Snowpack determines relative importance of climate factors driving summer lake warming

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Scientific Significance Statement

Lake thermal regimes are frequently decoupled from rising air temperature trends in their watersheds because interactions among climate factors and local features affect net heat gain or loss. Although snowpack has been empirically linked to temperature and warming rates in mountain lakes, the mechanisms governing this relationship remain unclear and untested, despite predicted declines in mountain snow across the world. We quantified how snowpack and ice cover regulate lake warming in summer primarily by controlling when and how long lakes receive solar radiation, and we demonstrate the relative role of snowmelt on the lake heat budget. We contrast lake thermal responses between extremely wet and dry years and illustrate the increasing importance of other climate factors in the absence of snow.

Abstract

Mountain lakes experience extreme interannual climate variation as well as rapidly warming air temperatures, making them ideal systems to understand lake-climate responses. Snowpack and water temperature are highly correlated in mountain lakes, but we lack a complete understanding of underlying mechanisms. Motivated by predicted declines in snowfall with future temperature increases, we investigated how surface heat fluxes and lake warming responded to variation in snowpack, ice-off, and summer weather patterns in a high elevation lake in the Sierra Nevada, California. Ice-off timing determined the phenology of lake exposure to solar radiation, and was the dominant mechanism linking snowpack to lake temperature. The relative importance of heat loss fluxes (longwave radiation, latent and sensible heat exchange) varied among wet and dry years. Declines in snowpack and ice cover in mountain systems will reduce variability in lake thermal responses and increase the responsiveness of lake warming to atmospheric forcing.

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Data Availability Statement: High frequency lake temperature and meteorology data used in the heat budget analysis (Data sets S1–S2), as well as hydrological and meteorological data used in multiple regression analyses (Data set S3) are archived on the Environmental Data Initiative (EDI; https://pasta.lternet.edu/package/data/eml/edi/445/1/c97b40cc0826203f2fc7386f80f40082).

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Water temperature regulates ecological processes in lakes by governing physical dynamics and biogeochemical reactions. Although many lakes are warming worldwide, thermal responses vary considerably (Schneider and Hook 2010; O’Reilly et al. 2015), in contrast to globally coherent increases in air temperature (IPCC 2013; O’Reilly et al. 2015). Climate change affects lake temperature by increasing heat gains or reducing heat losses, and there is evidence for both at a global scale (O’Reilly et al. 2015). However, the relative importance of these processes in driving lake responses to climate change remains poorly understood, which may have important implications for lake warming rates. For example, seasonally ice-covered lakes are warming faster than lakes without ice cover (O’Reilly et al. 2015), suggesting that lakes experiencing both increased heat gains (due to shorter ice cover and greater solar exposure) and reduced heat losses are the most sensitive to climate change.

Thermal responses of lakes to climate differ from observed trends in air temperature in part because: (1) air and water temperature are mechanistically linked via multiple heat flux mechanisms that depend on cloud cover, humidity, and wind (Edinger et al. 1968); and (2) lake and landscape features mediate climate responses of waterbodies (Toffolon et al. 2014; Winslow et al. 2015; Woolway et al. 2018). Lakes warm by absorbing shortwave and longwave radiation and cool primarily through surface exchanges with the atmosphere: emitted longwave radiation is a function of surface-water temperature; latent heat exchange (evaporation) is a function of air and water temperature, the air–water humidity gradient, and wind speed; and sensible heat exchange (conduction) is a function of the air–water temperature gradient and wind speed. Longwave radiation and latent heat fluxes are generally the largest heat losses from small to midsized lakes, with sensible heat exchange contributing less than 10% of total heat loss despite substantial diel, seasonal, and latitudinal variation in lake heat budgets (MacIntyre and Melack 2009; Woolway et al. 2018). Turbulent mixing caused by wind forcing or convection distributes heat vertically through the water column (MacIntyre et al. 2009; Monismith and MacIntyre 2009). The timing and strength of stratification, by determining mixed layer volume, can dictate the rate and magnitude of surface warming, especially in deep lakes (Piccolroaz et al. 2015). Last, stream or groundwater inflows can modify lake temperature, with the magnitude of effects dependent on the volume and temperature of incoming water and with the mixing extent (Rueda and MacIntyre 2009; Flaim et al. 2019). Predicting how climate change will alter lake thermal regimes requires understanding the interactions among climate factors that drive surface heat fluxes.

Mountain lakes are ideal systems to test hypotheses regarding lake-climate interactions because they experience extreme seasonal and interannual climate variability and have experienced rapid increases in air temperature (Sadro et al. 2018a). Snowpack and lake temperature are strongly correlated in mountain systems (Preston et al. 2016; Sadro et al. 2018a), likely due to tight coupling between snowpack and ice-cover in mountain lakes. Mountain lake “ice” is typically composed of alternating snow and ice lenses (Leppäranta 2015), and can be several meters thick in high snow years. Snowpack is expected to decrease in many mountain ranges as climate warms (Mote et al. 2018; Sun et al. 2019), with uncertain consequences for lake thermal regimes. Empirical relationships between catchment snow-water equivalent (SWE), snowmelt hydrology, ice-off date, and mean lake temperature suggest that snowpack influences lake thermal regimes through multiple heat flux mechanisms (Sadro et al. 2018a,b), including reduced inputs of solar radiation from prolonged ice cover, large inflows of cold snowmelt, and short water residence time during snowmelt. However, the relative importance of these mechanisms for lake thermal regimes is unclear, and previous analyses relied on statistical correlations rather than process-based approaches (Preston et al. 2016; Sadro et al. 2018a). Importantly, these analyses did not investigate interactions among climate factors that govern lake warming, nor how their relative importance shifts with reduced snow and ice cover.

Motivated by this context of rapid climate change and the sensitivity of mountain lake temperature to snowpack (Thompson et al. 2005; Sadro et al. 2018a), we investigated how heat fluxes, and thus lake warming during summer, responded to interannual variation in SWE, ice-off timing, and summer weather patterns in typical high elevation lake in the Sierra Nevada, California. We focused our analysis on the period between ice-off and maximum lake temperature (heating period) because lake temperature is most strongly linked to SWE during this time (Sadro et al. 2018a). We asked, (1) what are the relative roles of ice-cover and snow-melt for the lake heat budget during summer in high SWE (dry) years, and (2) what climate factors dictate lake heating in low SWE (dry) years, which may occur more frequently under future climate scenarios? Our results indicate the relative importance of mechanisms governing lake warming vary among wet and dry years in response to interactions among climate factors, and that declines in snowpack and ice cover in mountain systems will reduce the scale of variability in lake thermal responses and increase the responsiveness of lake temperature to atmospheric forcing.

Data and methods

Overview and site description

We estimated surface heat fluxes for Emerald Lake, a small lake in Sequoia-Kings Canyon National Park (SEKI), California (36°35′49″N, 118°40′29″W, 2800 m.a.s.l., maximum depth 10 m), for 4 yr (2014–2017), in order to investigate mechanisms linking snowpack and lake warming during summer. Most precipitation in the catchment falls as snow (86%; Sadro et al. 2018b), but SWE varies considerably among years (coefficient of variation (CV) = 50%; Fig. 1B, C). Ice cover develops in early to mid-December, whereas the date of ice-off can vary by over
100 d (April–August; Sadro et al. 2018b). 2014–2016 were dry years (< 1000 mm SWE), whereas 2017 was a wet year (SWE > 1700 mm; classifications relative to historic average; Sadro et al. 2018b). We used water temperature, hydrology, and meteorology data from Emerald Lake’s catchment (1992–2017) to assess empirical evidence for the mechanisms driving lake warming derived from the heat budget analysis, as well as to investigate potential interactions among climate drivers of lake warming. Data collection has been extensively described elsewhere (Sadro et al. 2018a,b), but we provide an overview in Supporting Information Text S1.

Surface heat budgets

We computed surface heat fluxes (W m$^{-2}$) from Emerald Lake in 2014–2017 at hourly time steps during the ice-free season, using vertically distributed water temperature measurements (instrument array in Fig. 1A; 20–22 depths) and meteorology data from an on-shore station (MET in Fig. 1A; Supporting Information Text S1; Data sets S1 and S2; Smits and Sadro 2019). Algorithms are as described in Imberger (1985) and MacIntyre et al. (2002, 2009); briefly, net surface heat flux $S = SW_{in} - SW_{out} + LW_{in} - LW_{out} - SE - LE$, where $SW_{in}$ is the downwelling shortwave radiation, $SW_{out}$ is the shortwave radiation reflected off the lake surface, $LW_{in}$ is downwelling longwave radiation, $LW_{out}$ is longwave radiation emitted by the lake surface, $SE$ is sensible heat exchange, and $LE$ is latent heat exchange.

The advective component of a lake’s heat budget captures the change in heat content due to lateral inflows and outflows. During the ice-free period, we estimated the advective component of the heat budget by computing the residual difference between the change in heat content each day calculated from temperatures in the lake, $Q_T$, and that calculated from the surface energy budget, $Q_M$, as in Augusto-Silva et al. (2018). We compared the heat content residuals (100* $[Q_T - Q_M]$/$Q_M$) among years, beginning at ice-off and ending with the transition from the descending limb of the hydrograph to summer base flow (snowmelt period; Sadro et al. 2018b).

To estimate energy flux “missed” by the lake due to prolonged ice cover in a wet year relative to a typical dry year, we summed net shortwave radiation ($SW_{in} - SW_{out}$) for 2017 between day of year (DOY) 144 (2014 ice-off DOY) and DOY 185 (2017 ice-off DOY) and assumed a constant loss flux of $-130$ W m$^{-2}$ ($LW_{net} + LE + SE$; mean value from DOY 144 to 185 in 2014). We chose 2014 rather than 2015 as a comparative dry year because summer weather conditions (cloudiness, rainfall) in 2014 were more typical relative to long-term averages.

Multiple regression analysis

We tested for statistical relationships between climate and hydrological variables and maximum lake temperature in the long-term data record (1992–2017; Data set S3; Smits and Sadro 2019) using multiple regression analysis. We chose annual maximum outlet temperature as a metric of lake warming after ice-off. Days in which air temperature is cooler than lake temperature correspond closely with periods of net heat loss (net loss days), providing a useful integrated measure of weather conditions associated with the lake’s cooling. For each year, we summed the number of net loss days during the heating period. We modeled annual maximum outlet temperature as a function of the following variables: SWE, ice-off day, number of net loss days during the heating period, total rainfall during the heating period, and mean air temperature during the heating period. We tested for additive and interactive effects of SWE and the other climate variables on lake temperature. We compared model fits using Akaike’s information criterion for small sample sizes ($AIC_c$; Burnham and 

![Fig. 1.](image-url)
Anderson 2002), using the “AICcmodavg” package in R (Mazerolle 2015; R Core Team 2015).

Results
Climate variation and lake heat fluxes
SWE varied by an order of magnitude from 2014 to 2017, resulting in variable ice-off timing, volume of snowmelt inflows, and divergent patterns of lake warming (Fig. 2; Table 1). Ice-off occurred almost 2 months later in 2017, a wet year, than in 2015. Discharge measured at the outlet varied appreciably, with significantly greater discharge in 2017 than in other years (Table 1, Fig. 3). In 2015, the snowmelt pulse was smaller than in other years, but multiple rain events occurred later in summer (after DOY 155; Fig. 3A). Mean theoretical residence time, or time required for cumulative discharge to equal lake volume, ranged from 8.5 to 68 d during the snowmelt period (Table 1).

Maximum lake heat content varied by up to 25% among years. Maximum epilimnetic temperature was lowest in 2017 (15.2°C) and highest in 2016 (20.8°C). \( \text{SW}_{\text{in}} \) was the largest source of heat for the lake; the major heat losses were \( \text{LW}_{\text{net}} \).
37–89% of total heat loss fluxes) and LE (14–34%; Fig. 2, right column). SE was a consistently small heat loss term from the lake (0–5%). The relative importance of different heat loss fluxes corresponded with SWE: in dry years, the proportion of heat lost via LWnet declined, whereas the relative importance of LE and SE increased (Table 1).

In all years, the lake was colder than predicted by the surface heat budget early in the ice-free season. Maximum daily heat content residuals ranged from $-7.3\%$ to $-16.5\%$ (Fig. 3A–D). Residuals (e.g., advective heat fluxes) were largest immediately after ice-off and diminished concurrently with the snowmelt pulse apparent in the lake outlet discharge (Fig. 3). While measurement or model error may contribute to residuals, seasonal decline in this term closely mirrored snowmelt discharge and approached zero later in summer, suggesting that advection was the dominant driver. The largest and most persistent advective fluxes occurred in 2017, when the magnitude and duration of snowmelt discharge were greatest.

Heating in the three dry years (2014–2016) did not respond linearly to SWE or ice-off date (Fig. 2). Despite becoming ice-free more than a month earlier than in the next driest year (2014), the lake warmed less and more slowly in 2015 than in 2014 or 2016 (Fig. 2A,B).

Prolonged ice cover in a wet year (2017) substantially reduced the total heat absorbed by the lake during summer. The potential energy “missed” by the lake due to an additional month of ice cover was a substantial fraction (70%; $1.71 \times 10^{13}$ J) of the net surface energy gained during the heating period. Conversely, in an extremely dry year with early ice-off (2015), longer exposure to solar radiation was offset by early season weather patterns that promoted heat loss.

### Table 1. Variation in catchment SWE, hydrology, outlet temperature, and heat budgets in Emerald Lake for the years 2014–2017.

|                      | 2014       | 2015       | 2016       | 2017       |
|----------------------|------------|------------|------------|------------|
| Maximum SWE          | mm         | Annual     | 530        | 241        | 1015       | 2319       |
| Ice-off date         | DOY        | Annual     | 144        | 105        | 155        | 184        |
| Total discharge      | m$^3$      | Annual     | $6.01 \times 10^5$ | $5.98 \times 10^5$ | $1.15 \times 10^6$ | $2.51 \times 10^6$* |
| Snowmelt period      | DOY        | After ice-off | 144–176    | 105–150    | 155–185    | 184–216    |
| Mean residence time  | Days       | Snowmelt period | 19.46 (15.7–27.5) | 68.18 (31.7–112.8) | 9.8 (5.8–14.1) | 8.9 (5.8–15.6) |
| Max epilimnion       | °C         | Annual     | 19.5       | 18.4       | 20.8       | 15.2       |
| Max temperature DOY  | DOY        | Annual     | 199        | 237        | 217        | 246        |
| Maximum heat content  | J          | Annual     | $1.19 \times 10^{13}$ | $1.18 \times 10^{13}$ | $1.21 \times 10^{13}$ | $9.83 \times 10^{12}$ |
| Max heat content DOY | DOY        | Annual     | 246        | 241        | 216        | 245        |
| Max daily heat residual | %        | Heating period | -7.31     | -11.29     | -11.22     | -16.56     |
| Mean daily heat residual | %        | Snowmelt period | -3.86     | -3.86      | -6.60      | -8.01      |
| LW$_{net}$: total heat loss | %          | Heating period | 63.52     | 64.02      | 75.61      | 89.75      |
| LE: total heat loss  | %          | Heating period | 33.67     | 30.68      | 25.04      | 13.59      |
| SE: total heat loss  | %          | Heating period | 2.81      | 3.3        | 0          | 0          |

*2017 discharge data only for days 1–273 due to instrument failure.

![Fig. 3.](a) Daily average outlet discharge (gray shaded polygon, m$^3$ h$^{-1}$, left y-axis) and the advective component of the lake’s heat budget (red solid line, right y-axis), computed as the daily residual (%) between measured whole-lake heat content ($Q_t$) and the expected heat content based on surface heat fluxes ($Q_m$). Dark gray bars show daily precipitation totals (rainfall; y-axis range 0–55 mm). Panels are ordered by SWE magnitude rather than by calendar year.
Emerald Lake gained only 28% of its net surface heat flux in the first month after ice-off (April–May) in 2015, whereas in 2014, the lake gained 60% of its surface heat flux within a month of ice-off (May–June).

Interactions among climate factors govern lake warming during summer

Annual maximum outlet temperature varied by almost 7°C over the last 25 yr (Fig. 4A). Ice-off date ranged from 15 April to 28 July. The length of the heating period ranged from 48 to 124 d; because ice-off date was strongly correlated with SWE ($R^2 = 0.86, p < 0.001$, df = 23), the heating period was longer in years with less snow.

Regression models confirmed that interactions between snowpack, ice cover, and summer weather patterns (cloudy days and rain events) drive lake warming during summer (model fits in Supporting Information Table S1). SWE and maximum outlet temperature were negatively related ($R^2 = 0.72, p < 0.0001$), but in dry years maximum lake temperature was unrelated to SWE (Fig. 4A). Ice-off date was negatively related to water temperature, though SWE was a better predictor ($\Delta AIC_c > 5$; Supporting Information Table S1).

Maximum lake temperature varied more among wet years (CV = 11.6%) than among dry years (CV = 2.6%). In dry years, there was appreciable variability in the number of net loss days, and unsurprisingly a greater number resulted in lower lake temperature (Fig. 4B). The best model of lake temperature included SWE, the number of net loss days, and their negative interaction (model $AIC_c$ weight = 1.0; Supporting Information Table S1). Summer rainfall had a similar cooling effect in low-snow years as the number of net loss days, and similar mechanisms of heat loss were likely operative. Mean air temperature also had an interactive effect with SWE on lake temperature, but this model explained substantially less variation than models including net loss days ($\Delta AIC_c > 12$).

Discussion

Recent climactic variation and corresponding thermal regimes in Emerald Lake spanned nearly the entire range of variation found over the last 30 yr (Sadro et al. 2018a). SWE varied by an order of magnitude, including the driest year on record (2015), ice-off date varied by over 2 months, and maximum epilimnion temperature varied by 5°C. In wet years, the lake is colder primarily because ice cover blocks incoming radiation and prevents heating, whereas in dry years, spring and summer weather patterns such as cloud cover and rainstorms determine lake warming by influencing the relative magnitude of heat gains and losses. Our results suggest that as snowpack declines in the Sierra Nevada and other mountain ranges (Mote et al. 2018; Sun et al. 2019), lake temperature will become increasingly sensitive to atmospheric forcing with reduced ice cover (Niedrist et al. 2018). Lakes in other warming regions may exhibit similar shifts in the relative importance of factors affecting their heat budgets as ice cover declines and lakes are exposed to spring weather patterns. Research on climate impacts in lakes has mainly focused on detecting decadal-scale temperature trends (Schneider and Hook 2010; O’Reilly et al. 2015; Winslow et al. 2015). However, as ice-covered lakes become increasingly sensitive to seasonal weather patterns, there is a need to explore finer-scale lake responses to interannual and seasonal climate variability and extremes, which can dramatically influence chemical and biological processes (Bertani et al. 2016; Perga et al. 2018).

In years with large snowpack, late ice-off, large inflows from snowmelt, and short water residence time reduced the lake’s temperature. Maximum heat content in Emerald Lake

![Fig. 4. (A) Maximum outlet temperature for each year from 1992 to 2017 (filled circles) vs. maximum annual SWE (x-axis). Years included in the heat budget analysis are identified as red-filled circles. SWE and maximum outlet temperature are negatively related in wet years (black solid line; SWE > 1000 mm; $p < 0.001$, $R^2 = 0.88$, df = 12), but unrelated in dry years (gray polygon corresponds to the y-axis extent in panel B). (B) Maximum outlet temperature in dry years (SWE < 1000 mm) vs. number of net loss days within the heating period (see “Data and methods” section for definitions).](image-url)
was 18% lower in 2017, a wet year in which ice-off occurred in early July, than in 2014, a dry year with ice-off in late May. Compared with a typical dry year, the potential energy “missed” by the lake in 2017 due to longer ice cover was a substantial fraction (70%; 1.71 × 10^{13} J) of the energy gained during the heating period. Snowmelt further reduced the lake’s heat content by up to 16% in 2017 (maximum advective Table 1). However, the effects of meltwater and residence time were small relative to the differences in lake heat content among years, and generally disappeared within a month of ice-off (Fig. 3A–D, Table 1). Regression analyses of the historical data set confirm the relative importance of these two mechanisms linking SWE and lake temperature: although ice-off day explained 58% of the variation in maximum temperature (p < 0.0001), SWE was a significantly better predictor (R^2 = 0.70, p < 0.0001, ΔAICc > 5, Supporting Information Table S1), likely because it integrates the effects of ice cover, meltwater inputs, and reduced water residence time.

In dry years, lake warming did not respond linearly to ice-off or snowmelt. In 2015, maximum lake heat content was 5% lower than in 2016, a year with almost four times greater SWE and 50 additional days of ice cover. 2015 was distinguished from other dry years by (1) an ice-free lake surface in April, when cold fronts are still common in the Sierra Nevada (Fig. 2B), and (2) the occurrence of multiple summer rainstorms (Fig. 3A). In spring, cloud cover and associated shifts in air temperature, humidity, and wind resulted in stalled lake warming or even periods of cooling through a combination of reduced radiation inputs (SWin), and to a lesser extent due to enhanced heat losses (Fig. 2B). Later in summer, rainstorms caused lake cooling by similar mechanisms. Comparing lake warming trajectories among dry years illustrates how early ice-off may not necessarily yield warmer temperatures: Emerald Lake gained only 28% of its net surface heat flux in the month after ice-off in 2015 due to the frequent number of cloudy days, whereas in 2014, the lake gained 60% of its total heat flux within the same time. Long-term measurements corroborate the influence of early-season weather patterns on lake warming in dry years. The best model of maximum lake temperature included a negative interaction term between SWE and number of net loss days, for example, the lake was colder than “expected” in dry years with frequent cloudy, cold days (Fig. 4B; model 1 in Supporting Information Table S1). 2015 represents an extreme example of the influence of spring and summer weather in years with early ice-off: maximum outlet temperature was 2.7°C colder than predicted by SWE alone, and the lake experienced 96 net loss days during the heating period, resulting in the lowest maximum temperature among all the dry years (Fig. 4B).

The Sierra Nevada are predicted to lose 70% of their snowpack by 2100 under “business as usual” scenarios (Sun et al. 2019), with significant losses below 2500 m elevation. Emerald Lake (2800 m) is typical of many high elevation lakes in terms of size and depth, and its responses to climate variation imply that lakes will warm as ice cover and snowmelt discharge decrease. However, as snowpack decreases below threshold values, spring weather patterns will exert a stronger influence on lake thermal trajectories, albeit within a narrower temperature range than historically observed (Fig. 4A; Sadro et al. 2018a,b). 2015 represents a potential future scenario for high elevation lakes, in which the warming influence of early ice-off and low SWE are partially offset by net heat loss during cold fronts and rainstorms. Somewhat counterintuitively, lakes will not necessarily be warmest in years with the earliest ice-off if exposure to spring fronts causes slower warming rates and delays stratification (Fig. 2A vs. Fig. 2E).

Predicting lake responses to future climate is constrained by uncertainty in climate expression at regional and local spatial scales, as well as how landscape features mediate climate responses by lakes. Air temperature in the western U.S. is predicted to rise (Hayhoe et al. 2004), and snowpack is expected to decline, but changes in cloud cover, wind speed, and summer precipitation are less certain. Furthermore, rates of climactic change differ across elevational gradients (Pepin et al. 2015), and topography influences climate expression at local scales in mountainous terrain, for example, through formation of temperature inversions in valleys (Lundquist and Cayan 2007), or through retention of snow and ice on steep, north-facing slopes (Tennant et al. 2017). Lake and watershed morphometry generate variable lake ice cover and thermal regimes within regions subject to similar climate forcing (Weyhenmeyer et al. 2011; Arp et al. 2013; Winslow et al. 2015; Rose et al. 2016). Lake depth mediates thermal responses to variation in SWE (Sadro et al. 2018a), with shallow lakes showing greater increases in surface temperature than deep lakes in dry years. Lake size influences heat budgets and mixing dynamics, with larger lakes more influenced by wind-driven mixing and heat fluxes than smaller lakes (Gorham 1964). LW_net is the largest heat loss flux in Emerald Lake, but for larger lakes in wind-exposed catchments, evaporative heat fluxes (LE) are likely more important (MacIntyre and Melack 2009).

Global syntheses show warming trends in high elevation lakes (O’Reilly et al. 2015; Christianson et al. 2019), which will affect a broad suite of physical, chemical, and biological processes (Moser et al. 2019). Yet, relative to their global abundance (>11 million above 2100 m; Verpoorter et al. 2014), temperature measurements in high elevation lakes remain rare (Christianson et al. 2019), and studies characterizing drivers of climate-sensitivity are even more so. The nonlinear relationship between SWE and peak temperature in Emerald Lake likely varies in form for other mountain lakes according to latitude, elevation, watershed attributes, and basin morphometry (Sadro et al. 2018a). Regional studies focused on mediators of lake climate-sensitivity, as have been undertaken in low-elevation lake districts (McCullough et al. 2019), will greatly improve our ability to predict changes in the ecosystem function of high elevation lakes under future climate regimes.
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279