Late Neogene and Quaternary Lacustrine History of the Great Salt Lake-Bonneville Basin

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ABSTRACT

The Great Salt Lake-Bonneville basin has contained lakes for many millions of years and has been hydrographically closed for most of its history. Lakes in the lacustrine system have ranged from saline to fresh, and from shallow to deep. Tectonics, specifically crustal extension, which began roughly 20 million years ago as part of the formation of the Basin and Range Province, is the cause of lake-basin formation. Much of the rock record of lakes from Miocene time is faulted and has been eroded and/or buried. Pliocene and Quaternary lakes are better known. For much of the past ~5 Ma the basin has probably appeared similar to today, with a shallow saline terminal lake in a dry desert surrounded by mountains. Freshwater marshes and fluvial systems existed on the basin floor during part of the past ~5 Ma, probably were caused by the lack of inflow from the upper Bear River during the Neogene Period and most of the Pleistocene Epoch (that river was diverted into the basin during the Late Pleistocene), combined with a warm and dry climate. The largest deep-lake cycles were caused by changes to a cold and wet climate, which affected the water budget of the lake system and were correlated with periods of global glaciation.

Based on limited data, the total length of time deep lakes existed in the basin is thought to be less than 10% of the past ~773 ka. Lake Bonneville, the most-recent of the deep-lake cycles, was probably the deepest and largest manifestation of the lake system in the history of the basin. Named deep-lake cycles during the past ~773 ka, are Lava Creek (~620 ka), Pokes Point (~430 ka), Little Valley (~150 ka), Cutler Dam (~60 ka), and Bonneville (~30 -13 ka).

Of the Quaternary deep-lake cycles, only Lake Bonneville is represented by lacustrine landforms, outcrops, and cores of offshore deposits; no landforms from older deep-lake cycles exist (some may be buried under Lake Bonneville deposits but are not visible at the surface), and pre-Bonneville lakes are represented by sediments in limited outcrops and drill holes (including a set of cores taken by A.J. Eardley in the mid 20th century). During the past ~773 ka, deep-lake cycles were correlated with changes in the total volume of global glacial ice; the available evidence indicates that prior to ~773 ka deep-lake cycles were rare.

INTRODUCTION

This paper discusses lakes of Pliocene through Quaternary age (Figure 1) that have occupied the Great Salt Lake-Bonneville basin (GSL-BB). The GSL-BB is located in the eastern Basin and Range Province and is part of the Great Basin (Figure 2). All lakes in the GSL-BB during its long history, which includes the past 15 or 20 million years (Ma, mega annum; Figures 1 and 2), should be thought of as parts of a single lacustrine system — this concept is extrapolated from that of Atwood and others (2016), who applied it to Lake Bonneville (LB) and post-LB Great Salt Lake (GSL). Lake size varied over time in response to tectonic and climatic changes; sometimes the lake was shallow and saline to hypersaline, and uncommonly it grew in depth, volume, and surface area to become brackish to fresh.

An important observation emphasized in this paper is that during the Pliocene and Quaternary Epochs the GSL-BB lacustrine system spent more time as a shallow lake than as a deep lake; deep-lake versions of the system have been relatively short lived and uncommon. A more quantitative approach to this observation is discussed below.

It is not possible to give precise definitions of "deep lake," and "shallow lake," but for this paper, "deep" lakes are regarded as being much bigger than modern GSL. In this general sense, "deep" lakes might range from a lake roughly the size of the Cutler Dam (CD) lake (see below for discussions of named lakes in the GSL-BB), roughly 60 m higher than the average elevation of modern GSL (1280 m), to the size of LB, almost 350 m higher than modern GSL in the middle of the basin. "Shallow" lakes would look similar to modern GSL, with average maximum depth near 10 m, but might be shallower than that or several tens of meters higher. With lake level constantly changing in the closed basin (on time scales longer than a few weeks), lake size is difficult to precisely define if shorelines are not available.

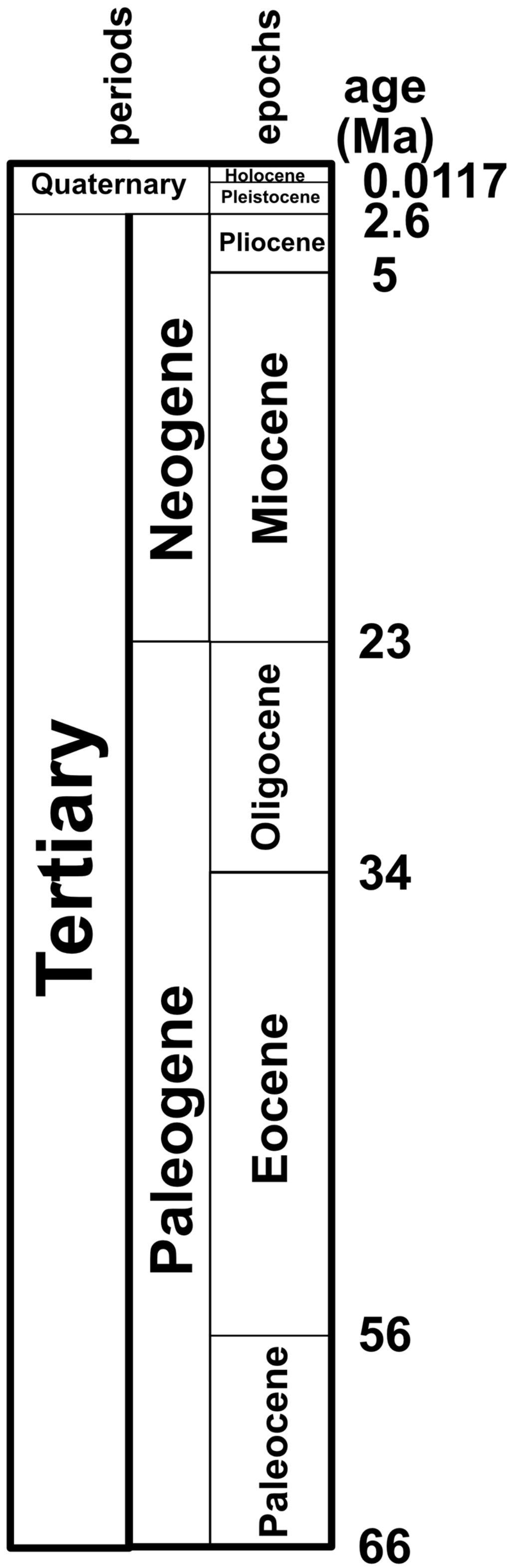

Figure 1. *Approximate ages (in Ma) for subdivisions of the Cenozoic Era (after Walker and others, 2018). In recent interpretations, the Tertiary Period (as it was called for many years) is now regarded as consisting of two geologic periods, the Paleogene and Neogene. The events discussed in this paper occurred during the Neogene and Quaternary Periods (the Miocene, Pliocene, Pleistocene, and Holocene Epochs).*

For this paper, the GSL-BB includes the subbasins that collectively comprise the Bonneville basin of Late Pleistocene to modern age. The subbasins are: (1) the Great Salt Lake (GSL) basin, (2) the Great Salt Lake Desert (GSLD) basin (separated from GSL by low divides), and (3) the Sevier basin (Figure 2). Major streams entering the system are the Sevier and Beaver Rivers in the Sevier basin, and the Provo/Jordan, Weber, and Bear Rivers in the GSL basin (Figure 2). All these rivers head in the high mountains and plateaus along the eastern margin of the basin. No major rivers flow into the GSLD basin, although a few rivers that are ephemeral today, were probably perennial during deep-lake episodes (streams such as Thousand Springs Creek, Grouse Creek, and Deep Creek [the Deep Creek that heads in eastern Nevada] built impressive deltas into LB). An upward component of groundwater flow (Stephens, 1974; Fitzmayer and others, 2004), and the observation that the mud of the mudflats is moist everywhere (except maybe for a few centimeters at the surface where the wind has dried it), indicates that the modern GSLD is a gigantic groundwater-discharge, or evapotranspiration area (in springs flow is concentrated).

Within the subbasins are smaller closed basins, such as Puddle Valley and Tule Valley in the GSLD basin, and Cedar Valley and Rush Valley in the GSL basin. All these hydrographically closed basins and subbasins exist because of Neogene and Quaternary faulting. The Wasatch fault bounds the eastern margin of the GSL-BB and the Great Basin (and Basin and Range Province), and has the greatest total offset of any fault system in the GSL-BB. The Wasatch fault accounts for the major mountain front of the Wasatch Range. The maximum thicknesses of Neogene and Quaternary sediment in the GSL-BB vary from place to place, and the sediments may be ~4 km thick, or more, in some places (Hintze and Kowallis, 2021). Details of the faulting history are beyond the scope of this paper, but faulting is an important long-term control on the lacustrine history.

This paper summarizes what is currently known about the lacustrine history of the GSL-BB for the past ~5 Ma. As is typical of geologic information, more is known about relatively recent events than about older events. The shapes and sizes of the older lacustrine basins within the GSL-BB are poorly known because of continued tectonic deformation.

MIOCENE TECTONICS AND DEPOSITION

Extension associated with the Neogene and Quaternary tectonics of the Basin and Range Province, including the GSL-BB in the eastern part of the Prov-

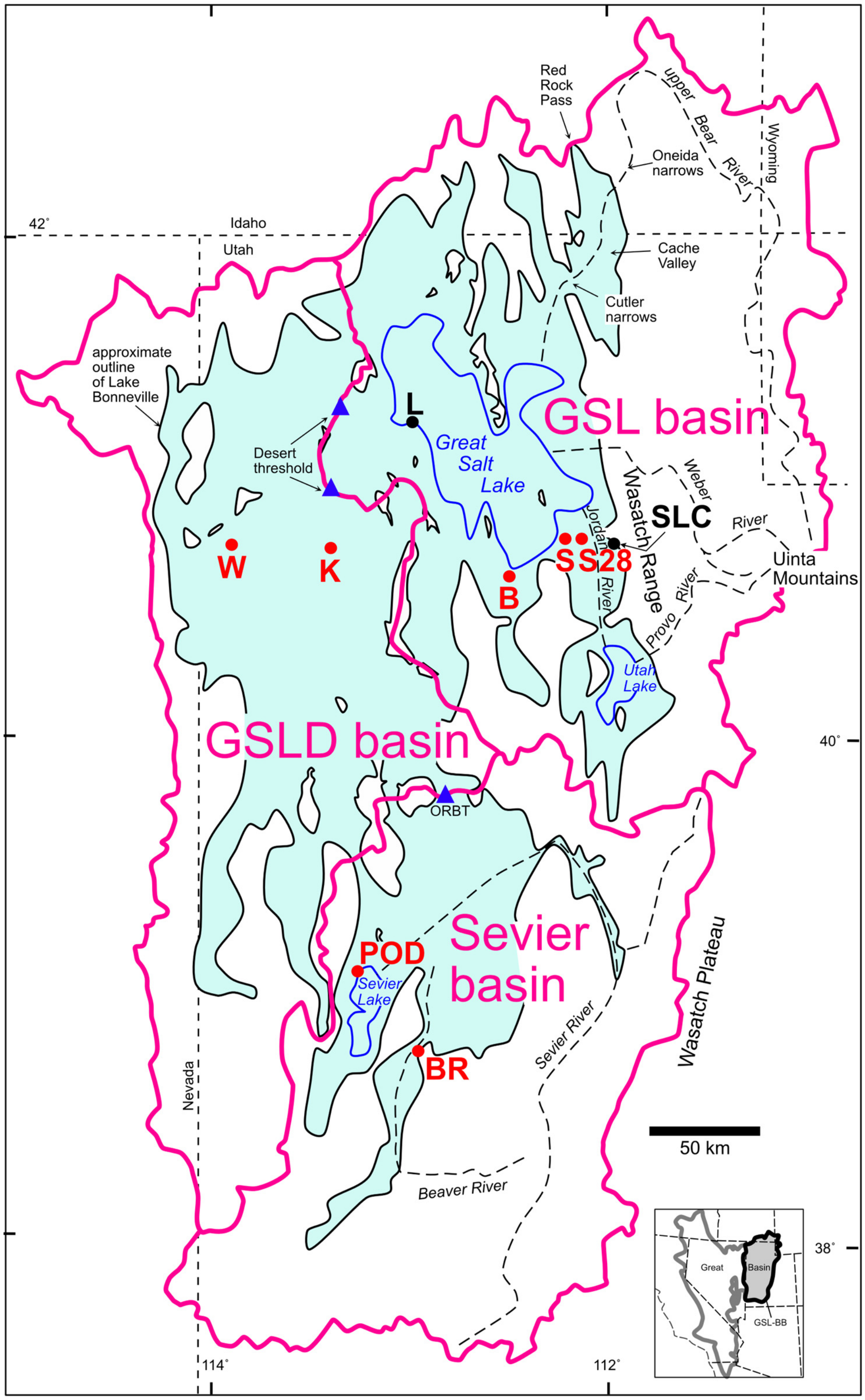

Figure 2. *Map showing approximate modern drainage divides for the subbasins within the GSL-BB. Pleistocene drainage divides were probably similar, but drainage divides for Neogene basins are not known. The GSLD basin has less than a meter of closure and it is separated from the GSL basin by two low thresholds, which are nearly imperceptible on the mudflats at identical elevations (1285 m); when first discussed by Eardley and others (1957) only one threshold, the southern one, was recognized and was called the "Desert threshold" (in some cases it is now called the "Eardley threshold"). The approximate outline of LB (the Bonneville shoreline) is shown for reference, as are major rivers that entered the basins from the east side. Modern lakes are labeled. Approximate locations of the Eardley cores and the Sevier-basin cores are shown with red dots (S28 = S28; S = Saltair; B = Burmester; K = Knolls; W = Wendover; POD = Pit of Death; BR = Black Rock). The low point on the divide between the Sevier basin and the GSLD, is the Old River Bed threshold (ORBT); flow from the Sevier basin entered the GSLD basin during the Late Pleistocene. Major rivers are shown schematically with dashed lines. L = Lakeside; SLC = Salt Lake City.*

ince, began roughly 20 Ma (Hintze and Kowallis, 2021). By at least 15 Ma, lake basins had begun to form in the eastern Basin and Range Province (Patton and Lent, 1980; Taylor and Bright, 1987; Oaks and others, 1999; Bortz, 2002; Janecke and others, 2003; Long and others, 2006; McClellan and Smith, 2020). Despite ongoing tectonism and many details of the topography that have changed between late Neogene time and the present, the general configurations of mountains and basins is probably similar now to what it was 5 Ma ago (Hintze and Kowallis, 2021). Some significant regional-scale changes have occurred in the SLC-BB during the time period in question, such as river diversions that have changed the water budgets of lakes (discussed below).

The Basin and Range Province is still tectonically extending today (WGUEP, 2016; Utah Geological Survey, 2023). Thick accumulations of lacustrine and associated deposits of Miocene age are exposed in such areas as Cache Valley, Utah and Idaho (Oaks and others, 1999; Janecke and others, 2003; McClellan and Smith, 2020), and Goose Creek, ID and NV (Perkins and others, 1995), and in many other places within or near the modern GSL-BB. Janecke and others (2003) present good evidence that Neogene lake basins developed in the area now called northeastern Utah and southeastern Idaho, many of which were associated with the evolving Bannock detachment fault system. It is likely that multiple individual basins were integrated into one large GSL-BB by Pleistocene times in response to continuing tectonism during the several-million-year period, but the details of the lacustrine history are still being discovered. Although there is no question that Miocene lakes existed in the GSL-BB, the outlines of individual basins and the shorelines of those old lakes are not preserved or are covered.

The GSL-BB is large (Figure 2), but it's not an ocean basin — because of the huge spatial variability in geology, biology, topography, etc., within the basin, a core taken from one point, or one outcrop, are unlikely to contain sediments that look similar to those in cores or outcrops several kilometers away. One core or outcrop, although it may contain valuable information, is not likely to record the geologic history of the entire basin. To construct a complete geologic history of the basin, information from multiple sources throughout the lake basin needs to be integrated, a process that takes a long time and efforts by multiple generations of scientists.

WATER BUDGET

The water budget of lakes in the GSL-BB is a fundamental consideration. Although precise measurements for many of the variables in water-budget equations for modern lakes are available, the values of important variables for older lakes can only be generally estimated. Water budget (or balance) can be expressed in many ways, but a simple equation shows water inflows equal to water outflows, plus-or-minus changes in storage of water in the lake (Hutchinson, 1957).

In the case of a hydrographically closed lake, water does not exit the system except by evaporation (there is no river or groundwater outflow). GSL is a closed-basin (or terminal, or endorheic) lake, so it has no surface outflow, and groundwater outflow is assumed to be zero (Arnow and Stephens, 1990). The relationship between volume and surface area (and elevation) in the modern GSL-BB is nearly linear (Wambeam, 2001). For most of its history the GSL-BB has been hydrographically closed and short-term changes in lake level have been correlated with changes in climate.

TECTONICS AND PALEOCLIMATE IN THE BONNEVILLE BASIN

The rate of tectonic deformation and sediment infilling compared to the water balance should be considered in tectonic basins (Bohacs and others, 2000). If climate in a basin favors a positive water balance, where inflows exceed outflows, a basin might appear to be open, but if tectonic subsidence of the basin floor is relatively rapid and the rate of sediment infill is low the basin might remain hydrographically closed even if inflows exceed outflows. The GSL-BB would be classified as "underfilled" by Bohacs and others (2000, their Figure 7; Bernau, 2022). In an underfilled basin plenty of space is available for water and sediment to accumulate, and that large volume of unfilled space keeps the basin from overflowing. In hydrographically closed basins, the water that remains in the basin after most of it has evaporated becomes increasingly salty over time (Hardie and Eugster, 1970).

Over its many-million-year history, the GSL-BB has remained underfilled with respect to sediment, and hydrographically closed most of the time. The rate of tectonic deformation in the GSL-BB is great enough that only one period is known where the basin was hydrographically open while remaining sedimentologically closed. This occurred when Late Pleistocene LB was overflowing at Red Rock Pass into the Snake River drainage basin as the Provo shoreline formed (Gilbert, 1890). During that period (possibly about 1000 to 3000 years in duration) climate was cooler and wetter than today and the lake was deep.

Neogene climate of the GSL-BB was probably similar to that of today, although the mean annual precipitation may have been generally lower and temperature somewhat higher (Moutoux, 1995; Moutoux and Davis, 1995, their Figures 3 and 4; Davis and Moutoux, 1998; Davis, 2002). These paleoclimate interpretations were based on pollen from samples of cuttings from drill holes in GSL (Table 1); the dating was not precise, but the pollen allowed for interpretations of generalized climatic conditions during the Pliocene and Early Pleistocene time.

Quaternary climate in the GSL-BB has been widely variable (Rhode, 2016). When deep-lake cycles occurred, climate was relatively cool and wet and during times when the lake system was shallow, climate was relatively warm and dry (Davis and Moutoux, 1988; Rhode, 2016).

FRESH- TO BRACKISH-WATER MARSHES ON THE BASIN FLOOR

Kowalewska and Cohen (1998), in an analysis of ostracodes (small crustaceans, typically about 1 mm in size) from cuttings taken from the same GSL drill holes that yielded the pollen samples mentioned above, found evidence of freshwater wetlands (marshes) and fluvial environments at various locations on the floor of the basin at different poorly dated times during the past 5 Ma. During the Holocene, the water of GSL has been hypersaline and has not supported ostracodes (Thompson and others, 2016), but at the locations of the drill holes studied by Kowalewska and Cohen (1998), freshwater conditions existed at times, and at other times the same places were occupied by shallow lakes, some of which were saline. Just the presence of freshwater ostracodes on the floor of the GSL-BB, which are not part of deep-lake faunas (Delorme, 1969; Forester, 1987), indicates hydrologic conditions much different than those of today.

Kowalewska and Cohen (1998) compared their ostracode results with pollen results described by Moutoux and Davis (1995), and they were not able to find meaningful correlations between the ostracode interpretations and pollen interpretations of the paleoclimate in the GSL-BB. One possibility to help explain why marshes and/or freshwater fluvial systems might appear low in the basin if it was hydrographically closed, is that, because of local tectonic activity, the basin floor was probably not smooth and uniform, but instead consisted of multiple shallow depressions separated by low ridges and hills. Fresh river water could flow into some depressions (and feed freshwater marshes and/or streams), but not into others, which might contain shallow saline lakes. The number and distribution of drill holes from which cut-

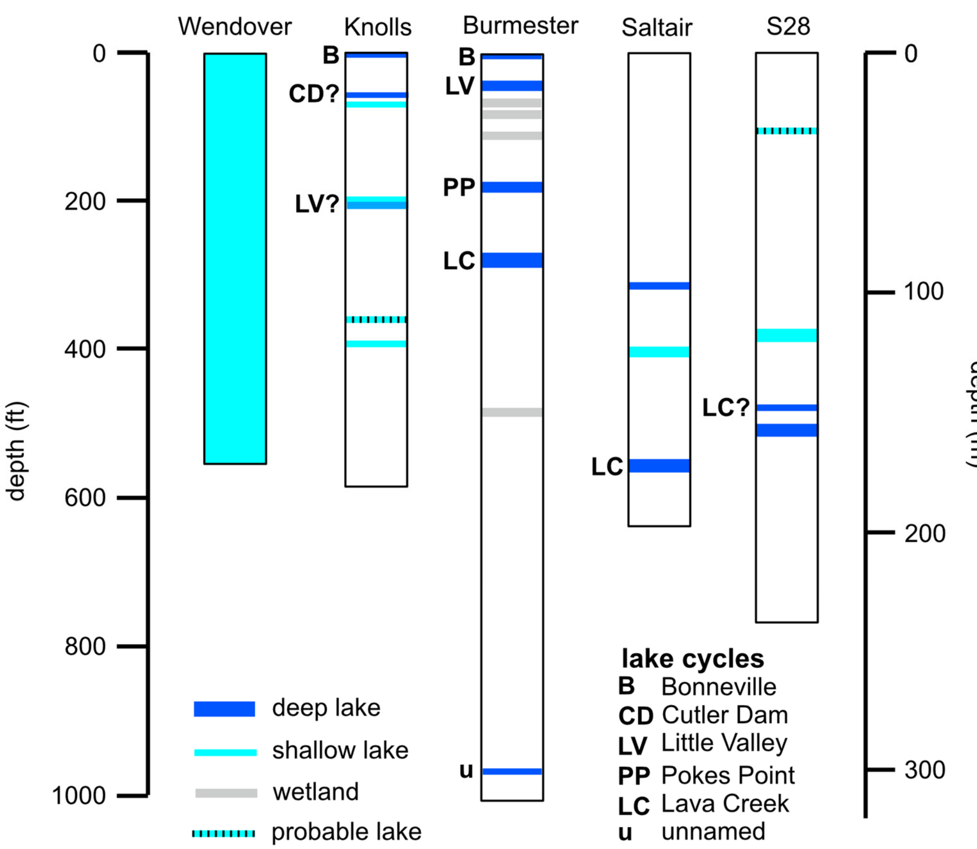

Figure 3. *The Eardley cores. This figure was assembled using data from published (Eardley and Gvosdetsky, 1960; Eardley and others, 1973; Williams, 1994; Oviatt and others, 1999) and unpublished sources (Shuey, 1971; Thompson and Oviatt, 1995, notes from core examinations; J. Bright, D.S. Kaufman, and R.M. Forester-- late '90s data on ostracode faunas and amino acid results for samples collected by Thompson and Oviatt).*

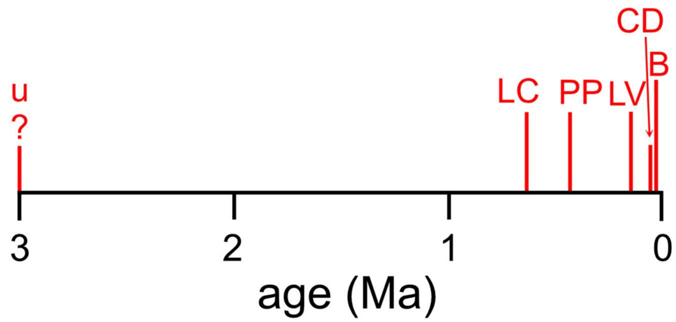

Figure 4. *Known lakes in the GSL-BB larger than modern GSL during the past 3 Ma. Chronologic data are from the Burmester core for the Bonneville, Little Valley, Pokes Point, Lava Creek, and unnamed lake cycles (Oviatt and others, 1999; unpublished information); the age of the CD lake cycle is from Kaufman and others (2001). The X axis of the graph marks the approximate elevation of modern GSL (~1280 m), and the vertical scale, which represents the relative maximum elevations of lakes, is not shown on the figure because insufficient information is available for most lake cycles. Approximations of the upper elevation limits of the CD and LV lake cycles are based on outcrops of lacustrine sediment. The upper elevation limits of the PP and LC lake cycles are interpreted as being similar to that of the LV lake cycle. The elevation of the unnamed lake cycle at about 3 Ma is unknown, but based on the ostracode fauna in sediments of that age from the Burmester core, the lake probably did not rise higher than the CD lake cycle. B = Bonneville, CD = Cutler Dam, PP = Pokes Point, LC = Lava Creek, u = unnamed lake cycle.*

tings were obtained is not sufficient to determine if this explanation is viable; also the available geochronological control is not good enough to make reliable correlations between cores. In a different data set (core GSL00-4; Balch and others, 2005, their Figure 6), the youngest ostracode fauna from a freshwater marsh on the basin floor is on the order of ~45 ka (kilo [1000] annum; presumably prior to the diversion of the upper Bear River and Cache Valley tributaries into GSL (see discussion below).

An important contributing cause of the appearance of marshes and/or freshwater fluvial systems on the basin floor involves the diversion into the GSL-BB of the upper Bear River plus the rivers that drain Cache Valley. These rivers contribute water and dissolved solids to modern GSL. The precise ages of incisions of canyons along the path of the Bear River have not been totally resolved, but it's likely that the incisions occurred during the Late Pleistocene.

According to Pederson and others (2016, their Table 2.1) Oneida Narrows (Figure 2; on the topographic divide of Cache Valley) was fully incised, allowing the upper Bear River to enter Cache Valley, based on optically stimulated luminescence ages, after 55.0 ± 5.6 ka and before 48.9 ± 6.9 ka (a round number near the middle of that overlapping range is 50 ka). Prior to the incision of Oneida Narrows, the upper Bear River had a complicated history involving flow into the Portneuf River (a tributary of the Snake River) and ponding upstream from Oneida Narrows to form Lake Thatcher (Gilbert, 1890; Bright, 1963; Pederson and others, 2016).

Another canyon through which the upper Bear River now flows into the GSL-BB is the Cutler narrows ("gate of Bear River," Gilbert, 1890, his Plate XXX), where the upper Bear River plus its Cache Valley tributaries exit Cache Valley. The exact timing of the incision of Cutler narrows, and the mechanism of the incision, has not been determined, but all the incision (it's possible the incision occurred in stages?) probably was not completed until sometime after the CD lake cycle (that is, after ~60 ka; Oviatt and others, 1987; Kaufman and others, 2001; Oaks and others, 2024).

The incision of Cutler narrows was traditionally interpreted to be the result of superposition probably combined with antecedence (Williams, 1958; Maw, 1968). The word "anteposition" was coined by Hunt (1982) to describe situations where incision began with superposition and continued because of tectonic uplift across the path of the river. Williams (1958), Maw (1968), and Hunt (1982) did not give specific ages or directly discuss which river was superposed to ultimately create Cutler narrows. Movement on the Wasatch and West Cache Valley fault zones would easily account for tectonic uplift of the Junction Hills bedrock block across a superposed river. If the anteposition interpretation were correct, however, the river that was superimposed across the Cutler divide could not have been the Bear River if the upper Bear did not incise Oneida narrows and enter Cache Valley until about 50 ka. More work is needed on the geologic history of the Cutler narrows.

Oaks and others (2018; 2024) suggested the presence of lakes in Cache Valley separate from lakes in the GSL-BB, but the precise ages and characteristics of those Cache Valley lakes have not been determined. This study adopts the relative age of incision of Cutler narrows as younger than the CD lake cycle and older than the LB lake cycle (possibly close to 30 ka, but this has not been scientifically tested).

The upper Bear River, plus the total discharge of rivers that enter Cache Valley from the nearby mountains, plus discharge from the Malad River, accounts for about a third of the modern annual inflow to GSL (Oviatt and others, 1987; Arnow and Stephens, 1990). Without input from the upper Bear River plus the Cache-Valley rivers, the river inflow to the GSL-BB lake system would have been significantly reduced.

Table 1. *Drill holes in the Great Salt Lake and Sevier basins that contain sediments of pre-LB age.*

Drill Hole ID	Collection Year	Latitude (°N)	Longitude (°W)	Elevation (m)	Depth of Hole (m)	Age at Bottom of Hole (Ma)	Core or Cuttings	Reference
GSL96-6	1996	41.0	112.4	1272	9	0.044	core	Thompson and Oviatt, unpublished, 1995-2022; Thompson and others, 2016
GSL96-4	1996	41.0	112.5	1272	5.5	0.04	core	Thompson and Oviatt, unpublished, 1995-2022
GSL00-4	2000	41.1	112.6	1271	120	0.280	core	Schnurrenberger and others, 2001; Balch and others, 2005
C	~1980	41.0	112.4	1272	5.5	0.035	core	Spencer and others, 1984; Thompson and others, 1990
AMOCO 1	?	41.5	112.8	?	?	?	cuttings	Moutoux, 1995
AMOCO 2	?	41.4	112.8	?	?	?	cuttings	Moutoux, 1995
AMOCO 3	?	41.4	112.8	?	?	?	cuttings	Moutoux, 1995
AMOCO 4	?	41.4	112.7	?	?	?	cuttings	Moutoux, 1995
AMOCO 5	?	41.4	112.7	?	?	?	cuttings	Moutoux, 1995
AMOCO 6	?	41.4	112.7	?	?	?	cuttings	Moutoux, 1995
AMOCO 7	?	41.4	112.6	?	?	?	cuttings	Moutoux, 1995
AMOCO 8	?	41.1	112.7	?	?	?	cuttings	Moutoux, 1995
AMOCO 9	?	40.9	112.3	?	?	?	cuttings	Moutoux, 1995
AMOCO 10	?	40.8	112.3	?	?	?	cuttings	Moutoux, 1995
South Rozel (J)	?	41.4	112.6	1272	?	~5	cuttings	Moutoux, 1995; Kowalewska and Cohen, 1998; Davis, 2002
Gunnison (P)	?	41.3	112.7	1270	?	~5	cuttings	Moutoux, 1995; Kowalewska and Cohen, 1998; Davis, 2002
Indian Cove (I)	?	41.3	112.6	1271	?	~5	cuttings	Moutoux, 1995; Kowalewska and Cohen, 1998; Davis, 2002
Bridge	?	41.2	112.5	?	?	?	cuttings	Moutoux, 1995; Davis, 2002
Carrington Island (H)	?	41.0	112.5	2171	?	~5	cuttings	Moutoux, 1995; Kowalewska and Cohen, 1998; Davis, 2002
Sandbar (N)	?	40.7	112.4	?	?	~2.3	cuttings	Kowalewska and Cohen, 1998
S28	1960	40.9	112.2	1286	224	~0.9	core	Shuey, 1971; Eardley and Gvosdetsky, 1960; Williams, 1994; Thompson and Oviatt, unpublished, 1995
Saltair	1956	40.8	112.1	1282	198	~0.8	core	Shuey, 1971; Eardley and Gvosdetsky, 1960; Williams, 1994; Thompson and Oviatt, unpublished, 1995
Burmester	1970	40.7	112.5	1285	307	~3.4	core	Shuey, 1971; Eardley and others, 1970; Williams, 1994; Oviatt and others, 1999; Thompson and Oviatt, unpublished, 1995
Knolls	1960	40.7	113.3	1289	152	~0.9	core	Shuey, 1971; Williams, 1994; Thompson and Oviatt, unpublished, 1995
Wendover	1960	40.7	113.9	1285	171	~1.7	core	Shuey, 1971; Williams, 1994; Thompson and Oviatt, unpublished, 1995; Bright and others, 2022
Clive	2019	40.7	113.1	1307	187	?	cuttings	Stantec, unpublished, 2019; Oviatt, unpublished, 2019
Black Rock	1993	38.7	112.9	1503	273	~3	core	Thompson and others, 1995
Pit of Death	1993	39.0	113.2	1383	140	~3.1*	core	Thompson and others, 1995

*This core contains an unconformity @~140 m, below which is a ~6 Ma tephra.

If the climate in the GSL-BB basin were dryer than today during the late Neogene and Pleistocene (except during deep-lake cycles), it is likely that climatically induced river inflow to the GSL would have been reduced at that time (following the logic of Bekker and others, 2014, who studied tree-ring reconstructions of late Holocene streamflow in the Weber River and the connections with climate). Climatically reduced inflow, combined with the lack of inflow from the Bear and Cache-Valley rivers, would have caused lakes in the GSL-BB to be smaller compared to Holocene GSL, and that reduced input would likely increase the probability of streams feeding marsh systems in isolated depressions on the basin floor.

The information reported by Balch and others (2005) suggests that the hydrologic budget of GSL about at 45 ka was different than it is today. The difference in budget could have been that the Bear and Cache Valley rivers were not entering GSL 45 ka, and/or that climate was dryer at that time, during marine oxygen isotope stage (MIS) 3. MIS 3 was an interglacial period.

It is interesting and seemingly paradoxical that a hypersaline condition for the lake system in the GSL-BB (such as modern GSL) probably requires the inflow volume to be relatively high compared to that required for freshwater marshes to appear on the basin floor. It's clear that a decrease in water inflow to the lake causes lake level to decline; if inflow were to decrease sufficiently a hypersaline lake would cease to exist. In 2023, the upper Bear River and Cache Valley rivers are contributing water to GSL, and the lake is dropping to alarmingly low levels, partly because of the very warm and dry climate we are now experiencing, but mostly because of water diversions by humans from the inflowing rivers before the water gets to GSL (Abbott and others, 2023). If the upper Bear River and Cache-Valley rivers were not presently entering GSL, what would be the condition of the lake in 2023?

PLIOCENE TO LATE PLEISTOCENE DEPOSITION

Sevier basin cores

The Sevier basin (Figure 2) has been part of the larger GSL-BB for at least the past ~3 Ma. Two sediment cores from the Sevier basin record sedimentation during the period from ~3 Ma to a few thousand years younger than the Brunhes/Matuyama paleomagnetic boundary (Thompson and others, 1995), currently dated at 773 ka (Channell and others, 2010). These two cores, the Black Rock and Pit of Death cores (Table 1; Figure 2), contain sediments of shallow lakes and muddy (playa) depositional systems. No deposits of deep lakes were encountered in those cores, an observation that is consistent with observations from the GSL basin farther north and reinforces the interpretation that lakes in the GSL-BB were low or did not exist during the period from ~3 Ma to 773 ka. The deep-lake cycle at about 3 ka in the GSL-BB probably did not get high enough to flood into the Sevier basin; the elevation of the topographic divide between the GSL basin and the Sevier basin (ORBT, Figure 2) was probably on the order of 1400 m.

Eardley cores

During the 1950s and 1960s, Armand J. Eardley, who was a professor of geology at the University of Utah, oversaw the drilling of four deep holes and the acquisition of sediment cores from those drill holes. The cores were called S28, Saltair, Burmester, Knolls, and Wendover (Figure 2; Table 2). Eardley and his colleague, Vasyl Gvosdetsky (University of Utah), published a description and interpretation of one of the cores (the Saltair core; they also commented on the S28 core; Eardley and Gvosdetsky, 1960). R.T. Shuey, a colleague of Eardley's at the University of Utah, obtained funding from the National Science Foundation (NSF) to study the paleomagnetism of the sediments in the Eardley cores and wrote an unpublished report for NSF (Shuey, 1971). In 1973, Eardley and a group of colleagues, published a description and interpretation of part of the Burmester core (Eardley and others, 1973). Lister (1975) described ostracodes from the Saltair and S28 cores. S.K. Williams, a Ph.D. student of B.P. Nash (also at the University of Utah and a coauthor on the Eardley and others, 1973, paper), studied the volcanic ashes from the cores and published important information about the Eardley cores (Williams, 1994).

In 1995, R.S. Thompson (USGS) and C.G. Oviatt (Kansas State University) examined the five Eardley cores looking for evidence of deep-lake cycles based on the presence of carbonate marl deposited in deep lakes and deep-lake ostracode faunas. In 1999 Oviatt and colleagues published a brief description and reinterpretation of the upper ~110 m of the Burmester core (younger than the Brunhes/Matuyama geomagnetic boundary; Oviatt and others, 1999). As part of that work, J. Bright and D.S. Kaufman (Northern Arizona University), and R.M. Forester (USGS), studied ostracode faunas and ostracode amino acid racemization in most of the Eardley cores, and some of that information was published in Oviatt and others (1999). More recently J. Bright and colleagues studied amino acid racemization in ostracodes from the Wendover

Table 2. *Information about the Eardley cores*

core ID	PLSS[1]	latitude[2]	longitude[2]	elevation (m)	depth (m)	sed. rate (m/Ma)[3]	approx. age at bottom (Ma)	year of drilling	recovery	references
S28	SW1/4, SE1/4, Sec. 28, T1N, R2W	40.79	112.07	1286	223	230	~0.9	1960	0% in some sec-tions, up to 40% in others	Shuey (1971); Williams (1994)
Saltair	SE1/4 Sec. 25, T1N, R3W	40.79	112.20	1282	198	260	~0.8	1956	50%	Eardley and others (1963); Shuey (1971); Williams (1994)
Burmester	SE1/4, Sec. 7, T2S, R5W	40.65	112.45	1286	306	3.4-2.6 Ma: 90 m/Ma; 2.6-0 Ma: 120 m/Ma	3.4	1970	90%	Shuey (1971); Eardley and others (1973); Williams (1994); Oviatt and others (1999)
Knolls	SW1/4, Sec. 15, T1S, R13W	40.72	113.30	1289	152	170	0.9	1960	30%	Shuey (1971); Williams (1994)
Wendover	SE1/4, Sec. 15, T1S, R18W	40.74	113.87	1285	171	130	1.7	1960	50% < ~120 m; 15% > ~120 m	Shuey (1971); Williams (1994)

[1]PLSS = Public Land Survey System
[2]datum for latitude/longitude coordinates is WGS84.
[3]data from Williams (1994); approximate sedimentation rates

core (Bright and others, 2022), the only core not studied for that purpose in the 1990s. Davis (2002) published pollen diagrams that had been constructed from data from the Wendover and Knolls cores in the 1960s, but which had not been previously published.

The Eardley cores are now completely dried out. They have been stored in cardboard boxes and sampled multiple times by different people for different purposes. Observations about the geologic history of the core sites, which would have been possible when the cores were fresh, are now difficult. The Eardley cores are now archived at the Utah Geological Survey Core Research Center.

The usefulness of the Eardley cores is limited because some of the core sections have crumbled. Although drilling technology has been vastly improved since the 1960s, the cost of drilling and the acquisition of even one new core that might build on what has been learned from the Eardley cores, would be huge. However, the scientific information (geologic, biologic, paleoclimatic, etc.) that could be obtained from a new core would be invaluable.

The following sections give summaries of published and unpublished information and interpretations concerning the Eardley cores (Figures 3 and 4; Tables 1 and 2). Eardley did not publish anything related to two of the cores (Knolls and Wendover). No independent studies of the sediments or changing depositional environments represented in the Knolls and Wendover cores have been published.

When Bob Thompson and I examined all the Eardley cores in 1995, we found that the core sections had not been split and the surviving sections of the cores were covered with dried mud from the drilling operations. In order to examine the sediments, we had to look at the ends or break apart dried core sections or scrape off the mud from the surfaces. We found this to be true for all the cores, including the Saltair and Burmester, so it was unclear to us how Eardley and his colleagues had observed any of the sediments in the cores.

S28 and Saltair cores

These Saltair and S28 cores were taken near each other (Figure 2; Tables 1 and 2). Although some important information about pre-LB lake cycles is preserved in these cores (Eardley and Gvosdetsky, 1960; recognizing that interpretations of global Quaternary history have changed considerably since the 1950s), the amount and quality of information about the lacustrine history of the GSL-BB the cores can provide is not great. Both the S28 and Saltair cores were drilled at locations dominated by the Jordan River and its precursors and were not suitable as complete records of sedimentation in GSL-BB lakes.

Deposits of the LB cycle are not present in either the S28 or Saltair cores, and it would now be difficult to determine whether LB sediments were not preserved at the coring sites or if LB sediments simply were not recovered during the drilling operations. Deposits of some older deep-lake cycles are present in the cores and deposits of some deep-lake cycles are missing. Although Lister (1975) defined some new ostracode species based on samples from the S28 and Saltair cores, and his descriptions of ostracodes are excellent and useful, he did not indicate the depths of the samples or say anything about the depositional environments of the samples he examined.

Burmester core

The Burmester core is the longest Eardley core at 306 m and covers the greatest amount of time (the age at the base of the core is ~3.4 Ma; Williams, 1994). In our examination of the core in 1995 we found many buried calcic soils, some with enough soil carbonate to whiten the core for many meters.

Eardley and others (1973; their Figure 1) showed 17 deep-lake cycles during Brunhes time based on their work on the Burmester core, whereas Oviatt and others (1999) found evidence in the Burmester core for only four deep-lake cycles during the same time period (an age of 750 ka for the Brunhes/Matuyama geomagnetic boundary was estimated by Eardley and others, 1973; in 2023 the age of that geomagnetic boundary is considered to be ~773 ka [Channell and others, 2010]). In the upper ~3 m of the Burmester core Eardley and others (1973; their Figure 1) interpreted the sediments as representative of shallow to dry lakes, overprinted by a soil, but Oviatt and others (1999; their Figure 1) found deposits of LB in that interval, including the Hansel Valley basaltic ash (Miller and others, 2008), which was erupted during the early transgressive phase of LB.

Recovery was good in the Burmester core (90%; Table 2) and that core has provided ages for middle and Late Pleistocene deep-lake cycles in the basin (Figure 4). The approximate drilling site of the Burmester core is low in the basin, but it is on land, not in the GSL, and no deposits of shallow lakes are preserved in the Burmester core.

Knolls core

The LB marl is present in the Knolls core. However, in a shallow pit about 4 km west of the approximate location of the Knolls core, only about 80 cm — approximately the lower half — of the LB marl (Gilbert's, 1890, white marl) are present, and the upper half has been deflated (Oviatt and others, 2020). It is unknown how much of the LB marl is present in the Knolls core and the section may not be complete. Sediments of pre-LB deep-lake cycles are present lower in the Knolls core, although it is not known if those deep-lake stratigraphic units are truncated or complete. Most of the core is dominated by sediments of shallow lakes (similar to the Wendover core, described below).

Wendover core

In the Wendover core the LB marl is completely absent, as are deposits of pre-LB deep lakes (unpublished observations by Thompson and Oviatt, 1995, and by Oviatt and D.L. Clark, 2019-2022; Bright and others, 2022; Clark and others, 2023; Bernau and others, 2024). Drilling recovery was not good (Table 2), but no non-lacustrine deposits have been observed; deposits of shallow lakes dominate the core. The Wendover core helps demonstrate the importance of deflation in the GSLD (Bernau, 2022; Bernau and others, 2023, this volume), but does not help with determining when deep-lake cycles occurred.

The sediments in the Wendover and Knolls cores reveal important information about the pre-LB history of the GSL-BB. In both cores, the most common sediment types are carbonate mud (grain sizes of clay, silt, some fine sand) and oolitic sands, where most of the oolitic grains are rod shaped. Also present are irregularly shaped carbonate lumps and gypsum grains (both primary and secondary precipitates). Some carbonate mud units (not the ones dominated by rod-shaped ooids) contain the ostracode *Limnocythere staplini*, but no other ostracode species are present.

L. staplini lives in brackish water with relatively low alkalinity. In this basin this means the lake was less than a few tens of meters deep — if it rose higher the water would have become diluted and other ostracode species would appear. The rod-shaped ooids probably indicate the presence of brine shrimp (Eardley, 1938); spherical ooids probably formed abiotically in the wave-agitation zone of a shallow saline lake (Eardley, 1938). These sediments indicate that in pre-LB times, lakes in the GSLD were shallow and varied in dissolved-solid content from being saline-enough to support brine shrimp at times (too saline for ostracodes), to being brackish and supporting ostracodes at other times (but no brine shrimp). Taking into account the poor recovery of the Wendover and Knolls cores (Table 2), the observations suggest that deposition in shallow lakes dominated in the GSLD for thousands or millions of years. Although dated shorelines of pre-LB lakes in the GSLD have not been found (and may not exist), fluctuating lakes with an average elevation of roughly 1300 ± 10 m would be suitable candidates for producing this kind of sedimentary record. A rise of GSL to about 1300 m today would cause widespread flooding and destruction of human infrastructure in the GSL part of the basin, but from a geologic perspective 1300 m is close to the average level of GSL. While “1300 m” is an arbitrarily chosen elevation, it’s within the possible range of elevations of closed-basin lakes that periodically flooded the GSLD during pre-B time.

This range of elevations is close to the maximum elevation of the latest-Pleistocene Gilbert-episode lake (~1297 m). The Gilbert-episode lake (about 12,000 years ago) formed after LB had evaporated, and was part of GSL. In the GSLD the Gilbert-episode lake was strongly influenced by fresh, cold water that flowed into the GSLD from the Sevier basin along the Old River Bed (Palacios-Fest and others, 2021; they referred to the Gilbert-episode lake as the "Old River Bed delta lake"), but in the GSL part of the system the same shallow lake was brackish (Thompson and others, 2016). Similar pre-LB lakes in the GSL-BB with elevations in the range of 1300 ± 10 m should be considered part of ancestral GSL, but it is unknown whether freshwater from the Sevier basin entered the GSLD in pre-LB times.

DISCUSSION

Figure 4 shows an estimate of the ages of known deep-lake cycles in the GSL-BB based primarily on data from the Eardley cores (Oviatt and others, 1999). Figure 5 shows correlations between deep-lake cycles in the GSL-BB and MISs (ages summarized by Lisiecki and Raymo, 2005). Deep lakes, other than the ones that have been documented so far, may have risen and fallen during additional even-numbered MISs during Brunhes time (even-numbered stages were glaciations, odd-numbered stages were interglacials), but further investigations are needed to decipher details. If samples of vein-fill calcite and aragonite from outcrops at Lakeside were deposited during deep-lake cycles, they may suggest deep-lake cycles during MIS 8 and MIS 10 (D. McGee, MIT, personal communication, 2019)(Figure 5).

It is possible to estimate the proportion of time that deep-lake cycles occupied the GSL-BB during the Brunhes geomagnetic Chron (730-0 ka). If each of the four largest deep-lake cycles lasted the same length of time as the Bonneville cycle, about 17 ka, the total proportion of time that deep-lake cycles occupied the GSL-BB during the past 773 ka was roughly 9%. Only one deep-lake cycle is poorly known from the period between 3 Ma and 773 ka (based on limited information from the Burmester core), and, based on its ostracode fauna, the lake

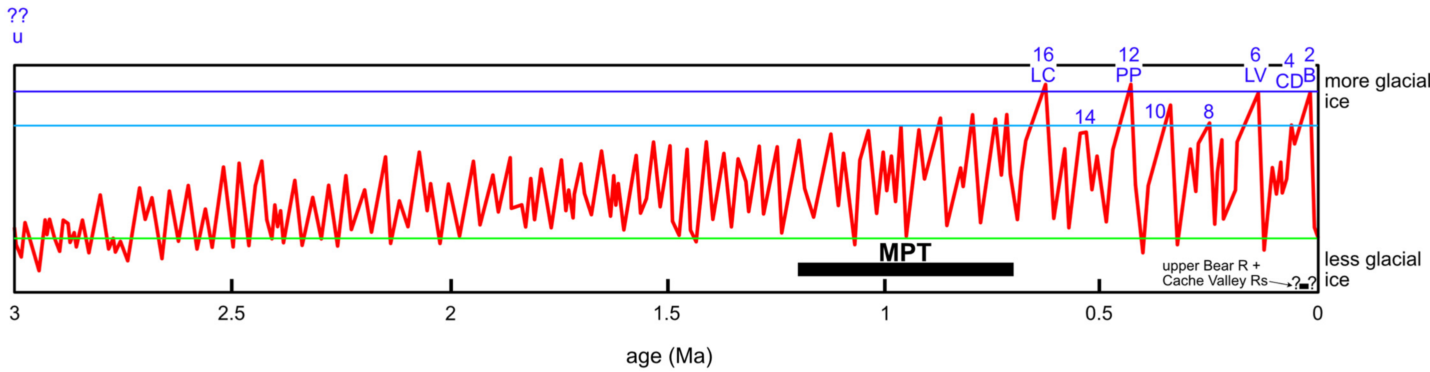

Figure 5. *Marine oxygen-isotope (MIS) record, which can be interpreted as representing the relative volume of global glacial ice (simplified from Lisiecki and Raymo, 2005), and known deep-lake cycles in the GSL-BB during the past 3 Ma. The red line is a stacked record of $\delta^{18}O$ in foraminifera fossils from 57 sites around the world where deep-sea cores have been taken (Lisiecki and Raymo, 2005; values of $\delta^{18}O$ in ocean water were relatively high at times when glacial ice attained large volumes on Earth's surface, and relatively low when ice sheets melted and the water flowed back to ocean basins; values of $\delta^{18}O$ are also correlated with water-temperature changes). Deep-lake cycles in the GSL-BB are shown in blue with their presumed correlative MIS stage numbers—B (Bonneville) ≈ MIS 2; CD (Cutler Dam) ≈ MIS 4; PP (Pokes Point) ≈ MIS 12; LC (Lava Creek) ≈ MIS 16 (Oviatt and others, 1999); the MIS stage number possibly correlative with the unnamed lake cycle ("u") about 3 Ma is unknown. Three other even-numbered stages are marked on the figure that are likely to have been correlative with deep lakes in the GSL-BB, but deposits of those hypothetical lakes have not been found. A possible age (very approximately 30 ka) of the diversion of the upper Bear River and Cache Valley rivers into the GSL-BB, is plotted. For reference, the green line is plotted at the level on the isotope curve approximately coincident with MIS 1 (the Holocene); the pale blue line is plotted at the level on the isotope curve approximately coincident with the CD lake cycle (MIS 4); the darker blue line is plotted at the level on the MIS curve approximately coincident with the LB cycle (MIS 2). The approximate duration of the middle Pleistocene transition (MPT; Clark and others, 2006) is shown.*

probably did not rise higher than the CD lake cycle; deep-lake cycles account for less than 1% of that period. Therefore, for over 90% of the past ~3 Ma lakes in the GSL-BB were shallow.

Of course, if further evidence is found for deep lakes other than the ones that have so-far been described for the past 3 Ma, the percentage of time during which deep lakes occupied the GSL-BB would be greater than 9 %. However, environmental conditions like what we see now (not including human influences) apparently were the rule rather than the exception for at least the past 3 Ma, and probably for a longer period (based on the MIS record of Lisiecki and Raymo [2005, their Figure 4], which extends back beyond 5 Ma). The domination of shallow lakes in the GSL-BB is not surprising considering that the upper Bear River and the Cache Valley rivers did not enter the GSL-BB until just a few tens of thousands of years ago (Figure 5).

As shown in Figure 5, very deep lakes in the GSL-BB were uncommon prior to the Middle Pleistocene transition (MPT; between about 1.2 Ma and 700 ka), which marked a change in the magnitude and frequency of Pleistocene glaciations (Clark and others, 2006; Clark, 2012). After the MPT, global climate varied with high-amplitude 100-ka cyclicity (as seen in MIS curves; Figure 5), and prior to the MPT, global climate varied with lower amplitude 41-ka cyclicity. After the MPT large Northern Hemisphere ice sheets began to attain great elevations and had larger volumes than earlier ice sheets (Clark, 2012). Very thick Northern Hemisphere ice sheets probably affected global atmospheric circulation patterns and may have been important in the growth of large lakes in the Great Basin (Antevs, 1948), although it's likely that the influence of ice sheets on global circulation was more complicated than that portrayed by Antevs (Oster and others, 2015).

The CD lake cycle and the post-LB Gilbert-episode lake are not represented by deposits in the Burmester core (or in any of the Eardley cores, except possibly in the Knolls core — Figure 3), but independently those lakes are known to have covered the Burmester core site and all other Eardley-core sites. Perhaps those lake cycles were quick (fast up, fast down), and little sediment was available at the core sites; or perhaps sediment from those lakes was present immediately after the lake cycles but was not preserved. If sediments of those lake cycles do not exist in the Eardley cores, maybe other major lake cycles occurred in the basin but have not yet been detected. Balch and others (2005) in a study of core GSL00-4 from GSL, including lake sediments that ranged in age from the present to as old as ~280 ka (Table 1), did not report evidence of those short-lived lake cycles. However, the spacing of the samples they examined averaged about 1 m (this represents an average of about 2400 years in that core). Even if sedimentation was continuous in some depressions on the floor of GSL (as at the site of GSL00-4), sampling at ~2400 years spacing may not have been close enough to intercept lake cycles that may have lasted only centuries or less. Clearly much remains to be learned about pre-LB lakes in the GSL-BB.

Based on what we know now, it is safe to say that the long-term appearance of the GSL-BB has been close to what we see today, with a shallow saline lake on the floor of the basin. LB was an anomaly, as were other deep-lake cycles in the basin. Our historic view of GSL (the past 170+ years) is occurring during a drop in the ocean of geologic time. From a perspective grounded in geologic time, GSL should be viewed as typical rather than as a "remnant" of LB.

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