spatial distribution of groundwater $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values in the eastern Bear River Range are (1) the location of the study area, which is on the leeward side of the Bear River Range; and (2) the topography of the Bear River Range on a regional, rather than local, scale. The negative correlation between stable isotopes in precipitation and altitude occurs as air masses rise, cool, condense, and rain out while passing over a topographic barrier—the Bear River Range in particular. Once an air mass impacts the western, windward slope of the Bear River Range it must traverse an additional 30–35 km before reaching the study area (Fig. 9). Additional rainout while crossing the remaining topography of the Bear River Range would produce increasingly negative precipitation at increasing distance from the windward range front (e.g., Moran et al., 2007, and references therein). The topography of the Bear River Range is such that the windward range front due west of the southern study area

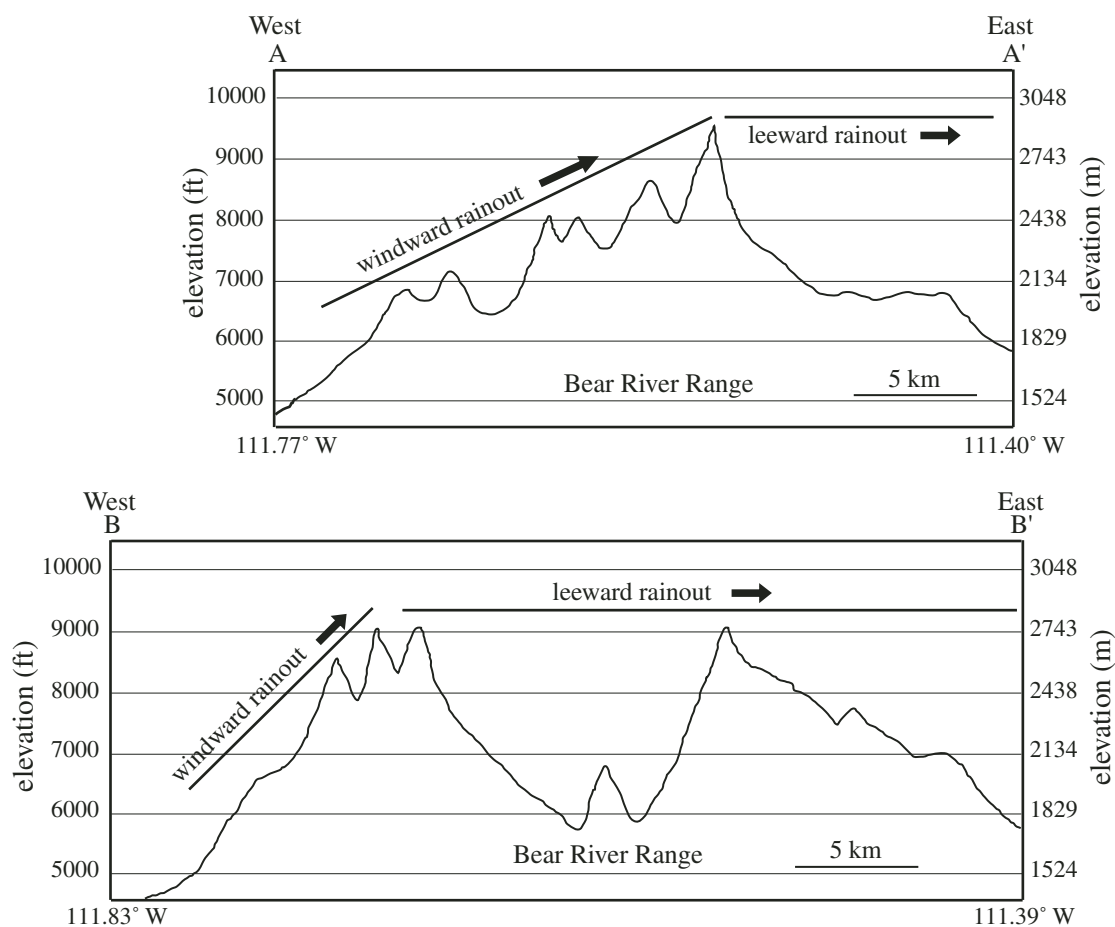


Figure 9. Elevational cross sections A–A' and B–B' through the Bear River Range. See Figures 1 and 2 for locations of sections. Vertical exaggeration 8 \times . Moisture moving west to east along cross section B–B' reaches maximum elevation farther west and experiences greater rainout before reaching Bear Lake than does precipitation passing over cross section A–A'. As a result, moisture condensing in the southern study area has experienced greater isotopic distillation (is more negative) than moisture condensing farther to the north. See text for discussion.

is very steep and maximum elevations are reached ~25 km west of the study area (Figs. 2 and 9; cross section B–B'). Farther to the north the elevational gradient of the windward range front is much more gentle and maximum elevations are not reached until ~12 km west of the study area (Figs. 2 and 9; cross section A–A'). Consequently, precipitation falling in the southern portion of the study area will have been subjected to greater leeward rainout and will be more isotopically negative (more depleted) than precipitation falling in the northern portion of the study area.

The decrease in $\delta^{18}\text{O}$ and $\delta^2\text{H}$ observed in springs and streams within an individual drainage may be explained, at least partially, by the same leeward rainout process. There is ~11 km of lateral distance between the maximum and minimum elevations in the northern portion of the study area (Fig. 9; cross section A–A'). If leeward rainout does occur, then the stable isotope values in precipitation would progressively decrease west of the topographic high. High-elevation springs and stream catchments in any given drainage would be recharged by slightly isotopically heavier precipitation than the more distant, lower-elevation springs and stream catchments. This relationship is supported by the limited data on hand, with the exception of Paris Creek drainage. A more comprehensive study of the stable isotope variability in local precipitation and spring discharge in the eastern Bear River Range is needed to further test this hypothesis, however.

The tritium data suggest that spring discharge from the Wasatch Formation on the Bear Lake Plateau is predominantly pre-bomb era water mixed with a small amount of modern recharge (Site 47; 2.6 TU). The age of the groundwater discharging at springs 48 and 53 along the East Bear Lake Fault is not known but the lack of detectable tritium indicates that no modern precipitation is present. Currently, there are no radiometric or other chronologic data to refine the "pre-bomb era" age for these springs. Consequently, a portion of the spring discharge on the Bear Lake Plateau is recharge that could be significantly older and no longer representative of modern climate dynamics.

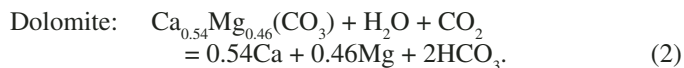
The $\delta^{18}\text{O}$ and $\delta^2\text{H}$ data from the Bear Lake Plateau springs fall below the GMWL on an evaporation line with a slope of 6.2 that crosses the GMWL at approximately $\delta^{18}\text{O} = -20\text{‰}$ and $\delta^2\text{H} = -150\text{‰}$ (Fig. 4B). These values indicate that the original isotopic composition of Bear Lake Plateau precipitation is significantly more depleted than the average precipitation collected at the Bear Lake County Airport (Fig. 2; Friedman et al., 2002). Precipitation that falls on the Bear Lake Plateau is sourced by storms passing over the Bear River Range, and the lower isotopic values are likely the result of continued rainout as storms pass over the range (e.g., Mayo and Loucks, 1995; Moran et al., 2007). Once storms pass over the Bear River Range the distance between cloud-base and the ground increases and the humidity is likely lower. Precipitation falling on the Bear Lake Plateau probably experiences evaporation (or sublimation) during air-fall before reaching the ground. Additional evaporation of snowmelt could occur during the spring if infiltration rates on the Bear Lake Plateau are slow. On the basis of the data presented here, it is reasonable to conclude that the stable isotope values on the Bear

Lake Plateau (excluding sites 48 and 53) result from local orographic and evaporative effects, although long-term Bear Lake Plateau precipitation isotope data and additional groundwater-age determinations are needed to test this hypothesis.

Solute Behavior of Swan Creek Spring

Swan Creek Spring (22) is one of the largest springs in Utah (Mundorff, 1971). The impressive discharge of Swan Creek Spring is indicative of a large and well-developed karst conduit system within the Bear River Range. The sensitivity of Swan Creek Spring to rainfall events (Kaliser, 1972) indicates a strong linkage to the surface. Infiltration into the Swan Creek Spring aquifer is likely quick, and given spring's cold temperature, passes quickly through the mountain range along shallow, and possibly short, flow paths. The spring is fed by a large solution channel (Kaliser, 1972) and is located along one of a series of north-south-trending faults in the Bear River Range, west of Bear Lake (Dover, 1995). The location of this spring along a fault provides an opportunity to study the effect of faulting on spring chemistry. The EPA has monitored Swan Creek Spring for nearly 30 years (epa.gov/storet/dw_home.html; station 4907200) and has generated a large chemical data set. Many of the EPA analyses do not include K, SO_4 or Cl, so several charge balance errors are greater than 5%. The majority of the more complete analyses have charge balance errors less than 5%, however. The most complete EPA analyses for Swan Creek Spring are presented in Table 8.

The molar ratio of calcium (Ca) to magnesium (Mg) in ideal dolomite is 1.0, with calcian dolomites having Ca:Mg values slightly above 1.0 (Goldsmith and Graf, 1958; Sperber et al., 1984). All but one of the dolomites (characterized by slight to very slight effervescence) in the Bear Lake drainage have Ca:Mg values slightly above 1.0 (Table 9). Aqueous dissolution of the local limestone and the average local dolomite follows the equations:



Water dissolving equal amounts of the local limestone and dolomite would acquire 1 mol of Ca from limestone and ~0.54 mol of Ca and 0.46 mol of Mg from dolomite, resulting in water with a Ca:Mg value of 3.35 (e.g., Szramek et al., 2007). The Ca:Mg value decreases as the mixture becomes more enriched in dolomite.

Dolomite dissolution kinetics are not well understood (Morse and Arvidson, 2002) but two variables likely explain most of the Swan Creek Spring solute behavior. First, dolomite is more soluble than calcite at temperatures below 15 °C (Langmuir, 1997), and second, dissolution of dolomite does not appear to be congruent, especially early in the dissolution process. The CaCO_3 phase

of dolomite is apparently more soluble than the $MgCO_3$ phase, with the dissolution of $MgCO_3$ being a slower, rate-limiting step (Busenberg and Plummer, 1982; Morse and Arvidson, 2002). The higher Mg concentrations in Swan Creek Spring base flow (Fig. 10) suggests that this groundwater has been in contact with the dolomitic bedrock for an extended period of time, likely on the order of several years (e.g., Herman and White, 1985). The relatively constant Ca concentrations and decreased Mg concentrations at higher discharge (Fig. 10) likely reflect cold snowmelt passing through the karst conduit network within the Bear River Range (e.g., White, 2002; Ozyurt and Bayari, 2007), and

increased dissolution of limestone and the $CaCO_3$ phase of dolomite during the peak snowmelt months.

Sodium (Na) and chloride (Cl) concentrations in Swan Creek Spring exhibit unexpected behavior in that the highest Na and Cl concentrations occur in a relatively narrow discharge window of $\sim 1\text{--}2.5\text{ m}^3\text{s}^{-1}$ (Fig. 11). Chloride concentrations at Swan Creek Spring generally track the snow-water-equivalent data in the Bear River Range for the same years (Fig. 12). Assuming that analytical errors are not the cause, then there are three likely causes for the discharge-dependent increase in Na and Cl at Swan Spring. These potential sources include an atmospheric

TABLE 8. DISCHARGE AND MAJOR-ION CHEMISTRY DATA FOR SWAN CREEK SPRING, UTAH

Collection (mo-yr)	Discharge ($\text{m}^3\text{ s}^{-1}$)	Ca (mg L^{-1})	Mg (mg L^{-1})	Na (mg L^{-1})	K (mg L^{-1})	HCO_3^- (mg L^{-1})	SO_4 (mg L^{-1})	Cl (mg L^{-1})	balance* (% error)	Ca (mM)	Mg (mM)	HCO_3^- (mM)	Na (mM)	Cl (mM)
2-75	N.D.	48	16	2.0	1	228	7.0	4.0	-2.0	1.20	0.66	3.74	0.09	0.11
5-75	N.D.	51	10	1.0	1	202	5.0	3.0	-0.9	1.27	0.41	3.31	0.04	0.08
6-75	N.D.	42	9	1.0	1	156	4.0	3.0	3.4	1.05	0.37	2.56	0.04	0.08
10-75	N.D.	46	17	2.0	1	222	8.0	2.0	-0.7	1.15	0.70	3.64	0.09	0.06
11-76	N.D.	45	16	3.0	1	212	8.0	3.0	0.1	1.12	0.66	3.47	0.13	0.08
1-77	N.D.	45	18	2.0	1	226	9.0	2.0	-1.3	1.12	0.74	3.70	0.09	0.06
3-77	N.D.	48	17	2.0	2	228	16.0	3.0	-2.6	1.20	0.70	3.74	0.09	0.08
5-77	N.D.	49	19	3.0	1	236	10.0	4.0	-0.2	1.22	0.78	3.87	0.13	0.11
7-77	N.D.	51	19	4.0	1	244	7.0	4.0	0.5	1.27	0.78	4.00	0.17	0.11
11-77	N.D.	46	17	4.0	N.D.	218	8.0	3.0	0.7	1.15	0.70	3.57	0.17	0.08
3-78	N.D.	42	19	5.0	1	222	16.0	3.0	-1.8	1.05	0.78	3.64	0.22	0.08
7-78	N.D.	40	14	2.0	N.D.	184	10.0	2.0	-0.8	1.00	0.58	3.02	0.09	0.06
8-78	N.D.	50	14	3.0	N.D.	222	5.0	3.0	-0.5	1.25	0.58	3.64	0.13	0.08
3-79	N.D.	46	18	3.0	1	216	8.0	3.0	1.9	1.15	0.74	3.54	0.13	0.08
8-79	0.42	50	16	3.0	N.D.	208	10.0	4.0	2.9	1.25	0.66	3.41	0.13	0.11
11-79	N.D.	42	18	4.0	1	N.D.	13.0	5.0	-1.3	1.05	0.74	N.D.	0.17	0.14
2-80	N.D.	46	20	3.0	N.D.	216	9.0	5.0	2.6	1.15	0.82	3.54	0.13	0.14
5-80	N.D.	50	7	1.0	2	169	4.0	3.0	3.9	1.25	0.29	2.77	0.04	0.08
8-80	N.D.	46	14	2.0	1	200	10.0	4.0	-0.4	1.15	0.58	3.28	0.09	0.11
10-80	N.D.	42	17	3.0	N.D.	202	10.0	4.0	0.0	1.05	0.70	3.31	0.13	0.11
12-80	N.D.	32	18	3.0	N.D.	172	11.0	3.0	1.3	0.80	0.74	2.82	0.13	0.08
2-81	N.D.	49	17	3.0	N.D.	224	11.0	2.0	0.3	1.22	0.70	3.67	0.13	0.06
12-81	0.37	32	20	4.0	N.D.	182	15.0	3.0	0.7	0.80	0.82	2.98	0.17	0.08
7-82	N.D.	38	15	2.0	N.D.	182	10.0	3.0	-0.8	0.95	0.62	2.98	0.09	0.08
9-82	N.D.	45	15	3.0	N.D.	218	13.0	3.0	-4.1	1.12	0.62	3.57	0.13	0.08
10-82	N.D.	48	17	3.0	N.D.	216	13.0	4.0	0.1	1.20	0.70	3.54	0.13	0.11
12-82	N.D.	47	17	7.0	N.D.	214	10.0	5.0	2.4	1.17	0.70	3.51	0.30	0.14
2-83	N.D.	46	17	2.0	N.D.	212	13.0	4.0	-0.8	1.15	0.70	3.47	0.09	0.11
6-83	N.D.	47	8	3.0	N.D.	182	5.0	3.0	-0.3	1.17	0.33	2.98	0.13	0.08
8-88	N.D.	43.8	16.7	3.0	N.D.	226	N.D.	3.5	-1.5	1.09	0.69	3.70	0.13	0.10
9-98	0.75	49.8	17.4	2.8	N.D.	222	N.D.	4.0	3.7	1.24	0.72	3.64	0.12	0.11
10-98	1.31	48.3	18	3.3	N.D.	222	N.D.	4.0	3.6	1.21	0.74	3.64	0.14	0.11
12-98	1.04	49	18	4.1	N.D.	222	N.D.	6.0	3.8	1.22	0.74	3.64	0.18	0.17
1-99	0.82	48	19.3	3.2	N.D.	232	15.2	5.0	-1.5	1.20	0.79	3.80	0.14	0.14
2-99	0.79	49	19.3	4.1	N.D.	228	N.D.	6.5	3.7	1.22	0.79	3.74	0.18	0.18
3-99	1.16	51.1	19	8.6	N.D.	234	N.D.	13.5	3.0	1.28	0.78	3.84	0.37	0.38
4-99	2.06	57	16.7	9.1	N.D.	226	N.D.	15.5	5.4	1.42	0.69	3.70	0.40	0.44
5-99	2.41	58.1	13.7	5.5	N.D.	212	N.D.	10.0	6.5	1.45	0.56	3.47	0.24	0.28
5-99	7.08	55.2	8.4	2.7	N.D.	198	N.D.	4.5	2.6	1.38	0.34	3.25	0.12	0.13
6-99	9.08	48.9	8.2	2.7	N.D.	181	N.D.	4.5	2.1	1.22	0.34	2.97	0.12	0.13
6-99	3.11	46.7	9.8	2.4	N.D.	176	N.D.	N.D.	5.7	1.17	0.40	2.88	0.10	N.D.
7-03	0.42	46.8	16.3	3.4	N.D.	220	N.D.	N.D.	3.0	1.17	0.67	3.61	0.15	N.D.
8-03	0.10	50.5	17.6	3.9	N.D.	228	N.D.	N.D.	5.1	1.26	0.72	3.74	0.17	N.D.
9-03	0.48	50.2	18	3.6	N.D.	241	N.D.	N.D.	2.2	1.25	0.74	3.95	0.16	N.D.
10-03	0.24	47	17.6	3.2	N.D.	236	N.D.	N.D.	0.9	1.17	0.72	3.87	0.14	N.D.
12-03	0.79	45.6	18.8	3.4	N.D.	224	N.D.	N.D.	4.1	1.14	0.77	3.67	0.15	N.D.
1-04	0.63	43.1	10.5	3.2	N.D.	214	N.D.	N.D.	-5.4	1.08	0.43	3.51	0.14	N.D.
2-04	0.30	47.4	18.8	3.4	N.D.	218	N.D.	N.D.	6.5	1.18	0.77	3.57	0.15	N.D.
3-04	1.03	50.2	17.6	7.4	N.D.	216	23.5	12.6	-1.4	1.25	0.72	3.54	0.32	0.36
4-04	2.09	55.2	13.5	4.7	N.D.	208	N.D.	N.D.	8.7	1.38	0.56	3.41	0.20	N.D.
4-04	1.59	46.2	13.6	4.0	N.D.	204	N.D.	N.D.	3.7	1.15	0.56	3.34	0.17	N.D.
5-04	3.70	49.8	10.5	3.0	N.D.	188	N.D.	N.D.	6.1	1.24	0.43	3.08	0.13	N.D.
5-04	1.16	47.1	11.4	3.3	N.D.	181	N.D.	N.D.	7.2	1.18	0.47	2.97	0.14	N.D.
6-04	1.97	48.3	11	3.2	N.D.	164	N.D.	N.D.	12.5	1.21	0.45	2.69	0.14	N.D.
6-04	0.78	50.9	13.6	3.7	N.D.	185	N.D.	N.D.	11.5	1.27	0.56	3.03	0.16	N.D.

Note: Data from Environmental Protection Agency (www.epa.gov/storet/dw_home.html; station 4907200); N.D.—no data.

*Charge balance percent error calculated as $[(\text{cation sum} - \text{anion sum})/(\text{cation sum} + \text{anion sum})] \times 100$, where units of measurement are in milliequivalents.

influx (e.g., dust from the Great Salt Lake basin), an influx of road salt from Highway 89, or another aquifer-solute source. Atmospheric Cl concentrations in precipitation at Logan, Utah (Fig. 1) tend to track the snow-water-equivalent data for the Bear River Range (Fig. 13). The Cl concentration in Logan precipitation is roughly an order of magnitude lower than the Cl concentrations in Swan Creek Spring, however. Some of the Cl at Swan Creek Spring is undoubtedly derived from atmospheric sources, but the atmospheric influx is not large enough to cause the Cl fluctuations in the spring discharge. Road maintenance along Highway 89 is another likely source of Na and Cl. Highway 89 is maintained throughout the year and salted during the winter. Salt-laden snowmelt could easily infiltrate into the Swan Creek Spring groundwater basin through the solution caverns that are in close proximity to Highway 89 (Fig. 2). Finally, solutes derived from the southern end of the valley may also be responsible. The Wasatch Formation is prevalent at the southern end of the valley, and spring water emanating from it has high Na and Cl concentrations. Wasatch Formation groundwater from the southern valley may be able to move northward along the faults that bound the western margin of the Bear River Range and discharge at Swan Creek Spring.

Differentiating between the road salt and Wasatch Formation groundwater hypotheses for the Na-Cl behavior at Swan

Creek Spring should be possible using SO₄ data. Wasatch Formation groundwater is also high in SO₄ whereas typical road salt would only be a source of Cl. The road salt hypothesis would be supported if there were a no correlation between SO₄ and spring discharge, whereas the Wasatch Formation groundwater hypothesis would be favored if these variables did covary. Unfortunately this approach is not currently possible due to a lack of SO₄ data for Swan Creek Spring (Table 8).

Paris Spring and Blue Pond Spring

Groundwater discharge at Paris Spring (3) is probably controlled by the Lead Bell Shale (Wylie et al., 2005) and discharge at Blue Pond Spring (12) is likely fault controlled (Oriol and Platt, 1980). Time-series data for the solute composition and discharge rates of these springs are not available. Several observations are possible with the limited data that are available, however. The immediate bedrock lithology surrounding Paris Spring is limestone and dolomite (Oriol and Platt, 1980). Paris Spring's major-ion chemistry has been reported twice, once from a September collection (this study) and once from an August 2002 collection (Wylie et al., 2005). Both collections were taken in the late summer and should reflect base flow conditions. The Ca and Mg concentrations from both collections suggest that dolomite

TABLE 9. CALCIUM (Ca) AND MAGNESIUM (Mg) ASSAYS FOR CARBONATE BEDROCK IN BEAR LAKE DRAINAGE

Formation	Lithology	Effervescence	CaO (wt%)	MgO (wt%)	Ca (mol)	Mg (mol)	Ca:Mg
Laketown Dolomite	dolomite	very slightly	22.68	15.95	0.40	0.40	1.00
Nounan Limestone	dolomite	very slightly	31.74	20.36	0.57	0.51	1.12
Nounan Limestone	dolomite	very slightly	33.71	21.52	0.60	0.53	1.13
Fish Haven Dolomite	dolomite	very slightly	31.51	19.77	0.56	0.49	1.14
Jefferson Dolomite	dolomite	very slightly	28.87	17.75	0.51	0.44	1.16
Wells Formation	dolomite	very slightly	31.31	19.37	0.56	0.48	1.17
Nounan Limestone*	dolomite	slightly	32.01	19.44	0.57	0.48	1.19
Langston Limestone [†]	limestone	very slightly	31.58	18.89	0.56	0.47	1.19
Bloomington Formation	limestone	very slightly	31.45	18.14	0.56	0.45	1.24
Fish Haven Dolomite/Laketown Dolomite	dolomite	very slightly	32.70	18.35	0.58	0.46	1.26
St. Charles Limestone [§]	dolomite	slightly	31.35	17.57	0.58	0.44	1.27
Brazer Limestone [¶]	dolomite	very slightly	26.00	13.20	0.56	0.33	1.39
Bloomington Formation	limestone	strongly	43.26	9.07	0.46	0.23	3.35
Ute Limestone**/Langston Limestone [†]	dolomite	strongly	44.42	2.26	0.77	0.06	13.17
Bloomington Formation	limestone	strongly	47.84	2.43	0.85	0.06	14.17
Blacksmith Limestone ^{††}	dolomite	very strongly	49.41	2.25	0.88	0.06	14.67
Madison Limestone	limestone	strongly	59.62	1.19	1.06	0.03	35.33
Garden City Limestone ^{§§}	limestone	strongly	49.51	0.62	0.88	0.02	44.00
Garden City Limestone ^{§§}	limestone	strongly	50.12	0.52	0.89	0.01	89.00
St. Charles Limestone [§]	limestone	strongly	50.64	0.47	0.90	0.01	90.00
Wells Formation	limestone	strongly	52.60	0.40	0.94	0.01	94.00
Average Ca, Mg, and Ca:Mg of slightly and very slightly reactive units (n = 12)					0.56	0.44	1.19 ± 0.10

Note: Formation, effervescence, lithology, %CaO, %MgO data from Kaliser (1972).

*Nounan Limestone is currently named Nounan Dolomite (Dover, 1995).

[†]Langston Limestone is currently named Langston Dolomite (Dover, 1995).

[§]St. Charles Limestone is currently named St. Charles Formation (Dover, 1995).

[¶]Brazer Limestone is currently named Brazer Dolomite (Dover, 1995).

**Ute Limestone is currently named Ute Formation (Dover, 1995).

^{††}Blacksmith Limestone is currently named Blacksmith Dolomite (Dover, 1995).

^{§§}Garden City Limestone is currently named Garden City Formation (Dover, 1995).

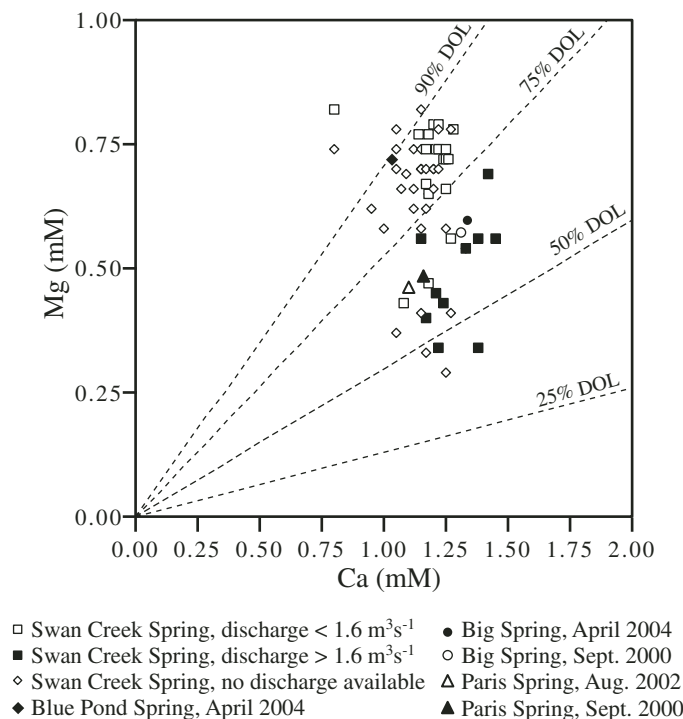


Figure 10. Magnesium (Mg) versus calcium (Ca) concentrations for four large springs in the Bear River Range. Open and solid squares represent data from Swan Creek Spring where spring discharge is known. Open diamonds represent data from Swan Creek Spring where spring discharge is not known. Solid diamond represents data from Blue Pond Spring, spring discharge not known. Open and solid circles represent data from Big Spring, spring discharge not known. Open and solid triangles represent data from Paris Spring, spring discharge not known. Dashed lines represent various Ca:Mg ratios created by the dissolution of different amounts of dolomite (DOL) and limestone. See text for discussion. Data are presented in Tables 3 and 8 and in Appendix DR1 (see footnote 1).

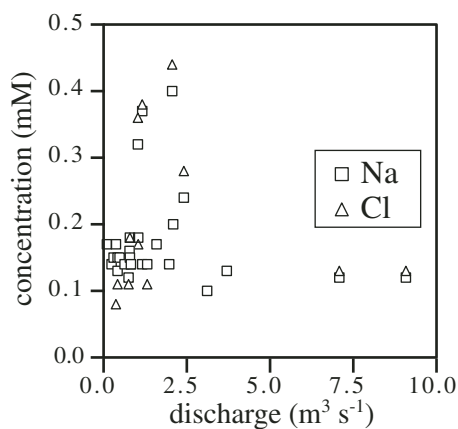


Figure 11. Sodium (Na) and chloride (Cl) concentrations versus spring discharge at Swan Creek Spring. Note the peak in concentrations for both ions during discharges of $\sim 2.0 \text{ m}^3 \text{ s}^{-1}$.

dissolution provides the majority of the base flow solutes (Fig. 10). In contrast, the bedrock lithology in the area of Blue Pond Spring is exclusively dolomite (Oriol and Platt, 1980). The major-solute chemistry of Blue Pond Spring (12) has been analyzed only once, during what should have been peak discharge conditions during April, 2004. The Ca and Mg concentrations from that collection indicate that Blue Pond Spring's solute chemistry is primarily derived from the dissolution of dolomite (Fig. 10), even during peak discharge. This implies that the groundwater basin that feeds Blue Pond Spring may be relatively small and local, or alternatively, if Blue Pond Spring's groundwater basin is large then the conduit system that feeds the spring is developed within dolomite. Repeated sampling and gauging of Paris Spring and Blue Pond Spring should be conducted during different seasons to test for discharge-dependent changes in major-ion chemistry, like that observed at Swan Creek Spring. Such information would be crucial for further understanding the karst development within the Bear River Range.

Big Spring

Streams in the eastern Bear River Range are conspicuously located in the carbonate terrain north of Garden City (Fig. 2). Perennial streams and large springs are absent in the Wasatch Formation terrain between Big Spring (30) and Garden City. Solution basins in the Bear River Range west and north of Garden City are thought to be primary recharge areas and conduits for snowmelt for the more northern springs and streams (Reheis et al., this volume). Similar solution basins (Bear Wallow and Peter Sinks) and one sinkhole region are mapped along the Bear River Range ridge crest west and southwest of Garden City (Dover, 1995; Fig. 2), yet with the exception of Big Spring, there are no substantial springs or streams in the area. The groundwater divide between Big Spring and Swan Creek Spring may lie relatively close to Big Spring such that a large portion of the infiltration from the southern portion of the valley flows northward and discharges at Swan Creek Spring. Additionally, the paucity of large springs and streams south of Garden City might be explained by the presence of the Wasatch Formation in this area: with its relatively low permeability it acts as a confining bed where it overlies the local Paleozoic carbonates so groundwater cannot reach the surface. Big Spring (30), the only major groundwater discharge point within the Wasatch Formation, emanates from a fault that may penetrate to the Paleozoic carbonate aquifer.

In solute chemistry, Big Spring is more similar to other large springs discharging from carbonate terrain, such as Paris Spring (3), Blue Pond Spring (12), and Swan Creek Spring (22) than to springs emanating from the Wasatch Formation (e.g., 39, 46, 47; Figs. 7 and 8). The $^{87}\text{Sr}/^{86}\text{Sr}$ value of Big Spring is indistinguishable from values of springs sourced in either Paleozoic carbonate or Wasatch Formation rocks, however, and is not a useful indicator of Big Spring's source. The tritium value from Big Spring (~ 5 TU) is half of the modern value from Swan

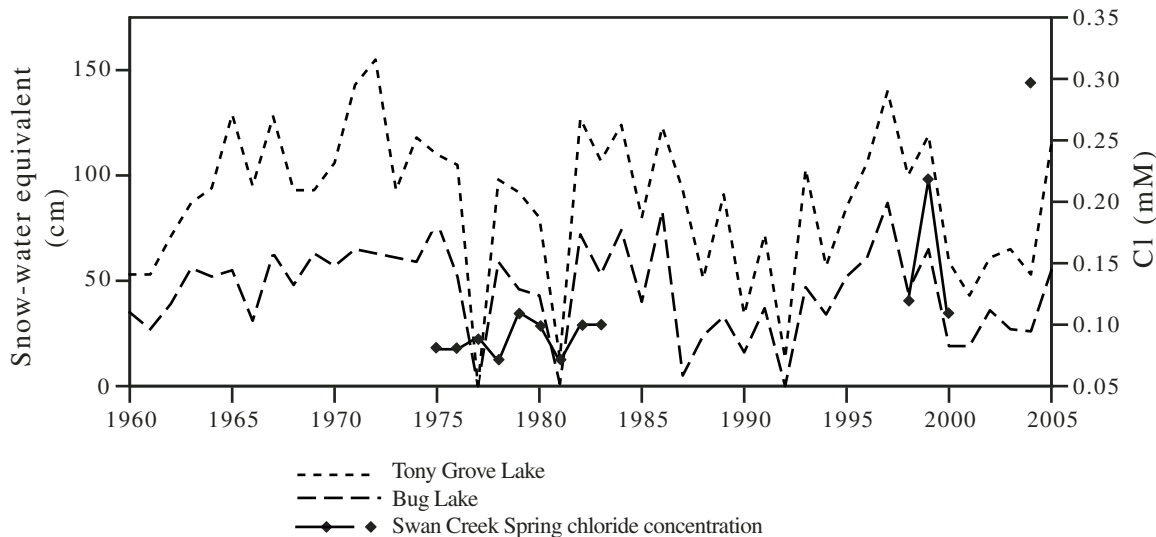


Figure 12. Snow-water-equivalent data for Tony Grove Lake (short-dashed line), Bug Lake (long-dashed line), and chloride concentration data from Swan Creek Spring (solid line with solid symbols).

Creek Spring, however, but nearly twice that of another Wasatch Formation spring (47).

The large volume of water issuing from Big Spring also suggests a strong connection to the Paleozoic carbonate aquifer, but note that the Ca and Mg concentrations at Big Spring during assumed peak discharge (April) and base flow (September) conditions are not discharge dependent (Fig. 10). The groundwater basin and conduit-fracture network that feeds Big Spring may be significantly different from the system that feeds Swan Creek Spring. Additional data are needed to test this hypothesis, however.

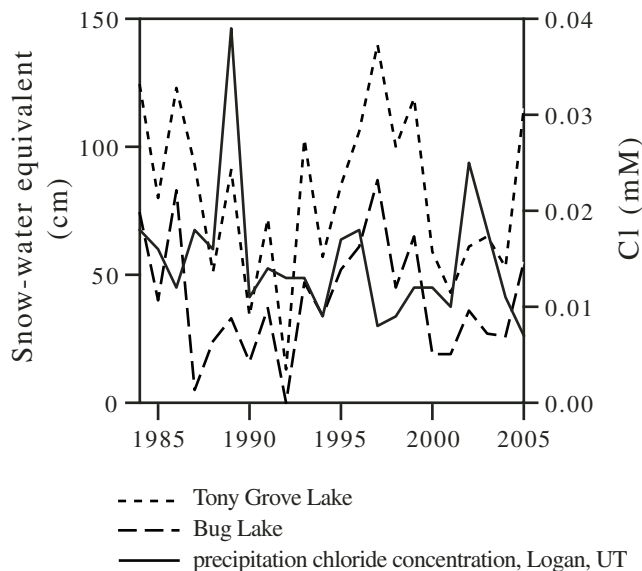


Figure 13. Snow-water-equivalent data for Tony Grove Lake (short-dashed line), Bug Lake (long-dashed line), and chloride concentration in precipitation at Logan, Utah (solid line).

Impacts of Groundwater on Bear Lake's Hydrologic and Solute Balance

The modern hydrologic balance of Bear Lake (excluding the Bear River) is probably balanced, or nearly so (Lamarra et al., 1986; Bright, 2003; Bright et al., 2006) even though the value of Swan Creek discharge used in previous estimates was underestimated by a factor of nearly three. Mundorff (1971) reported an average Swan Creek Spring discharge of $\sim 0.9 \text{ m}^3 \text{ s}^{-1}$, whereas Bright (2003) estimated $\sim 0.3 \text{ m}^3 \text{ s}^{-1}$. Given the higher Swan Creek Spring discharge, the hydrologic balance of Bear Lake is more easily explained by surface runoff and shallow subsurface sources, without the need for substantial amounts of groundwater influxes from lake-marginal or sublacustrine springs. Lake-marginal influxes to the lake do occur, but given the isotopic and solute chemistry of the eastern and southern lake-marginal springs reported here, they are evidently of minor importance.

Dean et al. (2007) used an analysis of pre-diversion lake chemistry and the $^{87}\text{Sr}/^{86}\text{Sr}$ values of modern water to conclude that $\sim 99\%$ of the solutes in Bear Lake prior to the diversion of the Bear River were derived from the streams sourced in the Bear River Range. This estimate is based on $^{87}\text{Sr}/^{86}\text{Sr}$ and Cl balances using mainly stream-solute data. The Dean et al. (2007) estimate did not include the lake-marginal springs that although small, contain concentrated solutes. The solute and isotope data from the lake-marginal springs presented in this study reinforce their interpretation. Using the average $^{87}\text{Sr}/^{86}\text{Sr}$ value of pre-diversion aragonite (0.71031; Table 10A) in sediment core BL96-10 as a reference, the influx of solutes (Na, K, and Cl) from the local streams and lake-marginal springs was mixed in various proportions to generate a hypothetical water body with a $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.71031 and with similar Na, K, and Cl concentration factors in relation to the lake's 1912 chemistry (Table 10B and 10C).

TABLE 10A. STRONTIUM ISOTOPE ($^{87}\text{Sr}/^{86}\text{Sr}$) VALUES
IN BEAR LAKE SEDIMENT CORE BL96-10

Depth (cm)	Status	$^{87}\text{Sr}/^{86}\text{Sr}$ ratio	St. dev. (\pm)
0	post-diversion	0.70942	0.00001
5	post-diversion	0.70942	0.00001
7	post-diversion	0.70943	0.00001
9	post-diversion	0.70951	0.00001
11	post-diversion	0.70975	0.00001
13	pre-diversion	0.71023	0.00001
16	pre-diversion	0.71031	0.00001
20	pre-diversion	0.71034	0.00001
25	pre-diversion	0.71035	0.00001
29	pre-diversion	0.71032	0.00001
pre-diversion avg.		0.71031	

TABLE 10B. MODELING DATA FOR BEAR LAKE SOLUTE SOURCES

Site (Fig. 2)	Name	Sr ($\mu\text{g L}^{-1}$)	f Sr total*	$^{87}\text{Sr}/^{86}\text{Sr}$	Contribution [†]	Cl (mg L^{-1})	Na (mg L^{-1})	K (mg L^{-1})
<u>West streams</u>								
14	St. Charles Creek	50	0.18727	0.71036	0.13303	1.8	2.1	0.3
22	Swan Creek Spring/Creek	68	0.25468	0.71079	0.18103	5.2	4.3	0.5
31	Big Creek	149	0.55805	0.71106	0.39681	9.1	6.4	1.0
sum		267			0.71087			
average		89				5.3	4.2	0.6
<u>West springs</u>								
20	spring	129	0.10768	0.71107	0.07657	7.0	7.6	1.3
21	spring	150	0.12521	0.71202	0.08915	7.8	6.9	1.4
26	spring	315	0.26294	0.71021	0.18674	7.6	9.8	0.8
27	spring	369	0.30801	0.71187	0.21926	46.0	38.1	2.7
28	spring	235	0.19616	0.71322	0.13991	23.0	19.3	2.0
sum		1198			0.71163			
average		240				18.3	16.3	1.6
<u>East streams</u>								
35	South Eden Creek	861	0.25809	0.70880	0.18293	55.4	32.5	2.0
40	North Eden Creek	645	0.19335	0.70901	0.13709	51.9	32.6	2.1
50	Indian Creek	1830	0.54856	0.70785	0.38830	22.4	21.1	1.8
sum		3336			0.70832			
average		1112				43.2	28.7	2.0
<u>East springs</u>								
48	spring	5933		0.70767		11.5	26.0	2.4
53	spring	4930		0.70977		74.1	157.0	41.3

*f Sr total = fraction of Sr total, calculated by (sample Sr concentration/Sr sum) for each geographic grouping.

[†]Contribution = (f Sr total * $^{87}\text{Sr}/^{86}\text{Sr}$) for each site.

TABLE 10C. TYPICAL MIXING MODEL RESULTS FOR PRE-DIVERSION BEAR LAKE

Source	Sr ($\mu\text{g L}^{-1}$)	$^{87}\text{Sr}/^{86}\text{Sr}$ (avg.)	Cl (mg L^{-1})	Na (mg L^{-1})	K (mg L^{-1})	Proportion	Cl:Na	Cl:K	Na:K
West streams	89	0.71087	5.3	4.2	0.6	0.9906			
East streams	1112	0.70832	43.2	28.7	2.0	0.0018			
West springs	240	0.71163	18.3	16.3	1.6	0.0014			
spring 48 (Fig. 2)	5933	0.70767	11.5	26.0	2.4	0.0023			
spring 53 (Fig. 2)	4930	0.70977	74.1	157.0	41.3	0.0039			
Mixture		0.71031	5.8	5.0	0.8	1.0000	1.2	7.3	6.3
Pre-diversion lake*		0.71031	78.5	66.3	10.5		1.2	7.5	6.3
Concentration factor [†]			13.5	13.3	13.1				

*Pre-diversion lake major-ion concentrations from Birdsey (1989).

[†]Concentration factor = Pre-diversion lake concentration for given ion/mixture concentration for same ion.

These results indicate that the western lake-marginal springs (20, 21, 26–28) and eastern streams (35, 40, 50) and springs (48, 53) collectively contributed only ~1% of the $^{87}\text{Sr}/^{86}\text{Sr}$ in the pre-diversion lake (Table 10C). The mixing models also reveal that Mud Lake Hot Spring-type water (53) is necessary to generate the K concentrations and Na, K, and Cl solute ratios reported for the pre-diversion lake (Table 10C). No other water sources reported in this study have the K concentrations needed to balance the model. This conclusion is based on the assumption that Na, K, and Cl behave conservatively in the lake (see Dean et al., 2007). Mud Lake Hot Spring (53) is, and has been, separated from the lake by a sandbar but prior to the construction of the water control structures at Lifton (15) ca. 1912 (Fig. 2) there was a small outlet on the west side of the lake that connected the lake to the Bear River via the marshes surrounding Mud Lake (McConnell et al., 1957). Solutes from Mud Lake Hot Spring must have entered the lake through this outlet, percolated through the sandbar, or, alternatively, there is hot spring-type water entering the lake from unlocated springs within the lake basin or possibly from groundwater leakage along the East Bear Lake Fault.

Sublacustrine springs were thought to discharge in the lake because isolated portions of the lake surface typically do not freeze during the winter. During calm lake conditions the surface of the lake at these ice-free areas visibly roils, preventing the lake surface from freezing in the winter. Sonar images at one of these ice-free areas revealed a strong reflector emanating from the lake floor (Figs. 2 and 14). Subsequent investigations by divers, however, indicated no detectable discharge of water at these sites. Bubble trails associated with the sonar reflections are composed of isotopically depleted methane gas (Dean, this volume). A density contrast between ambient lake water and methane-charged water is most likely responsible for the sonar reflection and surface-water disturbances. Another peculiar location in Bear Lake, termed “the rock pile” (Dean, this volume), may be an example of diffuse, sublacustrine spring discharge. Divers detected no noticeable groundwater influx at this site, however, and the site may no longer be active. To date, no large-volume springs have been identified on the floor of the lake. Sublacustrine spring discharge, if occurring, is probably of minor importance to the hydrologic balance of Bear Lake.

SUMMARY

1. The two primary rock types in southern Bear Lake Valley—Paleozoic marine carbonates, which are exposed primarily north of Garden City, and the Wasatch Formation, which is exposed at the southern end of Bear Lake Valley and on the Bear Lake Plateau—contain groundwater with two distinct solute compositions. The Ca-Mg-HCO₃ carbonate-terrain water contrasts sharply with the Ca-Mg-HCO₃-SO₄-Cl and Ba- and Sr-enriched water associated with Wasatch Formation springs. Water discharging at several fault-related, lake-marginal springs along the western margin of Bear Lake appears to be a mix of carbonate bedrock- and Wasatch Formation-sourced solutes. The

Ca-SO₄-rich groundwater in the northeast quadrant of Bear Lake is a third distinct water type in the watershed. Its distribution appears to extend southward along the eastern flank of Bear Lake along the East Bear Lake Fault. A hot spring with a Ca-Na-SO₄-Cl chemistry is located near the northeast quadrant of Bear Lake, also along the East Bear Lake Fault. Extrabasinal solute sources may be important to Bear Lake Valley, but additional data are needed to adequately address that issue. Faulting exerts a strong control on spring locations, including Swan Creek Spring, Big Spring, and Mud Lake Hot Spring. Faults in Bear Lake Valley are important conduits for groundwater flow.

2. The groundwater north of Garden City is derived from modern recharge with shallow flow paths. The solute chemistry of Swan Creek Spring varies in response to its discharge. Solutes derived primarily from dolomite dissolution dominate base flow and solutes derived from increased limestone dissolution dominate peak discharge. Paris Spring and Blue Pond Spring have solute chemistries that reflect the dissolution of the dominant carbonate bedrock in their source areas. Discharge-dependent increases in Na and Cl at Swan Creek Spring may be anthropogenic, or related to a northward migration of Wasatch Formation solutes

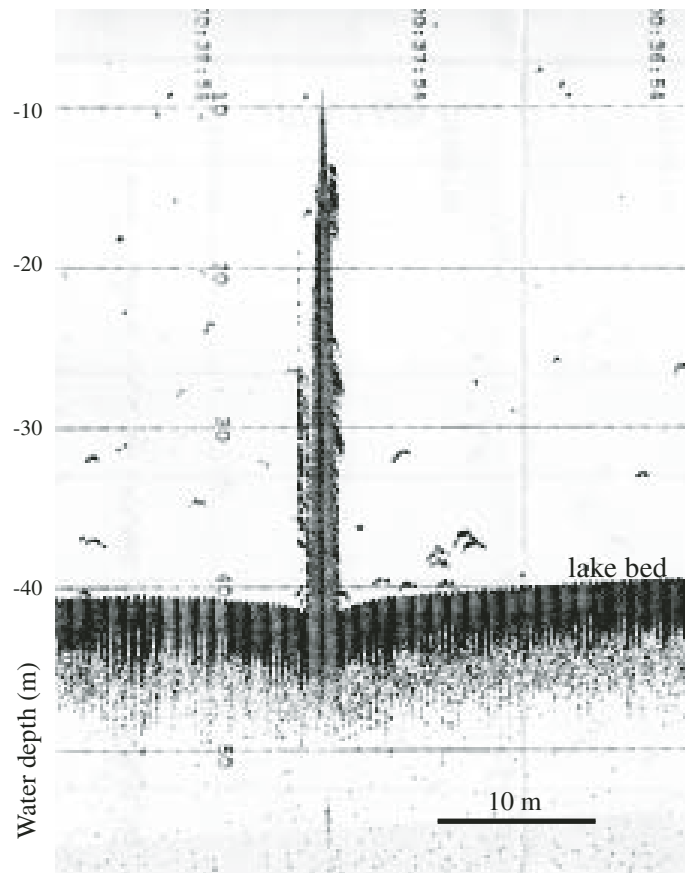


Figure 14. Sonar image of a sublacustrine methane seep initially thought to be a large spring. See Figure 2 for location.

along range-bounding faults. Spring discharge at Big Spring is fault controlled, and its solute composition is a mixture of both Paleozoic carbonate and Wasatch Formation derived ions.

3. The karsted Bear River Range aquifer north of Garden City is the primary recharge and discharge area for water and solutes entering Bear Lake. The solution features west of Bear Lake are important recharge areas, but recharge outside of the immediate watershed is also likely. Faulting exerts a strong control on the local and regional hydrology, serving in some cases as conduits and other cases as barriers for groundwater flow. A portion of the infiltration from the solution basins located in the southern part of the Bear River Range is probably discharged at Big Spring, but the remainder may be diverted to the north, where it discharges at Swan Creek Spring, or it may be diverted away from Bear Lake.

4. Groundwater in the Bear River Range is modern, but only a small portion of the groundwater in the Bear Lake Plateau is modern. Stable isotope ($\delta^{18}\text{O}$, $\delta^2\text{H}$) data indicate that the topography of the Bear River Range exerts a major control on the distribution of stable isotope values of groundwater in southern Bear Lake Valley.

5. The hydrologic balance of Bear Lake is apparently maintained by surface runoff and by shallow groundwater sourced within the Bear River Range. Groundwater leakage around the margins of the lake or through the lake floor is (was) probably not a major source of solutes into Bear Lake, although a unique K-rich water source is needed to generate the pre-diversion lake chemistry. A significant influx of solutes from the eastern and southern parts of Bear Lake Valley is incompatible with the solute and $^{87}\text{Sr}/^{86}\text{Sr}$ balance for the pre-diversion lake. If the inflow to Bear Lake has been overestimated and subsurface groundwater influx is a sizable component of the lake's hydrologic balance, then the northwest quadrant of the valley is the only source area with a compatible solute composition.

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