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Key Points:

- Modeling the water budget of lakes in Taylor Valley demonstrates their sensitivity to changes in available energy rather than precipitation
- The main source of water to the lakes is glacial melt; precipitation (snow) reduces inflow to the lakes because it increases glacier albedo
- Lakes are out of balance with the current climate; the differing rates of lake level rise are due to watershed glacier cover and hypsometry

Supporting Information:

Supporting Information may be found in the online version of this article.

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Physical Controls on the Hydrology of Perennially Ice-Covered Lakes, Taylor Valley, Antarctica (1996–2013)

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Abstract The McMurdo Dry Valleys, Antarctica, are a polar desert populated with numerous closed-watershed, perennially ice-covered lakes primarily fed by glacial melt. Lake levels have varied by as much as 8 m since 1972 and are currently rising after a decade of decreasing. Precipitation falls as snow, so lake hydrology is dominated by energy available to melt glacier ice and to sublimate lake ice. To understand the energy and hydrologic controls on lake level changes and to explain the variability between neighboring lakes, only a few kilometers apart, we model the hydrology for the three largest lakes in Taylor Valley. We apply a physically based hydrological model that includes a surface energy balance model to estimate glacial melt and lake sublimation to constrain mass fluxes to and from the lakes. Results show that lake levels are very sensitive to small changes in glacier albedo, air temperature, and wind speed. We were able to balance the hydrologic budget in two watersheds using meltwater inflow and sublimation loss from the ice-covered lake alone. A third watershed, closest to the coast, required additional inflow beyond model uncertainties. We hypothesize a shallow groundwater system within the active layer, fed by dispersed snow patches, contributes 23% of the inflow to this watershed. The lakes are out of equilibrium with the current climate. If the climate of our study period (1996–2013) persists into the future, the lakes will reach equilibrium starting in 2300, with levels 2–17 m higher, depending on the lake, relative to the 2020 level.

Plain Language Summary The McMurdo Dry Valleys (MDV) in Antarctica are a polar desert, characterized by a cold and dry climate. A number of permanently ice-covered lakes occupy the MDV. These lake are fed by melt from nearby glaciers, and many lakes occupy watersheds with no outlet. Lake levels have changed dramatically in the past. Precipitation in the MDV falls as snow, contributing very little water directly to the lakes. Lake hydrology is controlled by the energy available to melt glacier ice and to sublimate ice from the frozen lake surface. To understand the energy and water balance controls on lake level changes, we model the hydrology for the three largest lakes in Taylor Valley. We use a numerical model, based on real world hydrological processes, to estimate the water balance of these lakes. The model results show that lake levels are highly responsive to small changes in glacier albedo, summer air temperature, and wind speed. The lakes in Taylor Valley are out of balance with the current climate. If the climate of our study period (1996–2013) continues into the future the lakes will reach levels 2–17 m higher, depending on the lake, relative to the 2020 level.

1. Introduction

Lakes are recognized as sentinels of climate change owing to their sensitivity to small climatic perturbations (Adrian et al., 2009; McCullough et al., 2019; Williamson et al., 2009). The water budget of closed watershed lakes is particularly sensitive to weather and climate variations (Jones et al., 2001; Mason et al., 1994). Polar lakes, even those covered with perennial lake ice, are no exception (Castendyk et al., 2016). For example, some perennially ice-covered high Arctic lakes have experienced a shift from perennial to seasonal ice covers due to surface air temperature warming (Lehnherr et al., 2018). Likewise, some Antarctic lakes have dramatically increased in lake levels in a response to a few weeks of anomalously warm temperatures, replacing the water lost in the previous decade of water loss (Doran, Priscu, et al., 2002; Doran et al., 2008). However, the same weather/climate forcing can lead to different responses in close-proximity lakes (Gooseff et al., 2011; Obryk et al., 2016). Here we examine the physical processes controlling lake level change and its spatial variability at three polar lakes located in Taylor Valley, McMurdo Dry Valleys, Antarctica (MDV).

Writing – review & editing: J. M. Cross, A. G. Fountain, M. J. Hoffman, M. K. Obryk

The purpose of this study is to elucidate the physical processes that govern the hydrologic budget of the lakes. We address the questions of what drives seasonal and long-term changes in lake level, why the lakes respond in individual ways to the same weather and climate influences, and what future levels may be expected given the range of climate variations observed to date. A hydrological model is presented that couples a glacier ablation model to a lake ablation model to solve the lake budget. Model outcomes inform our understanding of the climate, hydrology, and controls on glacial melt in the region.

2. Site Description

The MDV are a polar desert located along the west coast of the Ross Sea at about 77.5° south and are the largest ice-free area, 4,500 km^2, on the continent (Figure 1; Levy, 2012). The region is of intense interest because the ice-free landscape is one of the few places on the continent in which the geology is exposed, showing geomorphic and geochemical evidence of past lake levels that possibly record paleoclimate conditions (Hall et al., 2010; Myers et al., 2021; Toner et al., 2013; Zieg & Marsh, 2012). The Valleys also hosts a unique microbial ecosystem that inhabits the various landscape elements including soils, streams, lakes, and glaciers (Fountain et al., 1999; Gooseff et al., 2011; Lyons et al., 2001; McKnight et al., 1999). Because of the dry and cold environment, the MDV are often used as a terrestrial analog for planetary processes (Doran et al., 1998; Levy et al., 2008; Marchant & Head, 2007).

The MDV are characterized by broad expanses of sandy gravelly soils, punctuated by bedrock outcrops. Ephemeral streams drain meltwater from the glaciers that descend into the valleys. Most streams terminate in the closed lakes and ponds that dot the valley floor. Ice-cemented permafrost is continuous below an active layer (soil that thaws each summer) 20–45 cm deep (Bockheim et al., 2007). No significant groundwater movement is known to occur below the active layer, but within the active layer, hyporheic zones along the streams, wetted margins surrounding lakes, and water tracks, water movement is common (Gooseff et al., 2007, 2011; Levy et al., 2011). Sunlight is continuous during the austral summer and absent during the winter months. Annual air temperatures average about −20°C while mean summer temperatures hover below 0°C, occasionally warming above freezing for a few hours to days (Doran, Priscu, et al., 2002; Obryk et al., 2020). Historically, precipitation falls as snow, and what little accumulates on the valley floor, 3–50 mm w.e. yr^{-1} depending on location, mostly sublimates before making a significant contribution to soil moisture (Chinn, 1981, 1993; Eveland et al., 2013; Fountain et al., 2010; Obryk et al., 2020). Occasionally, and with more frequency in the winter, foehn winds descend into the valleys from the polar plateaus reaching speeds of 35 m s^{-1} (Doran, McKay, et al., 2002; Nylen et al., 2004; Speirs et al., 2010). They adiabatically warm as they descend and rapidly increase air temperature in just a few hours.

The focus of our study is Taylor Valley because it has a wealth of hydro-meteorological measurements collected by the McMurdo Dry Valley Long Term Ecological Research project since the early 1990s (https://mcm.lternet.edu/; e.g., Doran & Fountain, 2019; Doran & Gooseff, 2018; Dugan et al., 2013; Fountain et al., 2016; Gooseff & Fountain, 2022; Gooseff & McKnight, 2016; Obryk et al., 2020). The valley is 24-km long and is open to the Ross Sea at its eastern end, although a ∼90 m tall glacial moraine blocks stream drainage to the ocean. The Kukri Hills confine the valley to the south and Asgard Range to the north, both of which rise abruptly to over 2,000 m in places. Alpine glaciers, up to 50 km^2 in area, descend from the surrounding mountains. These glaciers are considered polar glaciers with internal and basal temperatures below freezing. The cold, dry, and windy conditions typically keep the glacier surfaces below freezing, favoring sublimation over melt (Hoffman et al., 2014; Lewis et al., 1998). Only the near-surface ice warms to melting temperatures, feeding ephemeral streams that flow for about 10 weeks during the austral summer from November to February (Bergstrom et al., 2021; Hoffman et al., 2016; McKnight et al., 1999).

Three major lakes, all perennially ice-covered, occupy the valley bottom, Lakes Fryxell (7.81 km^2, based on the 2020 datum), Hoare (2.36 km^2), and Bonney (4.66 km^2). Although the lakes are ice-covered, a narrow moat of water forms in summer along the shoreline where the lake is shallow and the ice thin. The only significant water gain to the lakes is from glacial meltwater and water loss is through sublimation from the ice-cover and evaporation from the moats (Chinn, 1993; Dugan et al., 2013). Lake levels exhibit an annual cycle with increasing level in the summer, due to the flush of glacial meltwater, reaching peak levels in late January, and a slow winter decline as the ice surface sublimates reaching minimum levels in late-November (Chinn, 1981). The lake ice melts extensively in summer but the water returns to the lake. Model estimates of sublimation suggest an average loss rate

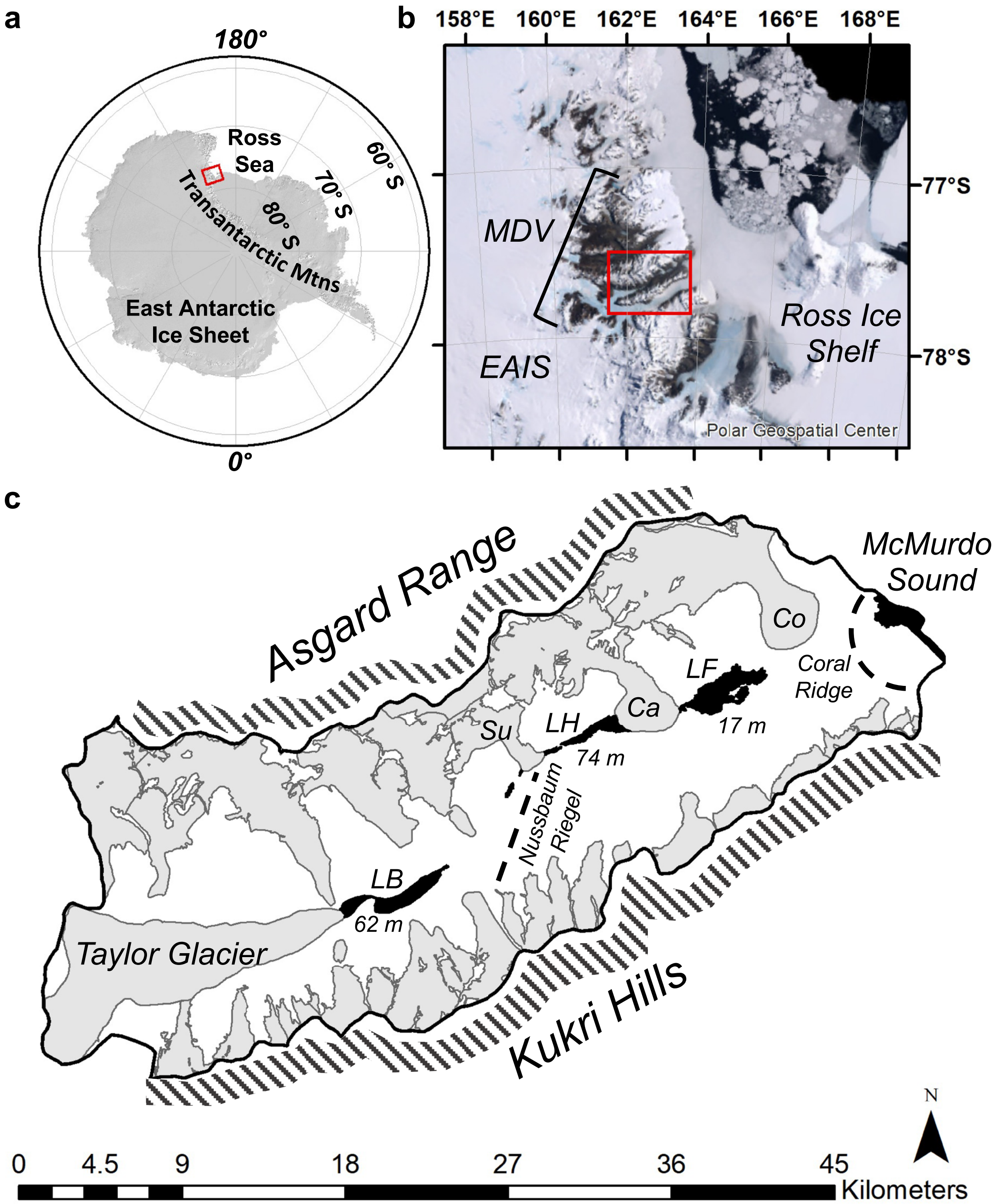

Figure 1. Map of the McMurdo Dry Valleys (MDV) in the Ross Sea region of Antarctica (a, b). The MDV are transverse valleys of the Transantarctic Mountains. Map of Taylor Valley study site (c) showing Taylor, Suess (Su), Canada (Ca) and Commonwealth (Co) Glaciers, and Lakes Bonney (LB), Hoare (LH), Fryxell (LF), and the East Antarctic Ice Sheet (EAIS). Approximate lake surface elevations (based on 1996 surveyed datum) are shown.

of about 0.35 w.e. yr^{-1} (Clow et al., 1988). The lake levels have been rising since 1903 (Bomblies et al., 2001; Chinn, 1993; Scott, 1905). From 1972, the start of near annual measurements, lake levels rose until 1993 at a rate of $+0.23$ m yr^{-1} at Lake Bonney and $+0.09$ m yr^{-1} at Lakes Hoare and Fryxell (Chinn, 1993; Doran et al., 2008). This was followed by a decade of gradual decline (-0.06 to -0.08 m yr^{-1}). In the summer of 2001/2002 an anomalous warm period of about 2 weeks dramatically increased glacial melt and rapidly increased lake level (Doran et al., 2008). Between 2001 and 2014 lake levels rose, with Lake Bonney rising at higher rate ($+0.32$ m yr^{-1}) than

Lakes Hoare or Fryxell (+0.13 m yr^{-1}). Between 2014 and the most recent measurement available, January 2020, lake levels have remained fairly constant.

3. Model

The hydrology of Taylor Valley is treated as a simple system in which glacial meltwater streams, with no instream losses, are the only water source to the closed-watershed lakes. Snowfall and groundwater contributions to the hydrological budget are ignored for the reasons given earlier. In-stream water losses, evaporation, and flow into hyporheic zone are compensated by melting wind-drifted snow that collects in the stream channels and the return of water frozen in the hyporheic zone over winter (Gooseff et al., 2013; Wright et al., 2021). The only mass loss from the system is sublimation from the lake ice, which occurs year-round (Chinn, 1993; Clow et al., 1988). Although lake ice melts in summer, the water largely percolates through the near-surface porous ice back to the lake water below (Adams et al., 1998; Dugan et al., 2013). Evaporative losses from the narrow moats surrounding each lake are minor and ignored because moat area accounts for a small fraction of the total lake area (Supporting Information S1; Chinn, 1993; Spigel & Priscu, 1998; Wharton et al., 1986). With these assumptions, the volume change rate of the lakes can be expressed as,

$$\frac{\mathrm{d}V}{\mathrm{d}t} = Q_{\text{glacier}} - S * A(h), \tag{1}$$

where Q_{glacier} (m^3 d^{-1}) is the meltwater inflow from the glaciers, S is sublimation from the lake surface (m d^{-1}), and A (m^2) is the area of the lake at lake level, h (m).

The change in lake volume V (m^3) for a given change in level, h (m), can be calculated from lake area, $A(h)$ (m^2), considering lake bathymetry and valley hypsometry through

$$\frac{\mathrm{d}V}{\mathrm{d}t} = A(h)\frac{\mathrm{d}h}{\mathrm{d}t}. \tag{2}$$

From Equations 1 and 2, the lake level change with time can be expressed as,

$$\frac{\mathrm{d}h}{\mathrm{d}t} = \frac{Q_{\text{glacier}}}{A(h)} - S. \tag{3}$$

$A(h)$ is known for all lake watersheds (Fountain et al., 2017; Obryk, 2014), and Q_{glacier} and S are calculated. MATLAB code, adapted from Obryk et al. (2017), is used to numerically solve (3) for the daily lake level change at Lakes Bonney, Hoare, and Fryxell using a finite-difference method.

Glacier melt contribution to the lakes is the sum of subaqueous melt at the glacier-lake interface, and glacial runoff from the sub-aerial ice surfaces,

$$Q_{\text{glacier}} = Q_{\text{subaqueous}} + Q_{\text{runoff}}. \tag{4}$$

Subaqueous melt is restricted to two glaciers, Canada Glacier (Lakes Hoare and Fryxell) and Taylor Glacier (Lake Bonney) and its flux is estimated from the speed of glacier at its terminus assuming the terminus position had not changed. Over the past ∼50 years the position of glacier termini in Taylor Valley have changed less than 1 m (Chinn, 1998; Fountain et al., 2016). We use a value of 5 m yr^{-1} (Pettit et al., 2014) and 1.5 m yr^{-1} (Telling et al., 2017) as an estimate for the terminus velocity at Taylor and Canada Glaciers, respectively. The melt volume is the product of the subaqueous melt flux, corrected for density (870 kg m^{-3} for Taylor Valley glaciers; Hoffman, 2011), and the area of ice-lake contact. The Taylor Glacier, however, had advanced 93 m between 1977 and 1997 (Fountain et al., 2004), and returned to its previous 1977 position by 2006 (personal observation). This transient advance results in a small additional volume of lake water and is ignored and we assume a constant flux of meltwater. Subaqueous glacial melt was estimated to be 31,800 and 31,500 m^3 yr^{-1} into Lakes Hoare and Bonney, respectively. Independently, Spigel et al. (2018) calculated a similar subaqueous water flux from Taylor Glacier of ∼34,000 m^3 yr^{-1}. The ice-lake contact area between Lake Fryxell and Canada Glacier is minimal, reaching at a maximum ∼1,500 m^2 (2022 datum). Based on this maximum contact area, subaqueous contributions to Lake Fryxell would amount to 0.1% of the mean annual inflow to the lake and is therefore ignored.

To estimate Q_{runoff}, we employ ICEMELT, a physically based energy balance model that predicts ice melt (Liston et al., 1999). Hoffman et al. (2008, 2014, 2016) modified the model and optimized the parameters for conditions in Taylor Valley. They showed an inclusion of solar penetration into the ice was required to accurately predict ice ablation and temperature and, where present, the formation and drainage of internal melt and associated density changes. Here, we apply the distributed form of the model (Hoffman et al., 2016) to predict streamflow from all the glaciers in Taylor Valley. The effect of snow cover in suppressing glacier melt is dealt with implicitly through albedo in that when snow is present the higher albedo is sufficient to shut off melt. Details of the model can be found in Hoffman et al. (2008, 2014, 2016) and our application of the model in the Supporting Information S1. In brief, ICEMELT solves the surface energy balance in the form of,

$$\chi(1-\alpha)\,Q_{\mathrm{si}} + Q_{\mathrm{li}} + Q_{\mathrm{le}} + Q_{\mathrm{h}} + Q_{\mathrm{e}} + Q_{\mathrm{c}} = Q_{\mathrm{M}} \tag{5}$$

where χ accounts for a fraction of absorbed solar radiation at the air-ice interface, α is the albedo of the surface, Q_{si} the incoming solar radiation flux, Q_{li} the incoming longwave radiation flux, Q_{le} the emitted longwave energy flux, Q_{h} the turbulent flux of sensible heat, Q_{e} the turbulent flux of latent heat, Q_{c} the conduction of heat in the ice, Q_{M} the energy available for melt. Incoming short- and longwave radiation (Q_{si} and Q_{li}) are directly measured and albedo (α) is calculated from the ratio of measured incoming and outgoing shortwave radiation. The other terms cannot be measured directly but relate to meteorological conditions at the surface and thus can be cast in form that leaves ice surface temperature (T_0) as the only unknown.

The modeling domain employs a 250-m grid covering only the ice-exposed surfaces of the glacier ablation zones (~75 km^2). The accumulation (snow) zones are not included because melt is uncommon and what does occur refreezes in the cold snow beneath (Fountain et al., 1999). The high snow albedo reflects much of the solar radiation and what is absorbed cannot compensate for the turbulent heat losses to the below-freezing atmosphere. Indeed, when a snow event blankets the ablation zone melt-runoff largely ceases and does not resume for days to weeks until the snow ablates (Fountain et al., 1998). For Taylor Glacier, which has large expanses of exposed ice, the summer melt limit is set at 400 m above sea level, above which little if any melt occurs. The model is run at hourly time steps because daily steps miss important melt events (Hoffman et al., 2014). Prior to each model run, the initial ice temperature for all cells is set to the mean annual air temperature based on the 17-year weather record (described below) and the spin-up period for the model is 17 years using the same record.

The topography of the ablation zones varies dramatically across any individual glacier and between glaciers, particularly at lower elevations, creating significant variations in the micro-meteorological conditions (Chinn, 1987; Hoffman, 2011; Johnston et al., 2005; Lewis et al., 1999). Hoffman (2016) characterized and mapped the roughness into two zones using a 2-m light detection and ranging (LiDAR) digital elevation model (DEM; Csatho et al., 2005). These zones correspond to relatively smooth surfaces with roughness <0.1 m, which cover much of the upper ablation zone, and deep basins with roughness of ~3 m, which occur on a number of glaciers at low elevations. The fixed meteorological stations are located on the smooth surfaces and come closest to the ideal for application of bulk energy methods. Rover meteorological stations were used to characterize conditions in the basins (Hoffman, 2011; Johnston et al., 2005; Lewis et al., 1999). Constant offsets between conditions in the basins relative to the fixed stations were estimated and used to predict basin meteorology over the period of record (Table 2 in Hoffman et al., 2016). Compared to the smooth surfaces, the basins have slower wind speeds, warmer air temperatures, and lower albedo due to partially frozen pools of water that typically cover the basin floors. The melt rate for the basins were increased by +20% to account for the basin walls, which are not resolved by the DEM (Hoffman et al., 2016). Finally, the glaciers terminate in 25-m tall cliffs and they were modeled separately using their average aspect for relative uniform segments along the perimeter. For glaciers that drain to multiple streams, surface catchments were defined using the same DEM of the valley (Csatho et al., 2005). Glacial meltwater is assumed to drain instantly to the streams.

Mass loss from the lakes is assumed to be solely through sublimation from the lake ice. To be consistent with our methodology for the glacier surfaces, we calculated sublimation from lake ice using ICEMELT's calculation of latent heat flux using the bulk aerodynamic method,

$$Q_{\mathrm{e}} = \rho_{\mathrm{a}} L_{v} D_{\mathrm{e}} \zeta \left(0.622\,\frac{e_{\mathrm{a}} - e_{\mathrm{s}}}{p_{\mathrm{a}}}\right), \tag{6}$$

where ρ_a (kg m^{-3}) is the density of air, L_v (J kg^{-1}) is the latent heat of vapourization, D_e is the turbulent exchange coefficient, ζ is a unit-less stability function, e is vapor pressure (Pa) and p is measured air pressure (Pa), measured at the Lake Hoare meteorological station, while subscripts s and a denote conditions at the surface and at the reference height in the air above the ice, respectively. The vapor pressure at the reference height was calculated from the measured temperature and relative humidity at 3 m and estimated at the lake ice surface. For air temperatures <0°C the humidity at the lake ice surface was assumed to be 100% with respect to ice and ice skin temperature equal to the air temperature; for air temperatures ≥0°C, 100% humidity with respect to water and ice skin temperature = 0°C. The stability function was assumed stable if the air temperature was >0°C, unstable if <0°C, and neutrally stable ($\zeta = 1$) if = 0°C.

The turbulent exchange coefficient, D_e, takes the form:

$$D_e = \frac{\kappa^2 u_a}{\left[\ln\left(z_a/z_0\right)\right]^2}, \tag{7}$$

where κ is von Kármán's constant and u_a is the wind speed at the reference height z_a. The surface roughness length, z_0, is the only adjustable parameter and is calibrated independently for each lake. Roughness was adjusted to match observed lake level change and held constant for the entire model run. However, different values were used for summer and winter. Summer months of December and January have a higher roughness to account for melt-roughening caused by patches of sediment preferentially melting into the ice and by the penetration of solar radiation melting the subsurface and collapsing the ice surface (Dugan et al., 2013; Jepsen et al., 2010). During winter, the ice is smoothed through preferential sublimation of the ridges and peaks of the rough surface, as has been observed in the field.

3.1. Model Forcing

The model is forced using Hoffman et al.'s (2016) grid of interpolated meteorological data from the eight stations in Taylor Valley. Four stations are located on the valley floor, three of which are adjacent to the lakes, and four are located on the glaciers. The variables include, air temperature, relative humidity, wind speed, and incoming and reflected short wave solar radiation, and were recorded at 30-s intervals then averaged over 15-min and stored (Doran & Fountain, 2019; Obryk et al., 2020). MICROMET (Liston & Elder, 2006) assimilated the meteorological data and interpolated values to a grid with 250-m nodal spacing based on known relationships between meteorological variables and topography. The results were averaged to hourly intervals for the model. Daily average albedo was calculated from the 15-min values of incoming and reflected solar radiation. For details on how meteorological input data were processed with MICROMET, see Hoffman (2011).

The albedo of glacier surfaces is highly variable due to spatial variations of supraglacial rock debris, sediment, and transient snow events (Bergstrom et al., 2020; Fountain et al., 1998). Acknowledging this variation, we initially used the daily albedo values from the in situ record averaging the values measured at Taylor and Canada Glaciers, as did Hoffman et al. (2016). By February 2000, global albedo data became available from the Moderate Resolution Imaging Spectroradiometer (MODIS) satellite instrument (Schaaf & Wang, 2015). Daily MODIS-derived albedo images (pixel size 500-m) covering Taylor Valley were downloaded and albedo values for cells contained within the domain boundary of each glacier were averaged for that glacier. Only Taylor, Canada, and Commonwealth Glaciers were large enough to fully encompass MODIS cells. These MODIS values were considered representative of smaller neighboring glaciers. For glaciers west of the Suess Glacier, the Taylor Glacier MODIS albedo values were used, while the Canada Glacier values were used for Suess Glacier. For the smaller alpine glaciers in the Lake Fryxell watershed the albedo from Commonwealth Glacier was used but reduced by 10% to account for the visually darker surfaces found on these. When albedo wasn't available due to cloud cover, the Canada-Taylor Glacier average of in situ values were used. The resulting time series of albedo was smoothed using a 7-day moving average. On average 58 days per year, around half the number of days in the melt season, had suitable MODIS values. Comparison of MODIS albedo to in situ measurements at meteorological stations on the three glaciers of sufficient size and with meteorological stations (Taylor, Canada, and Commonwealth Glaciers) is good, however, the in situ measurements were higher at all sites than the MODIS albedo by 0.18 on average (Supporting Information S1). This bias in the in situ values is not surprising as spatially variable snow accumulation on the glacier surface has less effect on the MODIS values since each pixel integrates over a much larger than the in situ pyranometer area (500 vs. ∼10 m^2). For this reason we did not use in situ measurements to adjust MODIS albedo values.

3.2. Calibration

The MDV Long Term Ecological Research project has collected hydro-meteorological data in Taylor Valley since the early 1990s. In addition to the meteorological stations, glacier mass balance is measured on four glaciers, stream discharge measured at 15 streams, and lake level and lake ice ablation on the three major lakes in the valley (Doran & Gooseff, 2018; Dugan et al., 2013; Fountain et al., 2016; Obryk et al., 2020). Glacier mass balance is measured twice a year, in late spring (November) and late summer (late January; Fountain et al., 2016). Seasonal values of snow accumulation and ablation are available at a number of stakes set into the smooth ice across the ablation zone and have a low degree of uncertainty, ≤10% (Fountain et al., 2006). No basin floors are measured.

Streams are continuously monitored in summer (Gooseff & McKnight, 2016). Flumes and weirs are present at all measured streams and stream stage is recorded at 15-min intervals. Discharge measurements, made several times each summer, are used to update the rating curve that converts stage to discharge. Daily values of discharge are provided with an uncertainty. Lake measurements include water levels and lake ice ablation (Doran & Gooseff, 2018; Dugan et al., 2013; Obryk et al., 2020). In the early 1990s, lake levels were determined by optical surveying once in spring and again in late summer. By the late 1990s (2004 for Lake Fryxell) installation of pressure transducers provided continuous lake level measurements. These methods typically have a low degree of uncertainty (±1%). Lake ice ablation was measured several times over a summer, by measuring the height of a stake set in the ice. Winter ablation, mostly sublimation, was estimated from the decrease in lake level.

The hydrologic model was tested in several different and independent ways. Modeled glacier ablation was compared to measurements on Taylor and Canada Glaciers (Gooseff & Fountain, 2022). Modeled glacial runoff was compared to the measured stream runoff. Estimated winter lake ice sublimation was calibrated by adjusting the value of winter ice roughness to match measured value inferred from the decrease in lake level. To close the hydrological budget the summer value of lake ice roughness was adjusted to match the modeled lake levels to the observed over the period of record. The upper limit of summer sublimation could not exceed the measured summer ablation.

4. Initial Results and Modifications

4.1. Meltwater

The model was run from 1 July 1996 to 1 February 2013, capturing 17 water years, defined as April to March. The initial application of ICEMELT, with only the meteorological station albedo, showed good agreement with observed values during the early years of the record (1996–2001) when the climate was cool and stream discharge was small. In warmer summers, starting in 2002, predicted discharge was much smaller than measured. Three adjustments to the model were made to better represent the glacier albedo, near-surface air temperature lapse rate, and melt within the basin sub-domain.

Application of MODIS albedo, including a valley-wide albedo reduction of −7%, improved the post-2007 streamflow predictions for glaciers in Lake Fryxell watershed. This adjustment was motivated by the hypothesized influx of sediment after 2006 suggested by Hoffman et al. (2016) and is within the uncertainty range of both MODIS and the in situ measurements. Hoffman et al. (2016) speculated that this sediment was incorporated into the subsurface ice, where it enhanced subsurface melting, already an important mode on these glaciers, but had less of an effect in lowering broadband surface albedo. No adjustment was applied to the basin floors because sediment is flushed from these surfaces and collects in the bottom of meltwater pools under the cover of partially frozen slush and probably does not affect albedo. The cumulative modeled ablation (1996–2013) for the smooth ice sub-domain underestimates measured by −3.1 cm w.e., representing 0.04% of total cumulative ablation. For reference, the average annual measured ablation across all stakes is 4 cm w.e.

Stream discharge from the glaciers in the eastern Kukri Hills, however, remained smaller than measured; even unrealistic reductions in albedo made little difference. Clouds are frequently observed in this part of the valley, particularly close to the coast (Acosta, 2016). They dissipate after advecting into the valley after a few kilometers. Increasing longwave radiation over the glaciers added little to the meltwater runoff. We argue that the

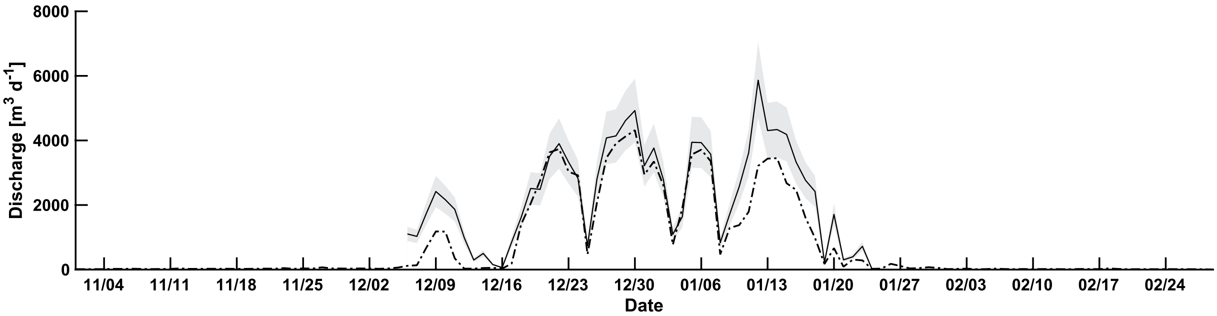

Figure 2. Daily discharge hydrograph for House Stream for the 2001/2002 water year. House Stream flows from the Suess Glacier into Lake Hoare. The solid line is observed stream discharge with ±20% measurement uncertainty (shaded area). Dashed line is predicted meltwater discharge.

near-surface air temperature lapse rate is different near the coast due to the higher humidity (Doran, Priscu, et al., 2002). Modeled lapse rate is assumed to be the dry adiabatic everywhere in the valley, an accurate assumption for environments up valley, westward of and including Canada Glacier (Doran, Priscu, et al., 2002; Hoffman et al., 2014). Near the coast, east of Canada Glacier we applied the wet adiabatic lapse rate. This adjustment had little effect on the air temperature and melting of low elevation ablation zones (Commonwealth Glacier) but it greatly increased temperature and melt at higher elevations on the glaciers the eastern part of the Kukri Hills.

Stream discharge from glaciers with extensive basin topography over elevation ranges of a few 100 m or so was also less than observed. We infer that the offset for basin microclimate was not constant but increased with elevation, implying that the basins do not cool as quickly as the surrounding atmosphere. Rather than adjust the various offsets, it was simpler to employ a scaling factor to the basin runoff, which increased linearly by 0.5 every 100 m.

The final meltwater model shows good skill in predicting daily streamflow variations (Figure 2 and Supporting Information S1). The Nash-Sutcliffe efficiency E (Nash & Sutcliffe, 1970), indicate good dynamical efficiency, $E = 0.84$, $r^2 = 0.85$ for all days with observed streamflow at all gauges. Total predicted streamflow accounted for 92% of total measured streamflow. Predicted annual streamflow totals, summed for all streams in Taylor Valley, fell within the uncertainty of observed values most years (Figure 3 and Supporting Information S1). Adjustments to account for evaporation in the stream channels did not improve the estimates.

4.2. Lake Ice Sublimation

Initial lake ice sublimation estimates, with calibrated surface roughness values, was sufficient to close the water budget at Lakes Bonney and Hoare, but not Lake Fryxell. Initial roughness values yielded summer values of 4, 8, and 0.05 mm for Lakes Bonney, Hoare, and Fryxell, respectively (Table 1). The value for Lake Fryxell is orders of magnitude too low based on field observations which suggest a roughness more comparable to Lake Hoare and rougher than Lake Bonney. These field observations are supported by an independent roughness metric that was calculated from elevations measured by an aerial LiDAR survey flown in December 2014 (Fountain et al., 2017). The standard deviation of elevations of 1-m cells for each lake showed that the roughness of Lake Fryxell was almost as large as Lake Hoare, and certainly larger than Lake Bonney (Table 2). We revised the roughness of Lake Fryxell to 7 mm, which caused sublimation losses to exceed inflow gains resulting in an annual water deficit averaging 450,000 $m^3\ yr^{-1}$ (∼1,200 $m^3\ d^{-1}$). Possible water sources to balance this deficit are discussed later. After adjusting the surface roughness at Lake Fryxell, estimated winter lake ice sublimation matched measured values at all three lakes (Figure 4). The calibration error at each lake was less than 15% of mean winter sublimation, with RMS errors of 12.6, 13.5, and 10.8 mm for Lakes Bonney, Hoare and Fryxell, respectively. Summer sublimation values stayed below measured ablation values for all lakes including Lake Fryxell (Figure 4).

4.3. Overall Model Results

The resulting modeled and measured lake level changes showed good correspondence with coefficients of determination greater than 0.87 and the root mean square error (RSME) equal to or smaller than 0.15 $m\ d^{-1}$ (Figure 5). Total difference in lake levels at the end of the 17-year period are less than 0.3 m. The seasonal performance of the model is good at reproducing the winter and summer lake levels. Examining the detail of an annual cycle reveals a

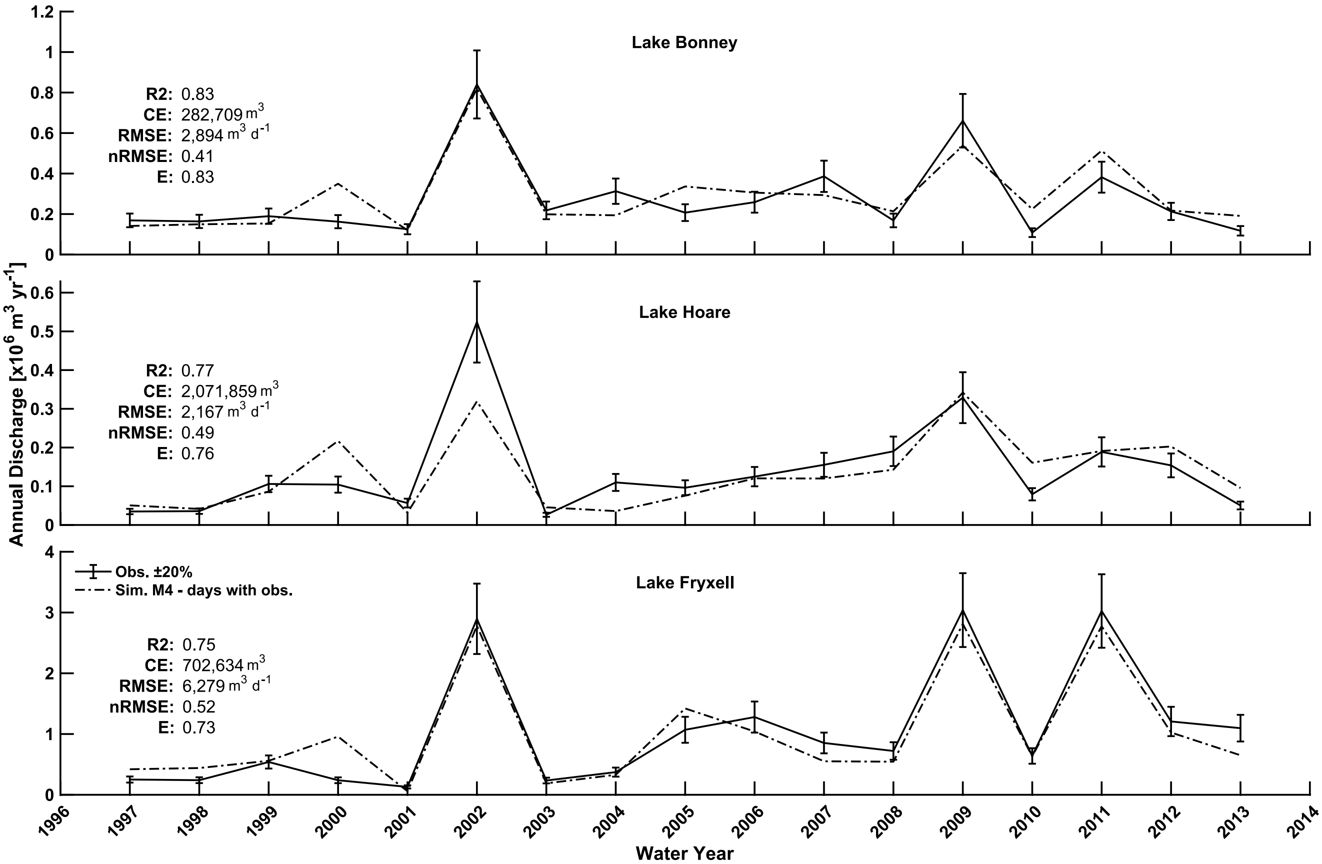

Figure 3. Annual inflow through measured streams into Lakes Bonney, Hoare, and Fryxell from water years 1996–2013. The *x*-axis label indicates the water year. The solid line is observed stream discharge with ±20% measurement uncertainty error bars. Dashed line is predicted meltwater discharge. Model performance is described for each lake using several metrics, including the coefficient of determination (R2), cumulative error (CE), root mean squared error (RMSE), normalized RMSE, and Nash-Sutcliffe efficiency (E).

number of important processes (Figure 6). Lake level decreases from April to December due to sublimation of the ice-cover in the absence of stream inflow. Winter periods of elevated sublimation correspond to warm temperatures and high winds during foehn events. Glacial streams begin to flow in November, but lake level continues to decrease due to elevated sublimation rates caused by warmer air temperatures. By mid-December, as air temperatures warm to near 0°C and solar radiation increases (not shown), meltwater inflow exceeds sublimation losses and lake levels rise. In mid-January air temperatures begin to cool. In this particular month, a late season snowfall blanketed the glaciers largely eliminating melt runoff and lake level begins its seasonal decline as sublimation losses exceed meltwater inflow.

Table 1
Winter (February–November) and Summer (December–January) Surface Lake Roughness (z_0)

	Winter roughness (mm)	Summer roughness (mm)	LiDAR roughness (mm)
Lake Bonney	0.086	4.00	35
Lake Hoare	0.122	8.00	48
Lake Fryxell	0.240	7.00[a] (0.05)	45

Note. Winter values were calibrated against lake ice ablation measurements, while summer values were found through closing the lake water budget. LiDAR-derived roughness, from early December 2014, is included to show the relative lake surface roughness. The LiDAR value is not the aerodynamic roughness but calculated from the standard deviation of elevation values within a 3-by-3 cell moving window. For reference, z_0 in the glacier meltwater model is 0.05 mm on smooth surfaces and 1 mm for basins.

[a]Based on the inclusion of an inferred snow melt flux of 450,000 m^3 yr^{-1} to close the water budget; original value shown in parenthesis.

4.4. Model Sensitivity to Environmental Variables

To examine the sensitivity of the model to the driving meteorological and environmental variables, changes were made to air temperature, wind speed, and albedo. The hourly wind speed and daily albedo values were adjusted by ±10%, while hourly air temperature was adjusted by ±0.275°C, which corresponds to 10% of the average summer air temperature. Because meltwater response will depend on how close the hourly air temperature value is to 0°C, a percent adjustment was not applied to the air temperature. The changes were applied to the entire period of record; the meltwater and sublimation changes were then averaged and the change in lake level is expressed for the last year of the record (Table 2). Meltwater response is similar for changes in

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Table 2
Glacier Meltwater and Lake Response Sensitivity to Changes in Environmental Variables

		Air temperature		Wind speed		Albedo	
	Base value	+0.275°C	−0.275°C	+10%	−10%	+10%	−10%
Glacier meltwater ($m^3\ a^{-1}$)							
Lake Bonney Watershed	3.79E+07	15%	−13%	−11%	14%	−53%	107%
Lake Hoare Watershed	1.13E+07	10%	−9%	−8%	9%	−53%	79%
Lake Fryxell Watershed	2.63E+07	8%	−13%	−12%	8%	−57%	92%
Taylor Valley	8.10E+07	12%	−12%	−11%	11%	−54%	99%
Lake sublimation ($m\ d^{-1}$)							
Lake Bonney	0.37	1%	−1%	10%	−9%	–	–
Lake Hoare	0.23	1%	−1%	8%	−8%	–	–
Lake Fryxell	0.23	1%	−1%	8%	−8%	–	–
Lake volume change ($m^3\ a^{-1}$)							
Lake Bonney	1.31E+07	40%	−35%	−48%	57%	−149%	297%
Lake Hoare	2.58E+06	35%	−32%	−50%	57%	−182%	303%
Lake Fryxell	5.45E+06	28%	−58%	−95%	69%	−241%	379%

Note. Values in table are expressed as the relative change ratio between adjustment conditions to normal conditions.

air temperature and wind speed, given the range selected but much more sensitive, by a factor of 8, to changes in albedo. Changes in air temperature had a much smaller effect on lake ice sublimation, whereas wind speed had a much greater effect. Clearly, changes in lake volume are more affected by changes in water inflow than sublimation due to changes in albedo of the glacier surfaces.

5. Analysis of Taylor Valley Lake Hydrology

Streams flowing over the valley floor (gaged and ungaged) supply the majority of inflow to Lakes Bonney and Fryxell, whereas Lake Hoare receives most of its inflow direct from Canada Glacier (Table 3). Subaqueous melt is only significant at Lake Hoare, accounting for 5% annual inflow. The inflow deficit (−450,000 $m^3\ yr^{-1}$) to Lake Fryxell accounts for 23% of the inflow. The mean annual estimated sublimation at Lakes Bonney, Hoare, and Fryxell were 0.37, 0.22, and 0.23 m yr^{-1}, accounting for 74%, 25%, and 35% of total lake ice ablation measured by Dugan et al. (2013), respectively (details of inflow contributions and lake ice sublimation and ablation are included in the Supporting Information S1).

The time series of lake hydrology can be characterized by two periods, before the summer of 2001/2002 and after. The first period is characterized by relatively cool air temperatures (mean December–January temperature −3.2°C), low stream discharge (Figure 3) and decreasing lake levels (Figure 5). During the summer of 2001/2002 a major foehn event in mid-December increased air temperatures to 4°C under cloudless skies, and the valley stayed warm for about 2 weeks (Doran et al., 2008; Foreman et al., 2004). During that time the glaciers melted rapidly and the glut of streamflow to the lakes more than made up for the sublimation losses of the lakes during the prior decade. In following years, summer air temperatures were slightly warmer (mean −2.6°C) and the albedo was considered to be 7% lower (this study, Hoffman et al., 2016), streamflow was greater and lake levels were rising. The warmer air temperatures and the lower glacier albedo were responsible for increased meltwater flow to the lakes.

During the cool period of the 1990s, when inflow did not compensate for sublimation losses, the magnitude of lake level decrease was about the same at each lake and sublimation of the lake ice dominated the water budget. This is consistent with the sensitivity analysis that showed similar responses of sublimation to changes in air temperature and wind speed. Since the summer of 2001/2002, lake levels have been rising due to the warmer mean summer air temperature and decreased albedo, but the rate of increase differs between lakes. Lake Bonney has risen much faster, by almost a factor of three, compared to Lakes Hoare and Fryxell. To explain this behavior,

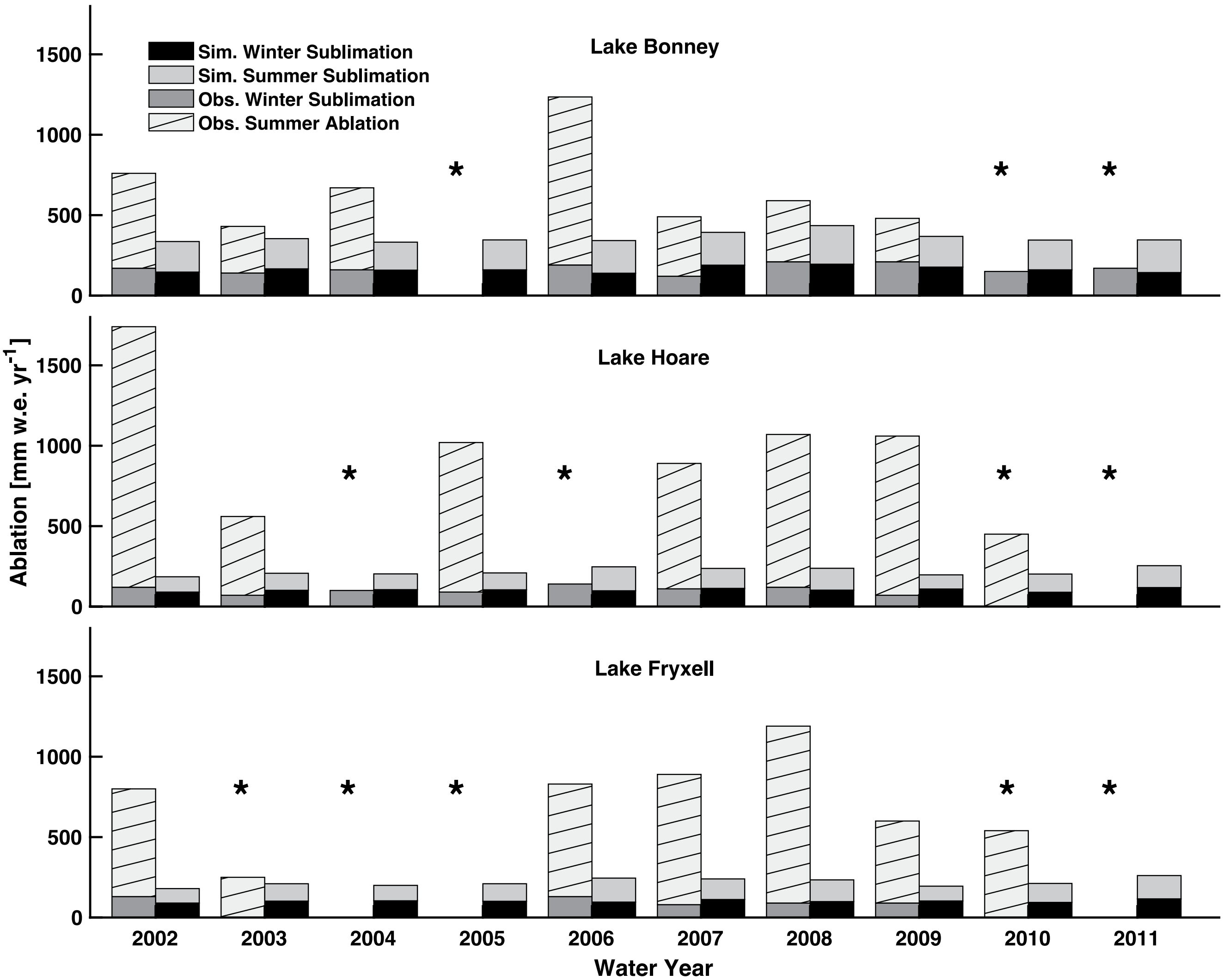

Figure 4. Sublimation model results compared to available measurements. Simulated annual sublimation, split into winter and summer, plotted with measured annual ablation, split into winter sublimation and summer ablation. Observed summer ablation (hatched area) includes both melt and sublimation. Observed winter values are measurements of sublimation only, as melt does not occur during the winter. All measured values are based on based on Dugan et al. (2013); an asterisks indicates a gap in the measurement record. The label on the *x*-axis indicates the ending year of the austral water year.

we first examined the effect of watershed hypsometry given that for the same increase in water volume, lakes in narrow valleys will rise more quickly than lakes in broad valleys. For each lake, the change in total water volume is calculated over the model period, 1996–2013 (Table 4). A hypothetical level change is then calculated for each lake assuming a constant lake area dating to 1996. The ratio of the observed level change to the hypothetical change yields the hypsometric effect such that ratios closer to 1 indicate more vertical walls and a greater sensitivity of lake level to volume change. Both Lakes Bonney and Hoare have ratios 0.97 and 0.98, respectively, whereas Lake Fryxell was half as large, 0.44. Therefore, for the same change in specific volume the level response at Lakes Bonney and Hoare would be twice that of Lake Fryxell.

Another factor is the specific melt, the total stream runoff divided by total contributing glacier area in each watershed. The inflow to Lakes Bonney and Fryxell is about the same but the contributing glacier area in the Lake Bonney watershed is about three times that of Lake Fryxell, indicating that the specific melt is smaller by a third. More ice is lost to sublimation per unit area compared to melt in Lake Bonney watershed than in either the Lake Fryxell or Lake Hoare watersheds due to the windier conditions (Hoffman et al., 2014). However, a small change in the energy balance due to warming air temperature or decreasing albedo, for example, can dramatically shift the balance from sublimation to melt (Hoffman et al., 2008, 2014). Therefore, the more rapid lake level rise

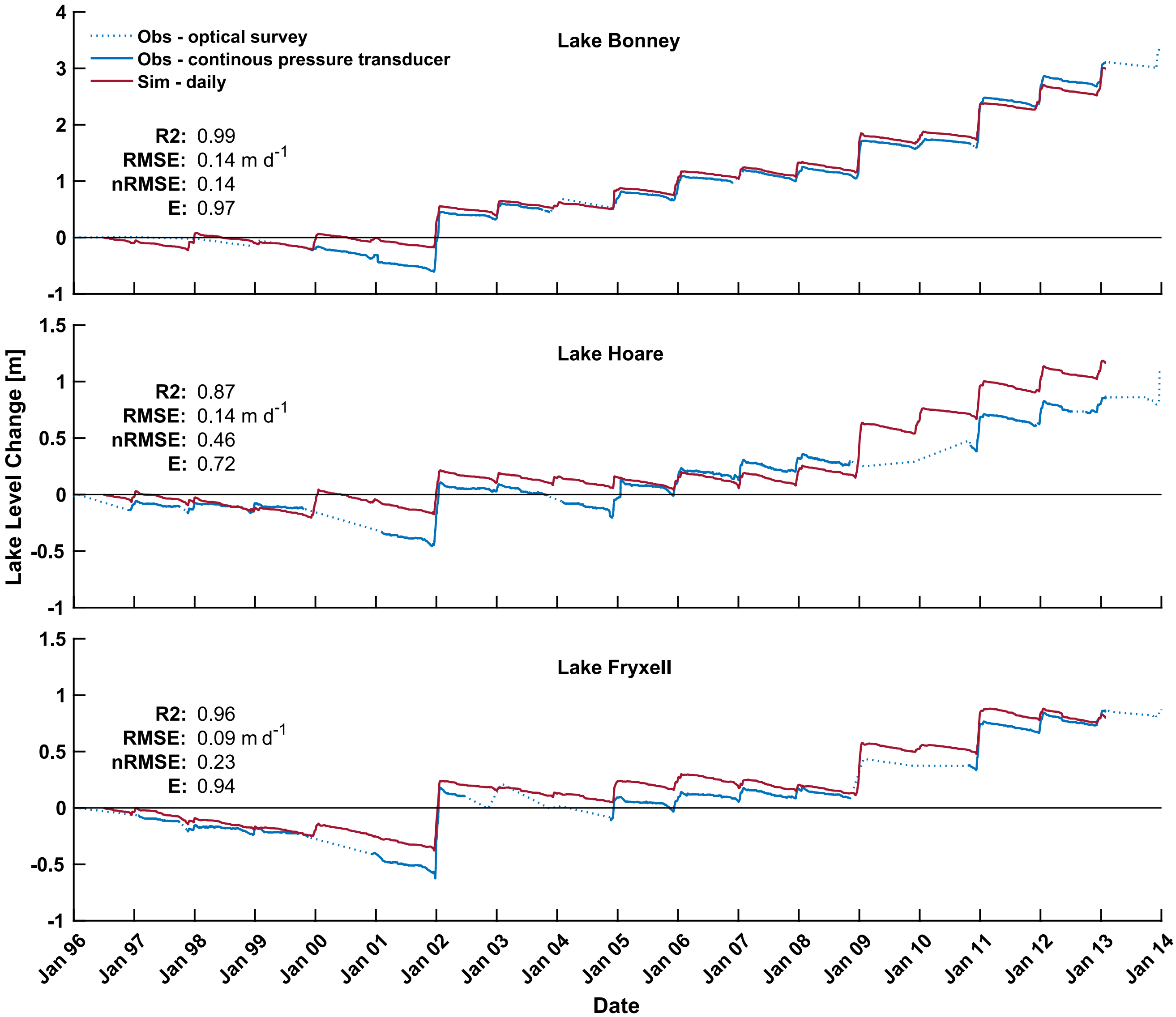

Figure 5. Simulated (red) daily lake level change, relative to the January 1996 datum, versus observed lake level change based on bi-annual optical surveys (thin dotted blue line) and continuous pressure transducer stage measurements (blue).

in Lake Bonney, compared to the other lakes, is caused by a combination of a larger ice-covered area contributing meltwater and a steep-walled watershed hypsometry that magnifies lake level response to changes in water volume.

As mentioned, the lake level change of Lakes Fryxell and Hoare are nearly identical, which is surprising considering differences in lake area, contributing glacier area, hypsometry, and response to weather forcing as shown by the sensitivity analysis. Their similar response is a coincidence in compensating effects of melt inflow and hypsometry. Lake Fryxell receives about four times the inflow as Lake Hoare, but its lake area is almost 4 times as large as Lake Hoare.

Year-to-year variability in lake level change is largely controlled by inflow variations due to changes in air temperature and magnitude of summer snowfall. The latter increases albedo of the ice surface, reducing melt. Within watersheds, the relative importance of various streams supplying meltwater to the lakes may also vary due to spatial patterns in snowfall. For example, in the Lake Fryxell watershed during cold summers (e.g., 1998, 2001), average temperature −4°C, streams flowing from Canada Glacier dominate the inflow (28% compared to 8% from Commonwealth Glacier), whereas during warm summers (e.g., 2002, 2009, and 2011), average temperature

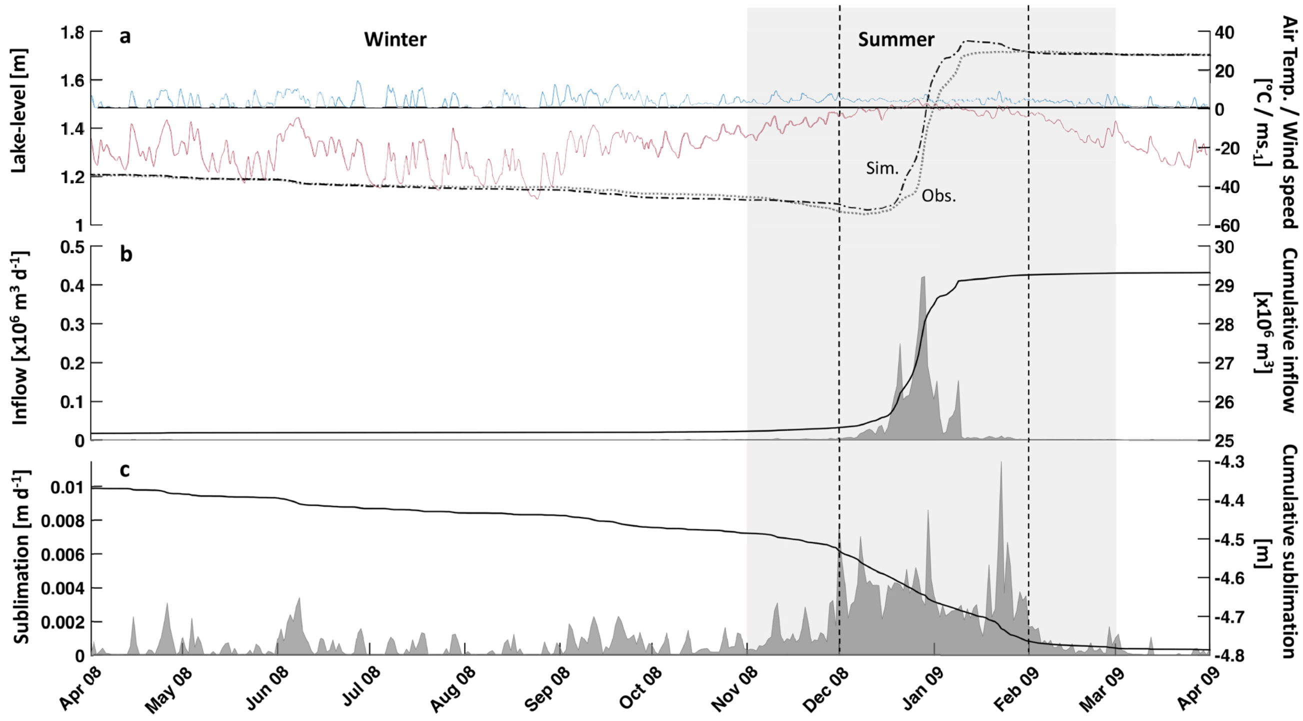

Figure 6. Simulated water year (April 2008–April 2009) at Lake Bonney. (a) Both the observed (dotted) and simulated (chain-dot) lake stage drop from April to December, and then rise rapidly until mid-January. Daily air temperature (red) and wind speed (blue) for the Lake Bonney meteorological station. (b) Daily (shaded) and cumulative (black solid line) predicted inflow into Lake Bonney. (c) Daily (shaded) and cumulative (black solid line) predicted sublimation of Lake Bonney ice-cover. Gray shaded box indicates the melt season in ICEMELT while dotted box indicates summer season with increased lake surface roughness in sublimation model.

−2°C, streams from Commonwealth Glacier dominate (21% compared to 18% from Canada Glacier). These effects are generally restricted to the Lake Fryxell and Lake Hoare watersheds because snow is less frequent in the Lake Bonney watershed (Fountain et al., 2010).

From the simple water balance model (1) lake level change can be projected through the end of the century assuming a steady state climate (Figure 7). Three estimates are made based on the entire study period (1996–2013) with an average summer (December-January) temperature of −2.7°C, the cool period prior to the 2001/2002 flood summer (1996–2001), summer temperatures, −3.2°C, and the warm period after the flood summer (2002–2013), summer air temperatures (−2.6°C) and an albedo lowered by −7% starting 2007. For each estimate, the streamflow and sublimation were modeled as before but the weather record over the period of interest was continuously cycled for 300 years. If the climate continues in the future as it has for the 1996–2013 period, all lakes will continue to rise with only Lake Fryxell reaching equilibrium by about 2200 (Figure 7 and Supporting Information S1). For persistent cool summers, lake levels decrease for all lakes and with all three lakes reaching equilibrium, Lake Bonney ~2250, Lake Hoare ~2150, and Lake Fryxell ~2250. For warm summers, the lakes continue to rise through 2300, reaching 93 m asl (+26 m from 2020 datum) at Lake Bonney, 86 m asl (+11 m) at Lake Hoare, and 24 m asl (+6 m) at Lake Fryxell. Under this warm summer scenario, only Lake Fryxell reaches close to equilibrium by 2300. In no scenario does a lake overflow into another lake watershed. The spill point between Lakes Fryxell and Hoare is 95 m, and between Lakes Hoare and Bonney is 156 m (Fountain et al., 2017).

Table 3
Sources of Inflow to the Three Major Lakes Based on Model Results

	Lake Bonney	Lake Hoare	Lake Fryxell
Gaged streamflow	45%	25%	64%
Ungaged streamflow	41%	6%	7%
Direct runoff	12%	65%	7%
Subaqueous melt	1%	5%	–
Inferred snow melt	–	–	23%

Note. Gaged streamflow is for the streams where the discharge is measured, using model data to fill in data gaps. Ungaged streams are unmeasured streams. Direct runoff is streamflow from the glacier surface and the subaerial portions of the glacier cliff face directly into the lake. Subaqueous melt is the ice loss from the subaqueous portions of the glacier cliff face abutting the lake.

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Table 4
Lake Level Sensitivity to Watershed Hypsometry

	Inflow	Loss	ΔV	A_L	A_C	Δh	Δz	$\Delta z{:}\Delta h$	SpMelt
Lake Bonney	44.4	29.8	14.6	4.4	52.7	3.28	3.22	0.98	0.84
Lake Hoare	11.5	9.0	2.4	2.2	5.2	1.10	1.07	0.97	2.21
Lake Fryxell	44.0	29.8	14.1	7.2	16.8	1.97	0.86	0.44	2.62

Note. Inflow (I $\times 10^6$ m^3), losses (L $\times 10^6$ m^3) and net volume change ($\Delta V \times 10^6$ m^3), are 1996–2013 totals. Lake area ($A_L \times 10^6$ m^2) is from 1996. A_c ($\times 10^6$ m^2) is the glacier area in each watershed that contributes melt to the lakes. The net volume change divided by initial lake area, $\Delta V/A_L$, yields the height change, Δh, ignoring watershed hypsometry. The observed height change, Δz, over that period divided by the hypothetical height change defines the hypsometric effect. Specific melt (SpMelt) is the Inflow divided by A_c.

6. Discussion

6.1. Taylor Valley Hydrologic System

Lake levels of the polar lakes in Taylor Valley are quite sensitive to climatic conditions, like closed watershed lakes globally (Street-Perrott & Harrison, 1985). That the lakes have a perennial ice cover has little bearing on their apparent sensitivity except the hysteresis between wet and dry years may be less pronounced than for

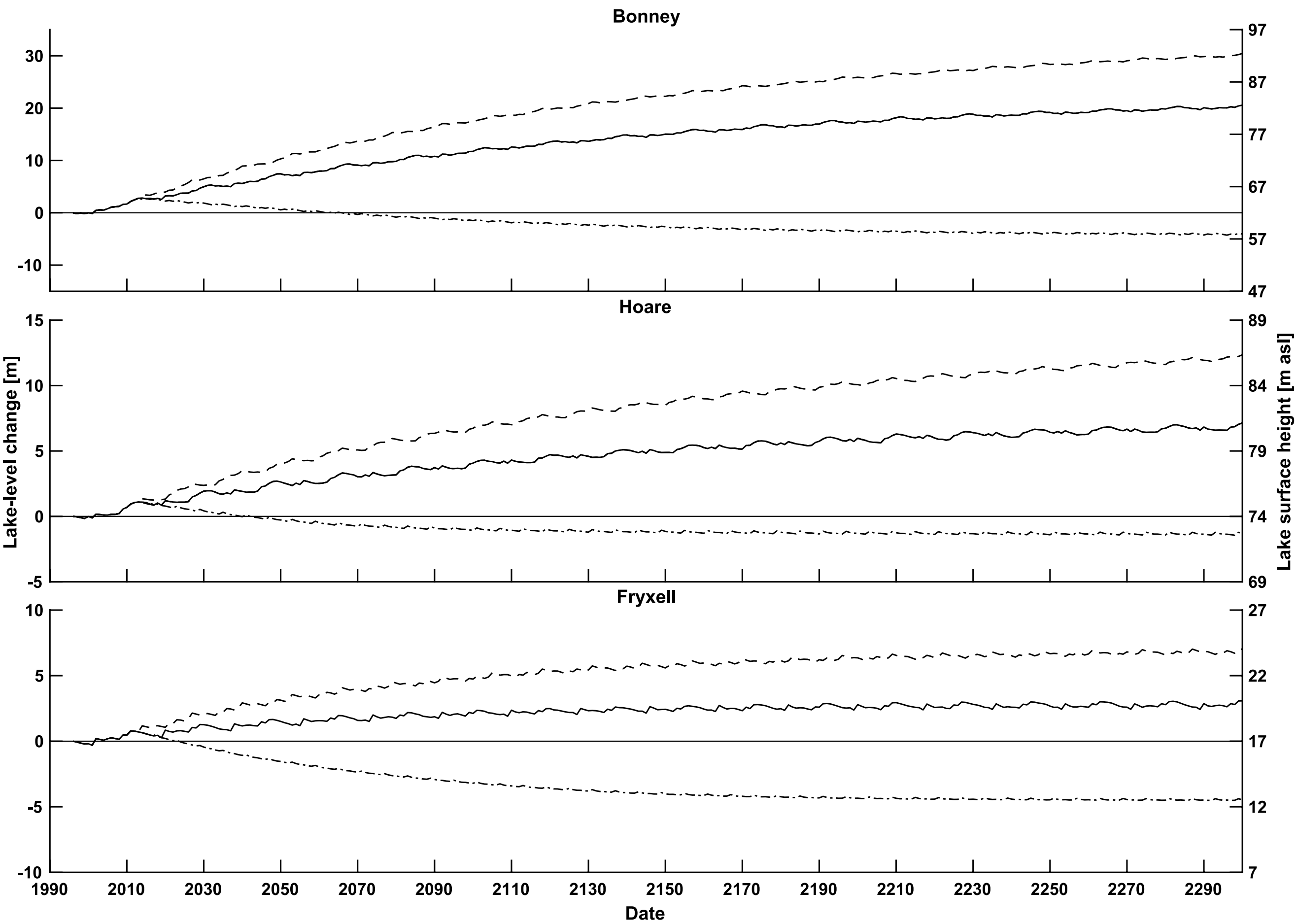

Figure 7. Projected lake levels 1996 to 2300 for Lakes Bonney, Hoare, and Fryxell. Lake levels were simulated based on mean conditions through the entire study period (1996–2013; solid), the cool period prior to the 2001/2002 flood summer (1996–2001; chain-dot), and the warm period after the flood summer (2003–2013; dashed). Lake level is expressed as change relative to the 1996 datum on the left and with orthometric surface height on the right.

temperate lakes. Losses from sublimation are +13% (at 0°C) more energetically expensive compared to evaporation and therefore require more time to lose the equivalent specific volume of water.

The surface level of temperate lakes can dramatically increase due to precipitation events via direct rainfall onto the lake and by rainfall-runoff from the surrounding watershed (Winter, 2003). The process is entirely different in the MDV. Direct precipitation, which falls as snow, adds little to lake level because the largest snowfall events are typically small (<0.1 m) and are low density (~100 kg m^{-3}; Fountain et al., 2010). More importantly, runoff from the glacial melt in the watershed is almost entirely suppressed during, and a few days after, a snowfall event. A thin snow cover on the ice-exposed zones of the glaciers dramatically increases albedo, and given the near-freezing summer air temperatures, the energy balance is insufficient to melt the snow resulting in the termination of melt-runoff from the glaciers except for the small discharge derived from the vertical ice cliffs that terminate at many of the glaciers (Chinn, 1993; Fountain et al., 1998; Hoffman et al., 2016). A relatively thin snow cover of a few centimeters can suppress runoff for weeks. Large runoff events are associated with warm air temperatures and clear skies (Conovitz et al., 2013; Doran et al., 2008; Hoffman et al., 2016). Therefore, the hydrology of the valleys is energy, rather than precipitation, limited. Should the region warm in future, it will become wetter with no change in precipitation required.

The hydrological model worked well predicting seasonal and annual lake level changes. Predicting total streamflow to each lake showed correlation coefficients between 0.75 and 0.83 with Nash-Sutcliff efficiencies of 0.71–0.83. Predicted lake levels matched measured values typically within 0.10 m, although larger excursions occurred occasionally. Sensitivity tests showed that meltwater runoff and lake level response are highly sensitive to differences in air temperature, albedo, and wind speed, all of which control the energy balance of the ice surfaces. These results support prior studies on components of the hydrologic system in this region (Chinn, 1993; Clow et al., 1988; Dugan et al., 2013; Lewis et al., 1999). Attempts to account for water losses from the streams to the atmosphere failed to improve the model results suggesting that the losses associated within channel processes are small relative to the larger uncertainties elsewhere in the system (discussed below).

Modeled lake level change shows that glacial meltwater streams account for the majority of water to Lakes Fryxell and Bonney, whereas direct runoff from the glacier is the major source to Lake Hoare. Subaqueous melt, which also includes direct calving of ice to the lakes is a minor source to Lake Hoare (5%) and less so for Lake Bonney (1%). Estimated sublimation losses from the lake ice surface are greatest for Lake Bonney, 0.37 m yr^{-1}, and essentially the same at Lakes Hoare and Fryxell, 0.22 m yr^{-1} and 0.23 m yr^{-1}, respectively.

Relative lake level change between lakes is partly due to mass balance and partly due to hypsometry. Lakes in broad shallow watersheds (Lake Fryxell) change less due to a difference in mass budget than a lake in a steep-walled valley (Lake Bonney). Indeed, the effect of hypsometry in the Lake Fryxell watershed is half that in the Lake Hoare and Lake Bonney watersheds. The sensitivity tests show that an increase in average air temperature by 10% (~0.28°C) increases meltwater flow in the Lake Bonney watershed by ~200% over the Lake Fryxell watershed and ~125% in the Lake Hoare watershed. This is due to the energy balance in the Lake Bonney watershed favoring sublimation (higher winds) and any increase in energy due to an increase in sensible heat (air temperature) or solar radiation (decreased albedo) or a decrease in sensible/latent heat losses (decreased wind), more dramatically increase melt flux. The meterological sensitivity in combination with the effect of hypsometry create greater and more rapid lake level changes in Lake Bonney than the other two lakes. That the lake level changes of Lakes Fryxell and Hoare are nearly identical is the coincidental result of compensating effects of volume of meltwater flow to lake area, and the difference in hypsometry of the two watersheds.

6.2. Processes Requiring Further Study

The model also highlighted processes requiring further study to improve our understanding of hydrology of this polar desert. Required adjustments of the near-surface air temperature lapse rate in the model suggests that the atmosphere is more humid within a few km of the coast compared to the known dry conditions farther up valley (Doran, McKay, et al., 2002; Fountain et al., 2014; Hoffman et al., 2016). This notion is supported by frequent field observations of clouds drifting into the region along the mountains on the south side of the valley. Considering sea ice with a below freezing (dry) surface covers the adjacent ocean, this adjustment was unexpected.

Another issue is the albedo of the ice-exposed (melting) glacier surface. We know from anecdotal experience and modeling outcomes that the magnitude of ice melt is very sensitive to albedo. Indeed, snow accumulation increases the albedo of glacier ice substantially and largely eliminates glacial melt, except on the vertical ice cliffs (Hoffman et al., 2016; Lewis et al., 1998, 1999). The sensitivity analysis of the model shows the most change in melt and lake levels when albedo is altered. Improved model performance would be achieved by defining spatial patterns of albedo. MODIS albedo has been very helpful in this regard but it is subject to missing data due to cloud cover and by coarse spatial resolution (500-m).

Sublimation is assumed to be the only major mass loss from the lakes, yet limited field data exist (Clow et al., 1988; Dugan et al., 2013). Modeling, consistent with anecdotal observations, suggests smaller roughness in winter and larger in summer. To better test the model, temporal field measurements of sublimation and lake ice roughness are necessary.

6.3. Differences at Lake Fryxell

The model did not close the hydrological budget in the Lake Fryxell watershed, despite good agreement with all measured streams and reasonable sublimation estimates from the lake ice. To close the budget either the roughness of the lake ice needed to be reduced or another source of water was required. A reduction in roughness required an unreasonably smooth value, 0.05 mm, two orders of magnitude smaller than required for either of the other two lakes and counter to field observations, which strongly indicate similar roughness between lakes. Using a more realistic roughness of 7 mm required another source of water averaging 450,000 $m^3\ yr^{-1}$ ($\sim$1,200 $m^3\ d^{-1}$) to close the hydrologic budget. One possible source is a deep saline aquifer detected under the lake (Mikucki et al., 2015). However, no known surface or subaqueous seeps of the discharge required are known. Also, the water is thought to be saline and would have created a much deeper saline pool in the bottom of the lake than what at present is known to exist. That pool has been shown to be a result of a paleo-lake desiccation event based on geochemical evidence (Hendy, 2000; Lyons et al., 1998). Furthermore, if there was a hydrologic connection between Lakes Hoare and Fryxell as hypothesized by Mikucki et al. (2015), water would flow from Lakes Hoare to Fryxell, requiring the model to supply extra water to Lake Hoare to close its budget and require no or little water to close the budget of Lake Fryxell. Alternatively, we hypothesize that a subsurface flux of melt water exists, probably conveyed through water tracks over the ice-cemented permafrost, and sourced from the distributed snow patches commonly found in this watershed. The Lake Fryxell watershed is closest to the coast and receives the most snow in the valley making snow patches relatively common compared to the other watersheds (Eveland et al., 2013; Fountain et al., 2010).

Water tracks and water patches are common in the Lake Fryxell watershed suggesting an active shallow groundwater system (Levy et al., 2011, 2012; Lyons et al., 2005). Although the flux from the largest known water track is only about 30 $m^3\ yr^{-1}$ (Levy et al., 2011), other undetected water tracks probably exist. A simple mass balance calculation indicates that only 2 mm w.e. snow accumulation per year over Lake Fryxell watershed in the valley bottom would be required to compensate for the deficit in lake budget. The estimated deficit is quite large, and the 7 mm roughness value may be an over estimate, reducing the roughness would reduce the deficit. No matter its final value, any reasonable estimate of roughness results in a deficit that requires resolution. Our hypothesis of a wide-spread shallow groundwater system within the active layer (e.g., O'Connor et al., 2019) challenges the accepted view of the hydrological processes in the Lake Fryxell watershed. One would expect that other watersheds, with similar climatic characteristics to Lake Fryxell, would also have a similar shallow groundwater systems and together have important implications for soil ecosystem structure and function and for lake chemistries (Barrett et al., 2006; Levy et al., 2011, 2014; Lyons et al., 1999).

6.4. Climate Sensitivity of the Taylor Valley Lakes

Simple mass conservation modeling shows that if conditions over the 1996–2013 period were to continue in the future, lake levels will continue to rise, with only Lake Fryxell reaching equilibrium by about 2300. That lake levels would continue to rise under these conditions shows that the lakes are out of equilibrium with the current environment and have been since measurements began in the 1970s (Chinn, 1993; Gooseff et al., 2011). As shown, lake levels are very sensitive to the energy balance as characterized by air temperature, wind speed, and glacier albedo. For example, if the cool period (1996–2001) is used to infer future climate the future climate lake levels decrease to an equilibrium level of 3–5 m lower and if the warm period is used, lake levels increase by 4–13 m. The air temperature difference between the cool and warm periods is only +0.6°C underscoring the extreme temperature sensitivity of the hydrologic system in this polar desert generally. Changes in lake level are not a recent phenomenon. Based on a measurement of lake width made by R.F. Scott in 1903 (Scott, 1905), the level of Lake Bonney was inferred to be 50 m (Chinn, 1993) and by 1990, the lake level had increased about 12 m, or about +0.14 m yr^{-1} (Bomblies et al., 2001). This is less than the current lake level rise over the study period of +0.17 m yr^{-1}. To explain that lake level rise, and considering air temperature alone, suggests an air temperature between −3.2 and −2.8°C lower than our cool period and overall average of our study period.

Under current conditions lake levels will not reach the spill points between lakes and they will remain separated. However, a number of fixed camps, study sites, and weather stations in the valleys may be flooded in the coming decades. Rising lake levels will not destabilize glaciers in contact with the lakes through increased buoyancy except possibly the lower tip of Taylor Glacier, which flows into Lake Bonney. The large magnitude of lake rise, relatively thin glacier terminus, and somewhat flat terrain around the terminus, makes the terminus vulnerable to possible flotation. Taylor Glacier has an intermittent hypersaline flow sourced from subglacial waters farther up-glacier (Carmichael et al., 2012; Mikucki et al., 2009), which, if submerged in rising lake water, could pressurize the glacier bed farther up-glacier. However, the glacier is reasonably thick where the subglacial waters are thought to be located and the increased pressure of a few tens of meters of water would likely not affect glacial flow.

7. Conclusions

The hydrological balance of the lakes in Taylor Valley, Antarctica, is simply governed by the flux of glacial meltwater and sublimation from the lake ice, neither precipitation nor instream evaporative losses play a significant role. In the Lake Bonney and Lake Hoare watersheds we infer no significant water gain or loss between the lakes and the shallow groundwater system that flows over the ice-cemented permafrost. For Lake Fryxell, we hypothesize that such groundwater is a significant component to Lake Fryxell sourced from the snow patches that dot the watershed. Glacial meltwater production and subsequent lake level response are very sensitive to small changes in the energy balance of the glaciers governed by air temperature, wind speed, and albedo. Because the energy balance of the glaciers hover near the melt threshold any increase in energy due to increased heat gain to the glaciers or a reduction in turbulent heat loss to the atmosphere, shifts the energy balance from the sublimation dominated mass loss to one with increasing melt causing a non-linear response in melt-runoff. Lake response to changing energy balance differs between lakes partly due to differences in the energy balance conditions in each watershed, the area of glacial ice, and watershed hypsometry. For these reasons the level in Lake Bonney is more responsive to the changing weather and climate conditions than the other lakes. All lakes are currently rising, and have done so at least since the 1970s, and perhaps as far back as 1903, suggesting the lakes are out of balance with the current climate. Assuming the weather conditions during the 17-year study period persist into the future, lakes levels will reach equilibrium starting in 2300.

Data Availability Statement

Meteorological data, glacier mass balance, streamflow, and lake level data used as input are available on the MCM LTER database (https://mcm.lternet.edu/) and appear as references in this manuscript (Doran & Fountain, 2019; Doran & Gooseff, 2018; Gooseff & Fountain, 2022; Gooseff & McKnight, 2016). The MCD43A3 V006 MODIS daily albedo product is available at https://doi.org/10.5067/MODIS/MCD43A3.006 (Schaaf & Wang, 2015), and was accessed via Google Earth Engine using scripts that are developed openly at https://github.com/geo-jules-cross/gee-albedo-extract. The input data and metadata for the MICROMET code used to distribute meteorological parameters in Taylor Valley are available at https://doi.org/10.6073/pasta/f5bc1d4c5925f389ee8220bbec823e8a (Hoffman, 2016); by request of the original author, the full code will be made available by request. Large meteorological input data files for the ICEMELT model are available at https://doi.org/10.5281/zenodo.6808785 (Cross, 2022). The adapted ICEMELT model code used for estimating Taylor Valley meltwater is preserved at https://doi.org/10.5281/zenodo.6808770 (Cross & Hoffman, 2022), available via MIT license and developed openly at https://github.com/geo-jules-cross/icemelt-cross. The lake model code and input data used in this study to estimate lake sublimation and simulate lake water balance are preserved at https://doi.org/10.5281/zenodo.7110777 (Cross & Obryk, 2022), available via MIT License and developed openly at https://github.com/geo-jules-cross/lake-model-daily.

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