Lake area constraints on past hydroclimate in the western United States: Application to Pleistocene Lake Bonneville

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ABSTRACT

Lake shoreline remnants found in basins of the western United States reflect wetter conditions during Pleistocene glacial periods. The size distribution of paleolakes, such as Lake Bonneville, provide a first-order constraint on the competition between regional precipitation delivery and evaporative demand. In this contribution we downscale previous work using lake mass balance equations and Budyko framework constraints to determine past hydroclimate change for the Bonneville and Provo shoreline extents of Lake Bonneville during the last glacial cycle. For the Bonneville basin we derive new relationships between temperature depression and precipitation factor change relative to modern. These scaling relationships are combined with rebound-corrected estimates of lake area and volume and macrofossil-derived surface temperatures to make quantitative estimates of precipitation and water residence times for the lake. For the Bonneville shoreline (~1552 m) we calculate that, prior to spillover to the Snake River drainage, precipitation rates were ~1.37 times modern, with a water residence time of ~185 years. For the Provo shoreline (1444 m), during the period of steady-state spillover, we calculate that precipitation rates were at least 1.26 times modern, with a residence time of ~102 years. These calculations suggest minimal difference in the hydrologic regime between the Bonneville shoreline highstand and the Provo shoreline stillstand during the last glacial termination. These estimates of hydroclimate scaling relationships differ in sensitivity with previous hydrologic modeling for Lake Bonneville and are complementary to those recently derived from glacier mass balance modeling from the Wasatch Mountains.

INTRODUCTION AND METHODOLOGY

The size of pluvial lakes in terminally draining basins of the western United States indicate a substantially different landscape during Pleistocene glacial periods. Building on the seminal work of G.K. Gilbert (1890) and J.C. Russell (1885), shoreline mapping and compilations of late Pleistocene lake surface areas in the western United States (e.g., Hubbs and Miller, 1948; Mifflin and Wheat, 1979; Williams and Bedinger, 1984; Reheis, 1999a; Orme, 2008; Grayson, 2011) has led to estimates of hydroclimate based on steady-state mass balance assumptions related to observed lake areas (e.g., Mifflin and Wheat, 1979; Hostetler and Benson, 1990; Reheis, 1999b; Broecker, 2010; Matsubara and Howard, 2009; Munroe and Laabs, 2013; Reheis and others, 2014; Ibarra and others, 2014; 2018; Barth and others, 2016). Geologic observations of shorelines or outcrop extent from ancient lake systems thus serve as constraints for to provide quantitative paleoclimate reconstructions for comparison to climate model experiments (e.g., Ibarra et al., 2014; Oster et al., 2015; Barth et al., 2016; Lora et al., 2017).

One such relationship, derived from the steady-state water balance assumption, is for the area of the lake (A_L) to the area of the basin (A_B) (Hudson and Quade, 2013; Ibarra and others, 2018):

$$\frac{A_L}{A_B} = \frac{P-ET}{E_L-P+Pk} = \frac{Pk_{run}}{E_L-P+Pk_{run}} \quad (1),$$

where:

A_L and A_B = the areas of the lake and basin, respectively (km^2).

P = the basin average precipitation (mm/year).

ET = the tributary average evapotranspiration (mm/year).

E_L = the lake average evaporation (mm/yr).

k_{run} = the runoff ratio of the subaerial portion of the watershed (i.e., $P–ET = Pk_{run}$), including both groundwater and riverine runoff into the basin.

The relationship between precipitation and runoff is non-linear across climate states, and thus necessitates increased proportional runoff (higher k_{run}) with increased precipitation. To impose this non-linear relationship, we use the Budyko framework (Budyko, 1974; Fu, 1981; Broecker, 2010; Roderick and others, 2014; Greve and others, 2015), where k_{run} is determined, using the Fu (1981) formulation, as:

$$1 - k_{run} = \frac{ET}{P} = 1 + \frac{E_p}{P} - \left[1 + \left(\frac{E_p}{P}\right)^{\omega}\right]^{\frac{1}{\omega}} \quad (2),$$

where:

E_p = potential evapotranspiration, the liquid water equivalent of the net downward radiation at the Earth surface derived from energy fluxes (Roderick and others, 2014; Ibarra and others, 2018).

ω = a free parameter that integrates the hydroclimatic properties of a watershed or basin (Fu, 1981; Greve and others, 2015).

The global average ω value is ~2.6 (Roderick and others, 2014; Greve and others, 2015). Recent work has shown that more complex, spatially explicit hydrologic modeling of Pleistocene lakes follow precipitation-runoff relationships imposed by the Budyko framework (Matsubara and Howard, 2009; Barth and others, 2016), justifying this scaling approach for steady-state calculations.

In this contribution, we use the methods outlined by Ibarra and others (2018) to analyze the distribution of the precipitation and energy fields from the gridded North American Regional Reanalysis dataset (NARR). We downscale the results presented by Ibarra and others (2018) for the Bonneville basin, defined by the NARR grid cells (resolution of NARR is 32 km, n=201 grid cells) spanning the region (37.5 to 43 °N, 114.5 to 110.5 °W) and present results based on the median change in lake area. We have confirmed that the average NARR precipitation fields (1979 to present) agree well with the 30-year normal PRISM precipitation dataset (1981 to 2010) and long-term weather station data archived by the Western Regional Climate Center. Lake evaporation (E_L) is calculated using the Priestley-Taylor equation and an E_p versus temperature scaling of 1.6%/K (Ibarra and others, 2018). To calibrate the lake area scaling relationships specifically for Lake Bonneville we lower the Budyko ω value to 2.465, which is calibrated to the total modern lake and seasonal playa lake surface area (8,924 km^2, modern A_L/A_B = 6.65%), which includes the Great Salt Lake, Bear Lake, the Bonneville Salt Flats and other seasonal playa lakes as compiled in Ibarra and others (2018).

For the latest glacial Provo and Bonneville shorelines we use rebound corrected lake area and volume estimates determined by Adams and Bills (2016), and temperature depression estimates based on macrofossil assemblages determined by Harbert and Nixon (2018) (see Table 1). Use of the non-rebounded corrected values would result in underestimates in both precipitation and lake residence times. We note that the Bonneville shoreline lake area and volume estimate reported by Adams and Bills (2016) agree well with new high-precision

differential GPS measurements reported by Chen and Maloof (2017) (see also Currey and others, 1982; Bills and others, 2002; 2016). All contours and scaling relationships shown are the median change in precipitation and temperature (for details see Ibarra and others, 2018), intended to represent the basin average change in hydroclimate necessary to explain the observed lake areas. Medians of the calculated A_L/A_B distributions are used due to the non-normal distribution in modern precipitation and energy fluxes used for the scaling analysis. We report uncertainty ranges on our estimates, where the uncertainty reflects the full range of temperature estimates, and is asymmetric due to the Budyko scaling relationship and the non-normal distribution in modern precipitation and energy fluxes.

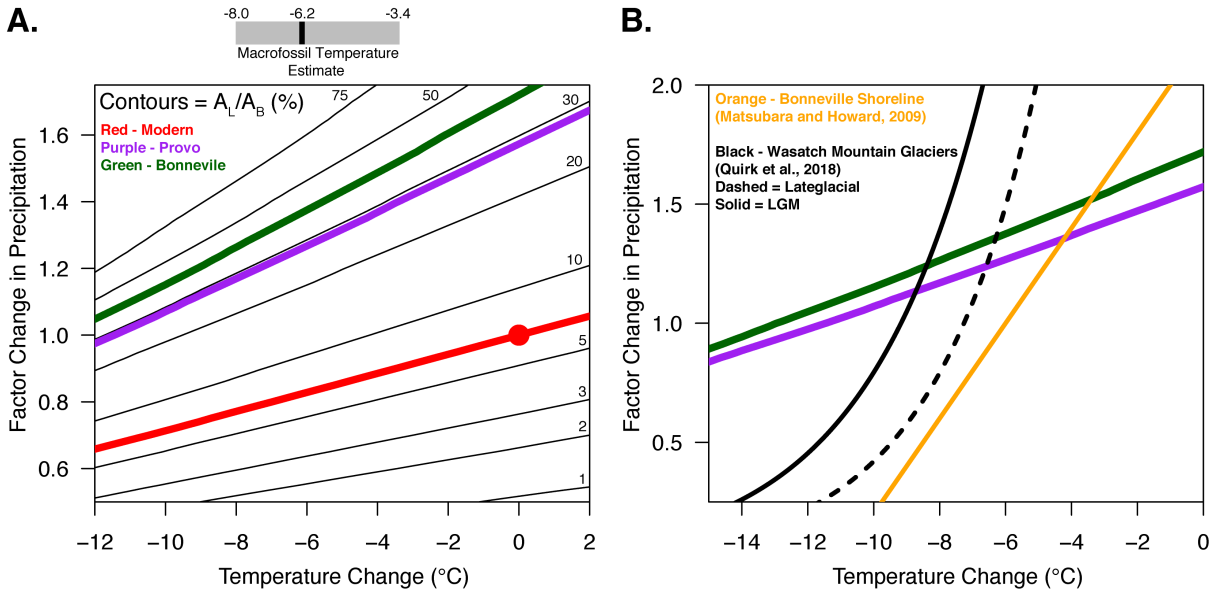

Figure 1. Hydroclimate change inferred from Lake Bonneville shoreline areas. (A) Contoured basin area-normalized lake area based on changes in precipitation and temperature over the Bonneville Basin (37.5 to 43 °N, 114.5 to 110.5 °W). Colored lines denote the modern (red), Provo shoreline (purple) and Bonneville shoreline (dark green). Macrofossil temperature depression estimates are from Harbert and Nixon (2018). Note that the black contours are not linear. **(B)** Comparison of lake area scaling relationships with previous temperature vs. precipitation estimates from hydrologic modeling of the Bonneville shoreline (Matsubara and Howard, 2009) and glacier mass balance modeling from Big Cottonwood Canyon in the Wasatch Mountains (UT) (Quirk and others, 2018), located on the eastern edge of Lake Bonneville. Note the axes difference between panels A and B.

RESULTS AND DISCUSSION

The results of our downscaled sensitivity analysis for Bonneville are displayed in Figure 1A and reported in Table 1. Solved for all combinations assuming a uniform temperature depression and factor change in precipitation over the basin, we plot the calculated precipitation factor change as a function of temperature depression needed to maintain a given lake area (expressed as A_L/A_B; black contours in Figure 1A). The colored contours on Figure 1A represent the Bonneville, Provo and modern surface area estimates.

Combining the temperature depression and lake area estimates (see above and Table 1) for the Bonneville shoreline surface area prior to spill over into the Snake River (~18.5 ka), we calculate precipitation rates that were ~1.37 (+0.15/−0.10) times modern with a water residence time of 185 (+13/−16) years. This estimate is comparable but higher than the Paleoclimate Modeling Intercomparison Project (PMIP) 2 and 3 Last Glacial Maximum (LGM; n = 15) ensemble average of 1.17 (±0.47, 1σ) for Lake Bonneville determined previously by Oster and others (2015), though we note that the best performing PMIP3 models (n = 5) suggest minimal change in precipitation (1.02 ±0.19) relative to modern. Our estimates of precipitation change for the LGM are higher than the central estimates from data to the west from Newark Valley (1.14 times modern, +0.15/-0.59) and Diamond Valley (0.97 times modern, +0.10/-0.51), derived from uranium-series systematics in soil opal (Maher and others, 2014; see their Table 2), but lower than estimates from lake balance modeling of Jakes Lake (~1.9 times modern produced by Barth and others (2016).

Similarly, for the Provo shoreline, we calculate a minimum precipitation factor change of 1.26 (+0.13/−0.09) times modern, and a water residence time of 102 (+7/−9) years. We note that minimal change in precipitation over the Bonneville region between the LGM and ~15.5 ka is simulated by the TraCE climate model simulation of the last 22 kyrs (Lora and others, 2017; see their Figure 4). Given ~2.5 kyr of stable shoreline development and continual overflow (Miller and others, 2013; Oviatt, 2015), this precipitation estimate is likely a minimum value considering the balance-filled nature of the Lake Bonneville system.

Table 1. *Modeling inputs and results for Lake Bonneville.*

Lake Stage	Lake Area [km^2][1]	Volume [km^3][1]	P/P_M vs. ΔT relationship[2]	Precipitation Factor Change	Water Residence Time (yr)
Bonneville	52,110 (A_L/A_B= 38.85%)	10,420	$P/P_M = 1.711 + 0.056\times\Delta T$	1.37 (+0.15/−0.10)	185 (+13/−16)
Provo	38,150 (A_L/A_B= 28.44%)	5,290	$P/P_M = 1.572 + 0.050\times\Delta T$	> 1.26 (+0.13/−0.09)	102 (+7/−9)

[1] Rebound corrected lake volume and areas determined by Adams and Bills (2016).
[2] P/P_M = precipitation factor change; ΔT = temperature change relative to modern. Linear relationships are approximate based on linear regression fit to contours in Figure 1.
Modeling assumptions: Budyko parameter, $\omega = 2.465$ (calibrated to total modern lake and playa areas; 8,924 km^2); Total basin area = 134,131 km^2; Residence time calculation assumes modern precipitation rate of 338 mm/yr (NARR median; comparable to PRISM value of 344 mm/yr); median temperature depression of −6.2 (−8.0 to −3.4 °C) based on filtering of macrofossil assemblage calculation from Harbert and Nixon (2018) for all deglacial and LGM locations (n=13) near Lake Bonneville; and temperature vs. potential evapotranspiration scaling of 1.6%/°C (Ibarra and others, 2018). Basin and modern lake/playa areas are from data tables in Ibarra and others (2018).

The residence of time of water in Lake Bonneville for the Bonneville and Provo shorelines place useful constraints on oxygen isotope measurements of sediments or shoreline tufa, though other geochemical systems, such as trace elements, are typically decoupled from water residence times. Our residence time calculations (Table 1) place minimum bounds on the timescale that oxygen isotope timeseries from cores or other high-resolution records could record rapid (sub-kyr) changes in climate (e.g., Dansgaard–Oeschger or Heinrich events) occurring within glacial-interglacial cycles (e.g., Benson and others, 1990; Oviatt and others, 1994; McGee and others, 2012; Ibarra and Chamberlain, 2015).

In Figure 1B we compare the Bonneville and Provo shoreline scaling relationships with two other estimates. Previously, Matsubara and Howard (2009) produced a spatially explicit model of the Great Basin and specifically analyzed the Bonneville region (see their Figures 4 and 5). Using their hydrologic modeling simulations Matsubara and Howard (2009) report an empirically derived multiplicative precipitation equation expressing precipitation factor change (P/P_M) as a linear relationship with temperature depression (ΔT), where $P/P_M = 2.2 + 0.2\times\Delta T$ (orange line in Figure 1B), a steeper slope than those derived using our approach (Table 1). In a similar fashion Quirk and others (2018) recently reported relationships for LGM and post-LGM glacier extents in the Wasatch range (shown as black lines in Figure 1B). Most significantly, our

relationships for the Bonneville and Provo shorelines intercept the time-equivalent post-LGM relationships from Quirk and others (2018) at a precipitation factor change of ~1.2 to 1.3 and a temperature depression of ~7 °C, comparable to the median value from the macrofossil estimates reported by Harbert and Nixon (2018) and in agreement with the PMIP climate model ensemble average temperature depression of 7.5°C (±2.6, 1σ; n = 15) for Lake Bonneville determined previously by Oster and others (2015).

CONCLUSIONS

In sum, we have downscaled the scaling analysis presented by Ibarra and others (2018) for the western United States for the Bonneville and Provo shoreline areas of Lake Bonneville during the last glacial period. These estimates are similar to previous hydrologic modeling and glacier mass balance modeling efforts. Future work will also constrain the scaling relationships associated with the pre-LGM Stansbury shoreline and post-Bonneville Gilbert episode. Furthermore, ongoing work to constrain changes in surface and lake temperature using pollen, macrofossil and carbonate clumped isotopic techniques (e.g., Mering, 2015; Harbert and Nixon, 2018) will provide independent constraints on the regional magnitude of temperature depressions for the last glacial period. Finally, we anticipate that future more advanced spatially explicit hydrologic modeling (e.g., Matsubara and Howard, 2009; Barth and others, 2016; Hatchett and others, 2018) and isotope mass balance approaches (e.g., Ibarra and others, 2014; Hudson and others, 2017) for the Bonneville basin will provide inter-comparison between different methods for determining past changes in hydroclimate from lake area extents.

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