Supplement of

PISM-LakeCC: Implementing an adaptive proglacial lake boundary in an ice sheet model

Sebastian Hinck et al.

Correspondence to: Sebastian Hinck (sebastian.hinck@awi.de)

The copyright of individual parts of the supplement might differ from the article licence.
S1  Sensitivity experiments

In this section a short overview of all sensitivity runs is given. These are also listed in Table S1. Temporal snapshots of all experiments are shown in Sect. S2. The main experiments discussed in the main manuscript (LAKE, CTRL and DEF) are not discussed here.

<table>
<thead>
<tr>
<th>Experiment name</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>IncCalv</td>
<td>increased calving; lacustrine thickness calving threshold set to 500m</td>
</tr>
<tr>
<td>RedCalv</td>
<td>reduced calving; lacustrine thickness calving threshold set to 20m</td>
</tr>
<tr>
<td>MR</td>
<td>adapted sub-shelf melting; melt parameter adapted to account for differences between marine and lacustrine environment</td>
</tr>
<tr>
<td>nSG</td>
<td>no slippery grounding line model</td>
</tr>
<tr>
<td>TWO</td>
<td>use of set_tillwat_ocean parameterization instead of slippery grounding line model</td>
</tr>
<tr>
<td>GIA</td>
<td>adapted Earth model parameters for the Lingle-Clark bed deformation model</td>
</tr>
<tr>
<td>FR5</td>
<td>max_fill_rate set to 5m year$^{-1}$</td>
</tr>
<tr>
<td>FR10</td>
<td>max_fill_rate set to 10m year$^{-1}$</td>
</tr>
<tr>
<td>FR50</td>
<td>max_fill_rate set to 50m year$^{-1}$</td>
</tr>
</tbody>
</table>

S1.1  IncCalv

As it has already been mentioned in the main text, vast ice shelves appeared in the LAKE experiment. However, we are uncertain in how far this aspect is realistic. We therefore conducted one experiment with increased thickness calving threshold for lakes. In this experiment almost all floating ice is immediately calved off. This value was set ten times higher than in the LAKE experiment: $\Delta h_L = 500$ m.

Increasing the calving rate speeds up the glacial retreat (see figures in Sec. S2). The break-up of the ice dome over Hudson Bay and the ice bridge connecting the LIS and CIS happen about 2000 yr earlier than in the LAKE experiment.

Figure S1: Mass losses for the LIS for the IncCalv experiment. Panel (a) shows the surface runoff for the entire LIS. Panels (b) and (c) show the calving and sub-shelf melt fluxes of the LIS at all lacustrine boundaries, respectively. Note, these fluxes are averaged over 100 yr. (d) The lines show the accumulated mass losses from the panels above.

Figure S1 shows the mass losses due to lacustrine calving and sub-shelf melting and relates them to the surface runoff. Since no major ice shelves are present, which are due to their low surface elevation subject to high melting, the runoff is
strongly reduced, compared to the LAKE experiment. The lack of ice shelves is also the reason why the sub-shelf melting is negligible. As long as major proglacial lakes are present, mass losses due to lacustrine calving is substantially higher than in the LAKE experiment. Surface runoff contributes most to the ice sheets mass balance.

S1.2 RedCalv

For the redcalv experiment the thickness calving threshold for lakes is reduced to 20m. Except for slightly larger ice shelves, the results are hardly different from the LAKE experiment (see figures in Sec. S2).

Figure S2: Mass losses for the LIS for the RedCalv experiment. Panel (a) shows the surface runoff for the entire LIS. Panels (b) and (c) show the calving and sub-shelf melt fluxes of the LIS at all lacustrine boundaries, respectively. Note, these fluxes are averaged over 100yr. (d) The lines show the accumulated mass losses from the panels above.

Figure S2 shows the mass losses due to lacustrine calving and sub-shelf melting and relates them to the surface runoff. The calving flux is reduced, compared to the LAKE experiment, while the sub-shelf melting is almost unchanged. Surface ablation is by far the dominant contribution to mass loss, and is slightly higher than in the LAKE experiment. When summing up the contributions from panels (a)-(c) (see panel (d)), it becomes clear, that the gain in runoff balances the reduction in sub-shelf melting.

S1.3 MR

The sub-shelf melting is calculated based on the melt pump scheme by [Beckmann and Goosse 2003]. Although the parameters of the model were adapted for a marine setting around Antarctica, it was used in the previous experiments in the lacustrine setting without adaptation. In the MR experiment we estimate the magnitude of the difference between marine and lacustrine melting and adapt a melt parameter of the model accordingly. The details are discussed in the following.

The melt flux is calculated as follows [Martin et al. 2011]:

\[ M = L^{-1} C \rho_{SW} c_p \gamma_{T,O} (T_O - T_f(S,d)) \]  

(S1)

Here, \( L \) is the latent heat of fusion, \( C \) is a tuning parameter, \( \rho_{SW} \) is the density of seawater, \( c_p \) is the heat capacity of the mixed layer, \( \gamma_{T,O} \) is the thermal exchange velocity, \( T_O \) is the temperature of the ambient ocean and \( T_f(S,d) \) the freezing temperature of water of salinity \( S \) at depth \( d \). We rewrite this formula for lakes:

\[ M = L^{-1} C \rho_{FW} c_p \gamma_{T,L} (T_L - T_f(S_L,d)) \]

\[ = L^{-1} C \rho_{SW} c_p C_\gamma \gamma_{T,O} C_T(T_O,T_L,S_L,d) (T_O - T_f(S,d)) \]  

(S2)
Here, $T_L$ is the temperature of the lake, $\gamma_{T,L}$ is the thermal exchange velocity for a lake, and the $C$ terms are factors to modify from lake to ocean parameterization:

$$C_\rho = \rho_{FW}/\rho_{SW} \approx 1 \quad \text{(S3)}$$

accounts for the difference in density. We neglect the density difference here and assume it is 1.

$$C_\gamma = \frac{\gamma_{T,L}}{\gamma_{T,O}} \approx \frac{1}{200} \quad \text{(S4)}$$

This value is not easy to estimate, it describes how efficient heat is transported away in the boundary layer. The thermal exchange velocity $\gamma_T$ is a function of the Nusselt number $Nu$ of the mixed layer (Holland and Jenkins [1999]), which is proportional to the velocity of the mixed layer. We assume that this velocity depends on the density contrast between the meltwater and the ambient water. In a marine environment the buoyancy of meltwater is about 200 times larger than in a lacustrine environment (Funk and Röthlisberger [1989]). We use this as our best guess: $C_\gamma = 200^{-1}$.

$$C_T = \frac{T_L - T_f(S_L,d)}{T_O - T_f(S,d)} \approx 5 \quad \text{(S5)}$$

This term cannot be expressed as a constant, as it depends on lake and ocean temperatures, depth and salinity. The marine values are set in the model: $S = 35$PSU and $T_O = -1.7^\circ C$, and also $S_L = 0$ for freshwater is fixed. The dependence on the other values is illustrated in Fig. $S3$. Even though the values vary a lot, we chose $C_T = 5$, which represents this value for $T_L = 2^\circ C$ and $d = 300$ m.

![Figure S3: Dependence of $C_T$ on the lake temperature $T_L$ for different depths $d$.](image)

Since we used $C = 0.01$ for the other experiments, we keep it at that value. With these estimates we can determine
our tuning factor $C^*$ for lacustrine settings:

$$C^* = C \cdot C_p \cdot C_\gamma \cdot C_T = 2.5 \cdot 10^{-4}$$

(S6)

Accordingly, the lacustrine sub-shelf melting is 40 times less effective than its marine counterpart.

We have to note that, when using this tuning factor sub-shelf melting is strongly reduced also at the marine boundaries. To prevent marine ice shelves from growing excessively large, we restrict the ice sheet in the MR experiment from expanding outside a prescribed margin. A respective mask was calculated in a way that the ice sheet growth was only restricted in marine areas, not terrestrial or lacustrine regions. This is the reason why the marine parts of the ice sheet differ from the other experiments (compare with Fig. S3).

Figure S4: Mass losses for the LIS for the MR experiment. Panel (a) shows the surface runoff for the entire LIS. Panels (b) and (c) show the calving and sub-shelf melt fluxes of the LIS at all lacustrine boundaries, respectively. Note, these fluxes are averaged over 100 yr. (d) The lines show the accumulated mass losses from the panels above.

Figure S4 shows the mass losses due to lacustrine calving and sub-shelf melting and relates them to the surface runoff. The mass loss due to lacustrine sub-shelf melting is strongly reduced, compared to the LAKE experiment, while the lacustrine calving flux is almost unchanged. Surface ablation dominates the ice sheet’s mass loss, while lacustrine sub-shelf melting is almost insignificant.

S1.4 nSG

For this run the slippery grounding line treatment of PISM, which was used for the other experiments, is disabled. Compared to the LAKE experiment, the transport of ice towards the marine and lacustrine boundaries is drastically reduced. As a consequence, the observed lowering of the ice surface upstream the lacustrine and marine boundaries is also smaller. The self-amplified feedback cycle, consisting of increased melting due to warmer air temperatures and warming temperatures due to surface lowering, that results in the rapid breakup of the ice sheet, as observed in the LAKE experiment, does not appear in this run (see figures in Sec. S3).

Although the mass loss is reduced compared to the LAKE experiment, the presence of lakes does increase the ice loss at the lacustrine boundaries compared to the CTRL experiment (see Fig. S5). At the marine ice margins, this lower mass flux leads to thicker ice shelves and an advanced grounding line.

Due to the thicker ice sheet, the rapid lacustrine advance underneath the grounded ice front, as observed over Hudson Bay for the LAKE experiment, does not happen here. Consequently, the self-amplified feedback mechanism, the PLISI, does not happen and the complete breakup of the LIS ice dome over Hudson Bay and the ice bridge to the CIS does not happen until the end of the run. In the final phase of the experiments the southern ice margin is a few hundred kilometers north of the no-lake experiments.
Figure S5: Ice thickness anomaly plot comparing the nSG and the CTRL experiment. Along the lacustrine margins and upstream of those the impact of the lake boundary, even in absence of the enhanced sliding parameterization, is obvious. At the marine boundaries, however, the mass loss is also strongly reduced, which can be seen in the strong positive anomaly.

S1.5 TWO

Another treatment of the grounding line is proposed in Albrecht et al. (2020). This method is called tillwater.ocean, hence the name of the experiment, and is not yet included into an official PISM release. This parameterization assumes the till at grounded ice cells, that have previously been ocean, to be saturated with water. This assumption is similar to the slippery grounding-line approach. As has also been reported by Albrecht et al. (2020), we find the results to be similar to the LAKE experiment, where the slippery.gl parameterization was used (compare with figures in Sec. S2).

S1.6 GIA

In this experiment different parameters for the Lingle-Clark bed deformation model of PISM were used. Our changes targeted the underlying Earth model, which was chosen accordingly to the parameterization used for NAICE (Gowan et al., 2016). The viscosity of the upper mantle is set to $4 \times 10^{20}$ Pa s and the lithosphere is assumed to be 120km thick. With these values the flexural rigidity of the lithosphere is calculated accordingly to Buier et al. (2007): $1.2627 \times 10^{25}$ Nm.

The results of this run are very similar to the LAKE experiment (see figures in Sec. S2). Differences in timing in the glacial retreat are due to the different calculated Earth response. The conclusions drawn from this experiment, however, are the same as for the LAKE experiment.

---

1 See the corresponding pull request on Github: [https://github.com/pism/pism/pull/425](https://github.com/pism/pism/pull/425) Accessed: 2021/08/15
S1.7 FR5, FR10 and FR50

Compared to the LAKE experiment for these three experiments (FR5, FR10 and FR50) only fill rates at which the water level is gradually adjusted is changed (5, 10 and 50 m yr$^{-1}$). Although the rates were drastically increased up to 50 times more rapid, the model stayed numerically stable. However, it can not be determined, whether this is because a reduced time step of 0.25 yr was chosen, or simply because no critical situation was triggered.

The more quickly adjusting water level comes closer to the original assumption that the lake basins are always filled. By looking at the overview maps in Sec. S2 no substantial difference in the ice sheet evolution can be determined, compared to the LAKE experiment.

References


S2 Overview maps
Figure S6: Overview map at 500yr for the different experiments
Figure S7: Overview map at 1000yr for the different experiments
Figure S8: Overview map at 1500yr for the different experiments
Figure S9: Overview map at 2000yr for the different experiments
Figure S10: Overview map at 2500yr for the different experiments
Figure S11: Overview map at 3000yr for the different experiments
Figure S12: Overview map at 3500yr for the different experiments
Figure S13: Overview map at 4000yr for the different experiments
Figure S14: Overview map at 4500yr for the different experiments
Figure S15: Overview map at 5000yr for the different experiments
Figure S16: Overview map at 5500yr for the different experiments
Figure S17: Overview map at 6000yr for the different experiments
Figure S18: Overview map at 6500 yr for the different experiments.
Figure S19: Overview map at 7000yr for the different experiments
Figure S20: Overview map at 7500yr for the different experiments
Figure S21: Overview map at 8000yr for the different experiments
Figure S22: Overview map at 8500yr for the different experiments
Figure S23: Overview map at 9000yr for the different experiments
Figure S24: Overview map at 9500yr for the different experiments
Figure S25: Overview map at 10000yr for the different experiments
Figure S26: Overview map at 10500yr for the different experiments
Figure S27: Overview map at 11000yr for the different experiments
Figure S29: Overview map at 12000yr for the different experiments
Figure S30: Overview map at 12500yr for the different experiments
Figure S31: Overview map at 13000yr for the different experiments
Figure S32: Overview map at 13500yr for the different experiments
Figure S33: Overview map at 14000yr for the different experiments
Figure S34: Overview map at 14500yr for the different experiments
Figure S35: Overview map at 15000yr for the different experiments
Figure S36: Overview map at 15500yr for the different experiments
Figure S37: Overview map at 16000yr for the different experiments
Figure S38: Overview map at 16500yr for the different experiments
Figure S39: Overview map at 17000yr for the different experiments
Figure S40: Overview map at 17500yr for the different experiments
Figure S41: Overview map at 18000yr for the different experiments
Figure S42: Overview map at 18500yr for the different experiments
Figure S43: Overview map at 19000yr for the different experiments
Figure S44: Overview map at 19500yr for the different experiments
Figure S45: Overview map at 20000yr for the different experiments
Figure S46: Overview map at 20500yr for the different experiments
Figure S47: Overview map at 21000yr for the different experiments
Figure S48: Overview map at 21500yr for the different experiments
Figure S49: Overview map at 22000yr for the different experiments
Figure S50: Overview map at 22500yr for the different experiments
Figure S51: Overview map at 23000yr for the different experiments
Figure S52: Overview map at 23500yr for the different experiments
Figure S53: Overview map at 24000yr for the different experiments
Figure S54: Overview map at 24500yr for the different experiments
Figure S55: Overview map at 25000yr for the different experiments
Figure S56: Overview map at 25500yr for the different experiments
Figure S57: Overview map at 26000yr for the different experiments