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Supporting Information for

Proglacial lakes control glacier geometry and behavior during recession

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Introduction

This supporting information comprises additional details of the datasets used in this study and their analysis. Text S1 provides additional information on the model initial conditions, including LGM bed topography and model parameters, supported by Figure S1 and Table S1. Text S2 outlines the procedure and results (Figure S2) for the model spin up. Text S3 discusses the experimental design of the model simulations, supported by Table S3, including how the idealised climate was prescribed and how the LAKE simulation was initiated. An extended description of the sensitivity analysis is given in Text S4 and Table S2, the results of which are given in Figure S3. Figures S4-S6 are included to provide additional illustration to the results. Movie S1 and S2 are the full retreat simulations for LAND and LAKE respectively. Movie S3 presents the LAKE simulation where ice thickness has been inverted to clearly show the effect of the lake on the ice front.
Text S1. Model initial conditions

S1.1 Bed topography

Before beginning the numerical modelling, we needed to consider the Last Glacial Maximum (LGM) bed conditions. We cannot assume that contemporary topography is an appropriate representation of the glacier bed beneath the Pukaki Glacier. The Pukaki Glacier occupied an area where Lake Pukaki now exists, for which standard digital elevation models (DEMs) represent the water surface and not the underlying bathymetry, thus obscuring the former glacier bed topography. To produce a DEM more representative of LGM conditions, the water depth of Lake Pukaki (Irwin, 1970) was subtracted from the modern DEM. The lake is only 98 m deep at its deepest, which is relatively shallow when compared to many other lakes in New Zealand that have beds below sea level (e.g. Lake Wakatipu, >300 m deep; Sutherland et al., 2019).

Geophysical data from Lake Ohau, in an adjacent valley to Lake Pukaki, indicate that the lake basin there contains up to 140 m of sediment deposited directly on top of bedrock (Levy et al., 2018). Sediment cores collected from Lake Ohau reveal that these sediments have accumulated since lake formation at the end of the LGM ~ 17 ka (Levy et al., 2018). Seismic surveys subparallel to the Pukaki basin provide estimates of the subsurface locations of bedrock also presently covered by thick layers of sediment (Kleffman et al., 1998; Long et al., 2003). Based on these studies, we contend that substantial sediment deposition has occurred in the Pukaki valley since the LGM and it is likely that the Pukaki Glacier bed is buried by Late Glacial and post-glacial sediments. Indeed, McKinnon et al. (2012) estimated post-LGM sediment thickness distribution (and area extent) within the Pukaki valley, revealing up to 384 m of post-LGM infill. Therefore, in this study, we used these estimates of bedrock elevation as a constraint on the modelled bed profile for the LGM. These bed elevation data were subsequently merged with the modern DEM to produce a surface that we suppose is more representative of LGM bed conditions (Figure S1) than simply using a modern DEM. It is highly unusual to know the bathymetry of a proglacial lake, especially one formed from the LGM, and this knowledge (in addition to the cosmogenic nuclide dating of the moraines that encircle Lake Pukaki) is further justification for our choice of site for this study.

S1.2 Model domain and parameters

We set up our model domain to cover 64 km by 128 km. For computational efficiency during the spin up simulation (Text S2) we set a 500 m x 500 m grid
refined three times around the grounding line of Pukaki to produce a maximum
horizontal resolution of 125 m (Figure S2). For the LAKE and LAND simulations we
used a 250 m by 250 m horizontal grid resolution across the entire model domain.
The simulations have 10 vertical levels. Ice surface temperature was held constant at
an isothermal value of 268 K in all simulations (Table S1).

Text S2. Model spin up

Numerical modelling results can be heavily influenced by the starting condition. In
this case it was imperative that the starting condition of ice thickness was that of a
glacier in equilibrium, to be sure that subsequent glacier changes were only a
product of a modelled forcing. In this study, ice extent in the spin up model run was
controlled by the surface mass balance (SMB). We imposed the initial SMB by the
following equation, where the Equilibrium Line Altitude (ELA) prescribed was 1465 m;

\[ \text{SMB} = (\text{surface elevation} - \text{ELA}) \times 0.0025 \]

This allowed the Pukaki Glacier to advance to its LGM position (Figure S2),
comparable to empirical reconstructions (e.g. Barrell et al., 2011), and other modelled
ice thicknesses of the Pukaki Glacier (e.g. Golledge et al., 2012; James et al., 2019).
The ice thickness at the end of this spin up simulation was used as the initial
condition for LAND and LAKE simulations which began at a stable ice volume with ice
grounded on the topography ~ 2 km down-valley from the lip of its over-deepened
basin.

Text S3. Model Experimental design

S3.1 Parameterisation of climate

The motivation for this study was to assess the impact of a proglacial lake on ice
dynamics, not to produce more realistic glacier changes or an absolute chronology
of events. Based on an accumulation area ratio (AAR) analysis of reconstructed
glaciers at the LGM, Porter (1975) estimated that the LGM ELA was 1225 m and that
the Late Glacial ELA inside the Birch Hill moraine limit was 500 ± 50 m lower than the
modern day (2100 m; Chinn et al., 2012). The ELA for the Birch Hill re-advance is
therefore taken to be 1600 m. The way in which we prescribed a steadily warming
climate is given as follows;

\[ \text{ELA} = 1465 + (10000 \times 0.05) \]
where 1465 is the initial starting condition, 10,000 is the length of the model run, and 0.05 represents the rate of ELA rise.

S3.2 Initiating the ‘LAKE’ simulation

Given that the present-day surface of Lake Pukaki lies at an elevation of 525 m a.s.l. we reset the elevations relative to the lake level. we lowered the topography of the whole model domain by 525 m to effectively bring the lake surface down to sea level (0 m a.s.l) in order to initiate the LAKE simulation. This method enabled us to then prescribe calving and subaqueous melt fluxes since BISICLES can only simulate such processes when the glacier margin is at zero elevation. The ELA in the LAKE simulation was also lowered by 525 m to account for the topographic lowering. Therefore, we take the ELA (initial starting condition) in the LAKE simulations to be 925 m.

Our choice of ice sheet model was based on accounting for grounding line dynamics induced by proglacial lakes and the issue in how we applied the model to simulate an inland lake is not a concern for the process-representation. Lacustrine termini are thought to experience fewer perturbations (e.g. tidal flexure, high subaqueous melt rates) and are therefore inherently more stable than tidewater termini. Water circulation in a proglacial water body determines when and how heat reaches glacial ice and affects melting. In marine-terminating environments, relatively warm ocean water can be drawn towards an ice margin by water circulation patterns caused by the relative buoyancy of that freshwater within the saline water. Water circulation in marine settings can be driven by density differences (Farmer and Freeland, 1983; Motyka et al., 2003), tides (Mortensen et al., 2011) or winds (Straneo et al., 2010), but since the water of a proglacial lake is fresh, such a circulation will not arise. All heating and cooling processes in a lake are therefore local and take place in a closed system, as opposed to marine environments where heat can be transported long distance from the ocean, which effectively acts as an infinite reservoir of heat (Truffer and Motyka, 2016).

Near-terminus surface slopes of tidewater glaciers are typically steeper than lake-calving termini, resulting in near-terminus ice speeds differing by an order of magnitude. Retreating tidewater glaciers often flow at speeds of 5-10 km a\(^{-1}\), compared to 100-1000 m a\(^{-1}\) for lake-terminating glaciers (Truffer and Motyka, 2016). Grounded tidewater termini are typically highly crevassed and characterized by steep topography, fast flow, high strain rates and frequent calving activity. In contrast, many lake-calving glaciers form floating tongues that are characterized by lower
surface gradients, flatter topography, slower flow, lower strain rates, less crevassing, and infrequent but massive calving activity, often in the form of large tabular blocks. Such is the case for lacustrine terminating glaciers in Alaska, New Zealand and Chilean Patagonia. However, this distinction does not hold universally, departures from this model are the large east Patagonian lakes, where they are exposed to relatively intense solar heating, and lake water temperatures can exceed 4 °C in summer (Truffer and Motyka, 2016). Subglacial discharge could be buoyant in such water, although density difference is significantly smaller than that between fresh and saline waters. The onset of circulation, together with the thermal forcing from entrained warm water, would lead to moderate rates of subaqueous melting that are below those observed at temperate tidewater glaciers, but significantly above those observed at smaller lakes. The largest of these lakes, Lago Argentino, lying at the terminus of Glaciar Upsala, has a maximum water depth at the grounding line of ~500 m, deep enough to allow part of the glacier tongue to float. Lago Argentino could serve as an analogue for Lake Pukaki in terms of energy balance, temperature and water circulations. However, the surface area of Lago Argentino is ~1400 km², an order of magnitude higher than that of Lake Pukaki (~180 km²). It is also noteworthy that glaciers that exit into the east Patagonia lakes (e.g. Upsala, Perito Moreno, and Viedma) are generally not afloat and do not calve tabular icebergs, while those in Alaska and Chilean Patagonia do. Generally, the east Patagonian lake-calving glaciers appear to have more in common with tidewater glaciers in terms of glacier speeds and terminus morphology (Venteris, 1999; Stuefer et al., 2007; Sakakibara and Sugiyama, 2014).

**Text S4. Sensitivity analysis**

A calving rate and a subaqueous melt rate needed to be calculated and prescribed in the LAKE simulations. Calving and subaqueous melt research is disproportionately focused towards tidewater glacier margins (Van der Veen, 2002; Benn et al., 2007). A sparsity of quantitative data means that these processes and their associated drivers at lacustrine ice-margins remain poorly understood (Purdie et al., 2016). Therefore, constraining rates based on present-day observations of exiting proglacial lakes is difficult. Table S2 shows highly variable calving and melt rates in proglacial lakes depending on many factors such as location and water depth.

Based on the differences and assumptions described above, we aimed to test the sensitivity of the model to (i) different types of calving model, (ii) the calving rate and (iii) the subaqueous melt rate (Table S3). The sensitivity simulations were run at a horizontal model resolution of 500 m for 4,000 years (from 18 ka to 14 ka). This
enabled enough time to force the terminus into the lake and well back through the
over-deepening in order to assess changes in model output.

\textit{S4.1 Calving rate}

Existing data from modern glaciers consistently show that calving occurs much more
slowly in lakes than in comparable tidewater settings. Lacustrine calving rates are
typically an order of magnitude lower than that of tidewater termini (Funk and
Rothlisberger, 1989; Warren et al., 1995; Warren and Aniya, 1999; Benn et al., 2007;
Truffer and Motyka, 2016). Such differences have been attributed to contrasts in
water densities, upwelling rates (and associated turbulent heat transfer), subaqueous
melt rates, frontal oversteepening and longitudinal strain rates (Funk and
Röthlisberger, 1989; Warren et al., 1995; Van der Veen, 2002). Warren and Kirkbride
(2003) confirm that calving correlates linearly with water depth in freshwater.

There is a strong contrast in calving mechanisms and rates between tidewater and
freshwater settings (Warren and Kirkbride, 2003). Thermal undercutting and
buoyancy-driven ice calving are the primary controls of retreat in most lakes.

Thermo-erosional notches in the calving front of glaciers that terminate in lakes may
be formed when rates of melting at the waterline are higher than subaerial or
subaqueous rates of melting. They have been observed at a variety of lake-calving
glaciers in New Zealand (Warren and Kirkbride, 2003; Röhl, 2006; Dykes et al., 2010),
Alaska (Trussel et al., 2013), Patagonia (Truffer and Motyka, 2016), and east
Greenland (Mallalieu et al., 2020).

\textit{S4.2 Subaqueous melt rate}

Many lacustrine subaqueous melt rates reported in the literature are conceptual or
have been reported from supraglacial lakes and ice cliffs, albeit a similar process but
on a much smaller scale. Several different methods have been applied to model
subaqueous melt rates and as such, their measured units vary from mm hr$^{-1}$ to m a$^{-1}$.
Some report a calving flux (e.g. m$^3$) whilst others report a calving rate (e.g. m a$^{-1}$).
Subaqueous melting is the least well-constrained term, however, could account for
significant portions of total ice retreat. The formulas for subaqueous melt rates in
numerical models, that are mainly derived from experiments and match inferred rates
from studies of Antarctic icebergs, apply to clean ice. The submerged faces of a
 glacier terminating in a lake are likely to be covered to varying degrees with
sediments. Besides other minor factors melt rates have been shown to decrease with
increasing water pressure at depth. The influence of water pressure is significant for
melting processes in ice-contact lakes as water depths often exceed 100 m.
Our sensitivity testing (Figure S3) revealed that varying the subaqueous melt rate produced morphological differences, such as the existence or absence of floating ice tongues. Subaqueous melting had a negligible effect on grounding line position in our model but had a significant effect on terminus position.

Where the combination of ice thicknesses and water depth satisfies the flotation criterion within the model, an ice shelf is formed. Low subaqueous melt rates (e.g. 0 - 10 m a\(^{-1}\)) result in a configuration where a large ice shelf was permitted to form during retreat with only a narrow band of exposed water. In contrast, when high or extreme subaqueous melting was prescribed (e.g. 100 m a\(^{-1}\)), the resulting configuration forced the removal of the ice shelf with little or no floating ice during retreat and a relatively large area of exposed water. The subaqueous melt rate therefore determines how much floating ice is present. Subaqueous melt drives retreat of terminus position, however, if the floating ice has a slightly larger extent, the impact on grounded ice extent, volume and velocity is still negligible. Changing the calving rate was also found to have a negligible effect on the overall pattern of retreat (Figure S3). This is because the ice terminus was wedge-shaped (when a melt rate >10 m a\(^{-1}\) was prescribed) and so the calving rate had little impact. Calving at the ice front plays only a minor role and our experiments are weakly sensitive to its representation in the model. Calving has much less control on grounding line retreat. As a result, a distinct calving model for lakes will not have any impact in our experiments. We acknowledge that this might not necessarily always be the case for different time intervals or settings (e.g. a colder climate). Most importantly, we show that both changing the calving or subaqueous melt rates have a negligible impact on grounding line position.
Figure S1. LGM bed profile. Profile taken along X-Y transect in Figure 3a.
Figure S2. Initial ice thickness and extent from the equilibrium spin up LGM ice simulation at 18 ka (a description of which is given in Text S2). Inset shows the evolution of ice volume and area during the spin-up simulation. Horizontal model resolution at the ice margin is 125 m with 3 levels of refinement.
Figure S3. Effects of changing subaqueous melt and calving rate on grounding line and terminus position, ice thickness, and velocity. Parameters and values reported in Table S3.
Figure S4. Model domain gridded into areas of open land (green), open water (dark blue), Grounded ice (red), and floating ice (light blue) for LAND and LAKE with -50 m a^{-1} subaqueous melt rate prescribed, and LAKE with 0 m a^{-1} subaqueous melt rate prescribed. Plotted every 500 years from 17.5 ka to 12 ka. Note difference in terminus position and extent of floating ice between both LAKE simulations, however, grounding line position remains the same.
Figure S5. Ice thickness maps for LAND and LAKE with -50 m a\(^{-1}\) subaqueous melt rate prescribed, and LAKE with 0 m a\(^{-1}\) subaqueous melt rate prescribed. Plotted every 500 years from 17.5 ka to 12 ka. Note difference in terminus position between both LAKE simulations but grounding line thickness remains the same.
Figure S6. Ice velocity maps for LAND and LAKE with -50 m a\(^{-1}\) subaqueous melt rate prescribed, and LAKE with 0 m a\(^{-1}\) subaqueous melt rate prescribed. Plotted every 500 years from 17.5 ka to 12 ka. Note difference in terminus position between both LAKE simulations, but ice velocity over the grounding line remains the same.
| Parameter                        | Value | Units   |
|---------------------------------|-------|---------|
| Sliding exponent                | 1     |         |
| Isothermal ice temperature      | 268   | K       |
| Ice density                     | 918   | Kg m\(^{-3}\) |
| Water density (freshwater)      | 1000  | kg/m\(^{-3}\) |
| Domain length                   | 128   | Km      |
| Domain width                    | 64    | Km      |
| Maximum refinement              | 0.25  | Km      |

**Table S1.** Key model parameters
| Location     | Glacier name | Calving Rate (m a⁻¹) | Subaqueous melt rate (m a⁻¹) | Reference                  | Explanatory notes                                                                 |
|--------------|--------------|-----------------------|------------------------------|----------------------------|-----------------------------------------------------------------------------------|
| New Zealand  | Maud         | 88                    | 18                           | Warwick and Kirkbride (2003)| Width and annually averaged measurements taken in Autumn 1994-95                  |
|              | Grey         | 47                    | 18                           |                            | Temperate                                                                         |
|              | Godley       | 47                    | 18                           |                            | Grounded                                                                          |
|              | Hooker       | 14                    | 18                           |                            | Debris-covered                                                                    |
|              | Ruth         | 18                    | 18                           |                            | Largely un-crevassed                                                              |
|              |              |                       |                              |                            | Calving face typically 20-40 m                                                   |
|              |              |                       |                              |                            | Grounded in shallow water (<20 m)                                                |
|              | Tasman       | 34                    |                              | Röhl (2006)                | Measurements taken between 2001 and 2003                                           |
|              |              |                       |                              |                            | Maximum water depth and average water depth along ice cliff of 180 m and 50 m respectively |
|              |              | 17.7                  |                              | Kirkbride (1995)            |                                                                                   |
|              |              | 40                    |                              | Roehl (2002)                | Average taken from different water temperatures                                    |
|              |              | 25 ± 5                |                              | Hochstein et al. (1998)     | Measured change in perimeter of a large, tabular iceberg which grounded in front of the ice cliff over 3 years |
| Argentina    | Ameghino     | 275                   |                              | Warren et al. (1995)        |                                                                                   |
|              | Moreno       | 800                   |                              | Warren et al. (1995)        |                                                                                   |
| Chile        | Leon         | 880                   |                              | Haresign and Warren (2005)  |                                                                                   |
| British Columbia | Bridge    | 70                    |                              | Chernos et al. (2016)       | Study period 1983-2013 estimated calving flux of 0.0362 km³                        |
| Himalaya     | Lirung       | 4-14 m a⁻¹            |                              | Sakai et al. (1998)         | Average observed supra-aqueous ice cliff melt during the melt season               |
|              | Ngozumpa     | 2.1 cm hr⁻¹           |                              | Benn et al. (2001)          | Waterline notch measurement taken in 1998                                         |
| Norway       | Svardisheibreen | 5                   |                              | Kennett et al. (1997)       |                                                                                   |

Table S2. Calving and subaqueous melt fluxes from different settings. Note change in units for Ngozumpa Glacier (Benn et al., 2001) in cm hr⁻¹
| Experiment Name         | Initial thickness | ELA (m a.s.l.) | Subaqueous melt rate (m a\(^{-1}\)) | Calving model     | Calving rate (m a\(^{-1}\)) |
|-------------------------|-------------------|---------------|--------------------------------------|-------------------|-----------------------------|
| SPIN UP                 | James et al. (2019) | 1465          | N/A                                  | N/A               | N/A                         |
| SENSITIVITY_CALVING_FLUX | SPIN UP           | 0.05          | 0                                    | Crevasse          | 10                          |
|                         | SPIN UP           | 0.05          | 0                                    | Crevasse          | 50                          |
|                         | SPIN UP           | 0.05          | 0                                    | Crevasse          | 100                         |
|                         | SPIN UP           | 0.05          | 0                                    | Crevasse          | 200                         |
|                         | SPIN UP           | 0.05          | 0                                    | Crevasse          | 500                         |
|                         | SPIN UP           | 0.05          | 0                                    | Crevasse          | 1000                        |
| SENSITIVITY_CALVING_MODEL | SPIN UP          | 0.05          | 0                                    | Flotation         | 100                         |
|                         | SPIN UP           | 0.05          | 0                                    | Rate proportional to speed | 100                 |
|                         | SPIN UP           | 0.05          | 0                                    | Crevasse          | 100                         |
|                         | SPIN UP           | 0.05          | 0                                    | No Calving        | 100                         |
|                         | SPIN UP           | 0.05          | 0                                    | Cliff Collapse    | 100                         |
|                         | SPIN UP           | 0.05          | 0                                    | Damage            | 100                         |
|                         | SPIN UP           | 0.05          | 0                                    | Maximum Extent    | 100                         |
|                         | SPIN UP           | 0.05          | -1                                   | Crevasse          | 0                           |
|                         | SPIN UP           | 0.05          | -10                                  | Crevasse          | 0                           |
|                         | SPIN UP           | 0.05          | -50                                  | Crevasse          | 0                           |
|                         | SPIN UP           | 0.05          | -100                                 | Crevasse          | 0                           |
| LAND                    | SPIN UP           | 0.05          | N/A                                  | N/A               | N/A                         |
| LAKE                    | SPIN UP           | 940           | -50                                  | Crevasse          | 100                         |

Table S3. Summary of experimental set-up, sensitivity analysis and forcing
Movie S1. LAND simulation

Movie S2 LAKE simulation using the Crevasse Calving Model where the crevasse depth was set such that it removed ice shelves and floating ice thinner than 100 m. The calving flux and subaqueous melt flux were set to 100 m a$^{-1}$ and -50 m a$^{-1}$ respectively.

Movie S3. LAKE simulation using the Crevasse Calving Model where the crevasse depth was set such that it removed ice shelves and floating ice thinner than 100 m. The calving flux and subaqueous melt flux were set to 100 m a$^{-1}$ and -50 m a$^{-1}$ respectively. Thickness has been inverted to clearly show the effect of the lake on the ice front.
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