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Key Points:

- Meltwater forcing and sea ice feedbacks influence the evolution of the Laurentide Ice Sheet (LIS) surface mass balance (SMB) during deglaciation
- Simulated SMB is largely consistent with ICE-6G ice mass loss, but overestimates at 20–18 ka, 14.7–14.2 ka, and 13.3–12.7 ka
- Strongly negative SMB during the Bølling–Allerød is consistent with rapid mass loss in ICE-6G, although ice flow likely contributed as well

Supporting Information:

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

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Comparing the Simulated Surface Mass Balance of the Laurentide Ice Sheet During the Last Deglaciation With a Geophysical Reconstruction

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Abstract Despite decades of research, our understanding of the role of climate feedbacks in shaping Pleistocene glacial cycles remains incomplete. Here, we investigate the decay of the Laurentide Ice Sheet (LIS) by calculating its surface mass balance (SMB) during the last deglaciation using the isotope-enabled transient climate model experiment (iTraCE). To do this, we compare two frameworks for calculating the melt rate, employ a parameterization of the refreezing of snow melt and liquid precipitation, and apply statistical downscaling. We compare the SMB to the ice mass rate of change deduced from the ICE-6G reconstruction from Last Glacial Maximum (LGM; ~21 ka) until 12 ka. We find that the SMB derived from iTraCE overestimates mass loss from 20 to 18 ka (corresponding to an approximately 60% excess in cumulative mass loss) and more closely aligns with the ICE-6G after 18 ka. A key finding is that meltwater forcing plays a critical role in shaping SMB evolution. Its inclusion produces abrupt shifts in SMB and improves agreement with ICE-6G during Heinrich Stadial 1 and the Younger Dryas, highlighting the importance of ocean circulation and sea ice feedbacks in modulating melt rates. During the Bølling–Allerød, strongly negative SMB is consistent with mass loss in ICE-6G, but dynamic ice loss is required, perhaps triggered by surface melt. These results demonstrate that SMB derived from transient climate simulations can reproduce key features of deglacial ice sheet evolution when meltwater forcing is appropriately included, and underscore the importance of coupled surface and dynamic processes in shaping ice sheet evolution.

Plain Language Summary This study investigates how climate feedbacks influenced the melting of the Laurentide Ice Sheet (LIS) during the last deglaciation, between about 20,000 and 12,000 years ago. We use a climate model simulation to calculate how the ice sheet gained or lost mass at its surface (surface mass balance, or SMB), and compare these results to an independent reconstruction of ice sheet changes. We find that the model overestimates ice loss early in the deglaciation (20,000–18,000 years ago), though this is not obvious in the raw data. Later, the model results agree more closely with the reconstructed ice loss. A key finding is that meltwater entering the ocean plays an important role in controlling how fast the ice sheet melts. When included in the model, this meltwater forcing changes sea ice, reduces surface melting, and improves agreement with the reconstruction. During the Bølling–Allerød, strong surface melting may have increased ice flow, accelerating ice loss. Overall, this study shows that both surface melting and ice dynamics are important, and that ocean–ice feedbacks can strongly influence ice loss.

1. Introduction

The retreat of the Laurentide Ice Sheet (LIS) following the Last Glacial Maximum (LGM; ~21 ka) profoundly reshaped the North American landscape and offers a recent example of interactions between ice sheet dynamics and climatic warming. The history of this continental-wide ice retreat is reconstructed from uplift rates resulting from continental isostatic adjustment (e.g., Peltier et al., 2015). The deglaciation spanned approximately 21 ka to ~6.5 ka (Clark et al., 2009; Ullman et al., 2016) and involved significant climatic shifts. Heinrich Stadial 1 (HS1, 18 ka to 14.5 ka) involved severe cooling and the discharge of icebergs into the North Atlantic (Hemming, 2004). Meltwater pulse 1A (MWP-1A), occurring approximately at 14.6 ka, raised global mean sea level by roughly 14–18 m in no more than 500 years (Deschamps et al., 2012; Fairbanks, 1989; Hanebuth et al., 2000; Peltier & Fairbanks, 2006). Although its sources remain uncertain, recent sea-level fingerprinting suggests contributions of approximately 3 m from the Laurentide Ice Sheet, 7 m from the Eurasian Ice Sheet Complex, and 5 m from the West Antarctic Ice Sheet (Coonin et al., 2025). The rapid warming brought by the onset of the Bølling–Allerød

(BA) interstadial period (14.5 ka to 12.9 ka) accelerated North American ice-sheet retreat and contributed to the ice-saddle collapse (Gregoire et al., 2016). This warm period was then interrupted by the Younger Dryas (YD) stadial (12.9 ka to 11.7 ka), marked by rapid cooling and a temporary return to glacial conditions (Broecker et al., 1989). The dramatic retreat of past ice sheets may serve as an intriguing analog for future warming consequences, given that the retreat of LGM ice sheets led to an average sea-level rise of approximately 1 m per 100 years, comparable to some future climate projections (Fox-Kemper et al., 2021; Horton et al., 2014).

One important control of ice sheet growth and decay is the surface mass balance (SMB), whose variations affect the rate of change of ice sheet mass (Fyke et al., 2018; Khan et al., 2015; Kapsch et al., 2021; van den Broeke, 2007). SMB is controlled by snowfall rate, with snow eventually compacting into ice, and by mass loss by melting, evaporation, and sublimation. Meltwater and liquid precipitation can either run off the surface of the glacier/ice sheet, penetrate via moulins to the ice sheet base, or percolate into the snow layer and refreeze (Pfeffer et al., 1991). In addition to the SMB, ice sheets are also influenced by ice flow, which redistributes mass and drives dynamic losses via calving and basal melting; together, these processes account for the remaining mass loss during deglaciation.

Atmospheric and oceanic feedbacks strongly influence the spatial distribution and magnitude of SMB. Sea ice cover may affect moisture sources for the ice sheet, influencing its growth and retreat (Gildor & Tziperman, 2000). The presence of the LIS is thought to have profoundly influenced the atmospheric circulation and climate over North America (Broccoli & Manabe, 1987; Fu, 2023; Kutzbach & Wright, 1985; Liakka & Löfverström, 2018; Löfverström et al., 2015), affecting the SMB of the LIS (e.g., Liakka & Löfverström, 2018; Löfverström et al., 2015; Roe & Lindzen, 2001). The effects of the LIS on ocean circulation have also been examined in several studies (e.g., Gong et al., 2015; Ullman et al., 2014; J. Zhu et al., 2014). Still, our understanding of the atmospheric and oceanic feedbacks that affect the evolution of ice sheets during the last deglaciation remains incomplete.

Several studies have examined the SMB of ice sheets during and after the LGM using the outputs from general circulation models (GCMs) or Earth system models (ESMs). Carlson et al. (2009) explored the SMB of what was left of the LIS during the early Holocene (~9 ka) using an energy–mass balance model from Anslow et al. (2008) and found that the 9 ka LIS SMB contributed to a sea-level rise rate of 0.96 ± 0.19 cm/yr, consistent with geological records. Using the same energy-mass balance model and refreeze scheme but with a different GCM, Carlson et al. (2012) investigated the response of the LIS to the Bølling-Allerød warming and MWP-1A using a calculation of SMB using the GCM output. Their study suggested that the ICE-5G model (Peltier, 2004) might overestimate the LIS contribution to MWP-1A. In another study, Ullman et al. (2015) found that the SMB of the LIS for key time slices (24, 21, 19, 16.5, 15.5, 14, 13, 11.5, and 9 ka), simulated by a GCM (Schmidt et al., 2014), was positive (i.e., contributing the ice sheet growth rather than decay) throughout much of the deglaciation. Ice flow could lead to a reducing ice mass even in the presence of a positive SMB, and Ullman et al. (2015) concluded that ice flow and the resulting dynamic discharge were indeed the main cause of mass loss until about 9 ka. This study therefore implied that the sign of the SMB is not a good predictor of ice sheet growth or decay. In direct contrast, Bradley et al. (2024) found that the SMB for the North American Ice Sheet Complex (including the LIS) was negative during the LGM (21 ka and 26 ka using ice sheet geometry history reconstruction GLAC-1D; Tarasov et al., 2012), pointing to a possible discrepancy between the GCM SMB estimates and the reconstructed ice volume history.

Kapsch et al. (2021) extended the analysis beyond specific time slices, examining the SMB and equilibrium line altitude (ELA) of various ice sheets during the last deglaciation. They used a transient climate model simulation (MPI-ESM; Mauritsen et al., 2019) at different resolutions, combined with an energy balance model that included a parameterization for albedo evolution. Kapsch et al. (2021) found that the SMB changes at the beginning of the deglaciation are a direct result of compensating effects of increasing accumulation rate and melt rate, with a decrease in SMB occurring around 14 ka, indicating the onset of the deglaciation coincided with the onset of the BA warming. They used ice topographies from both GLAC-1D (Tarasov et al., 2012) and ICE-6G (Peltier et al., 2015), focusing on the SMB evolution of the Greenland Ice Sheet (GrIS), and of the ELA of the LIS. Additionally, Gregoire et al. (2015) evaluated the relative roles of orbital forcings and greenhouse gases during deglaciation using a set of transient dynamical ice sheet model (GLIMMER-CISM; Rutt et al., 2009) runs forced by a GCM. They found that by 9 ka, orbital forcing caused 50% of the deglaciation of Northern Hemispheric ice sheets, whereas changes in greenhouse gas concentrations caused only 30%, with the remaining mass loss caused

by the interaction between the two. Given that the above-mentioned studies find differing SMB of the LIS throughout the last deglaciation, one concludes that there is still no consensus on whether the SMB of the LIS was positive or negative during the last deglaciation.

Our goal here is to study the SMB history during deglaciation, produced by the isotope-enabled transient climate experiment simulation (He et al., 2021, hereafter iTraCE), a sophisticated coupled climate simulation covering 20–11 ka, introducing several novel elements to the analysis: First, we compute the SMB of the LIS continuously throughout the deglaciation period. Second, we compare the area-integrated SMB with the ice mass loss rate inferred from independent geophysical constraints as represented by the ice sheet reconstruction in ICE-6G (Peltier et al., 2015). For this purpose, we calculate a time series of the SMB of the LIS utilizing climate model output from iTraCE, and downscale it to higher spatial resolution to better represent melting in the steep ice sheet margins. We supplement this offline SMB calculation with a refreeze parameterization (Colleoni et al., 2009; Krinner et al., 2007; Thompson & Pollard, 1997) to account for the fate of surface meltwater or rain that percolates into the thermally active snow layer and is later refrozen. The overall purpose of this study is to assess whether the iTraCE simulations produce the forcing required to drive the prescribed ice sheet changes that serve as boundary conditions for these simulations. Establishing this consistency is critical for determining whether the simulated climate fields can be used to draw physically meaningful conclusions about the mechanisms of deglacial ice sheet retreat. Specifically, any mismatch would imply that the simulated climate is not dynamically consistent with the imposed ice sheet evolution, limiting its usefulness for interpreting deglacial processes.

We find that ice mass rate of change deduced from the SMB overestimates that implied by ICE-6G between 20 and 18 ka, potentially reflecting either an overly warm iTraCE climate or inaccuracies in the prescribed ICE-6G geometry and/or ice loss history. Afterward, the two estimates are in closer agreement. The SMB also captures the major features of deglacial mass loss, including the acceleration during HS1 and the BA, and a slowdown during YD only when meltwater forcing is included. Our results provide a basis for evaluating the role of SMB in driving ice sheet change during the last deglaciation.

This paper is organized as follows: we describe the methods used in Section 2 where we first briefly summarize the ICE-6G reconstruction and describe the iTraCE experimental setup in Section 2.1. We then describe the SMB calculation in Section 2.2 and report our results in Section 3. Finally, we discuss our results in Section 4 and conclude in Section 5.

2. Methods

2.1. Inputs: The ICE-6G Reconstruction and the iTraCE Experiment

The ICE-6G ice sheet reconstruction (Peltier et al., 2015) uses a glacial isostatic adjustment model to calculate global ice thickness history, providing an estimate for the evolution of ice volume and ice sheet topography over the last deglaciation. This ice thickness and volume history are given on a $1^\circ \times 1^\circ$ global grid every 500 years. A key strength of ICE-6G is its integration of geological age control and uplift rate observations within a glacial isostatic adjustment framework, which constrains the timing and spatial pattern of ice sheet retreat. This reconstruction, however, does not incorporate ice flow and SMB, rendering it not necessarily glaciologically self-consistent (Tarasov et al., 2012). However, it provides a constraint on the ice sheet mass evolution, independent of that deduced from the SMB. We compare the ice mass changes deduced from the SMB to the net mass changes calculated from changes between ICE-6G snapshots. The mass evolution of the LIS, as estimated from ICE-6G, results from both the SMB and ice flow. The deglaciation mass loss rate calculated from SMB alone should, therefore, be consistently smaller than that estimated from the geophysical constraints, with the residual accounted for by ice flow.

We calculate the SMB during the deglaciation using the output of iTraCE, a transient ESM simulation of the last deglaciation performed using the isotope-enabled version of the Community ESM version 1.3 (iCESM1.3, Brady et al., 2019). While this version of the model also simulates water isotopes, this study uses only the physical climate variables from the simulation. The iTraCE runs use the ice sheet geometry from ICE-6G (interpolated to the grid of the GCM), shown in Figure S8 in Supporting Information S1. Therefore, and for consistency with the iTraCE boundary conditions, we use ICE-6G (rather than other ice reconstructions, e.g., Tarasov et al., 2012) as an upper bound on the magnitude of the deglaciation mass loss rate, and a point of comparison for the SMB mass

loss rate. The horizontal resolution of the iTraCE atmosphere and land components is 2° . The runs do not include a dynamically evolving ice sheet, and instead prescribe the ice sheet geometry as boundary conditions that are updated every 1,000 years (except at 20 ka where LGM ice topography is used).

The atmospheric model used in iTraCE, the Community Atmospheric Model (CAM5, Neale et al., 2012) uses the Rapid Radiative Transfer Method parameterization (RRTMG, Iacono et al., 2008; Mlawer et al., 1997), which uses a correlated k distribution method for calculating radiative fluxes and heating rates. RRTMG is a widely used radiative transfer code, and shows improvements in its agreement with line-by-line radiative calculations compared to the older CAM radiation package (CAMRT, Neale et al., 2012). The Snow, Ice, and Aerosol Radiative Model (SNICAR) is used to simulate snow albedo and the absorption of solar energy within individual snow layers. The snow albedo is influenced by the solar zenith angle, IR band, the albedo of the surface beneath the snow, concentrations of aerosols deposited from the atmosphere (e.g., black carbon, mineral dust, and organic carbon), and the effective grain size of ice (r_e), which is modeled through a snow aging process (Flanner et al., 2007; Flanner & Zender, 2005; Oleson et al., 2013). Glacier albedos are set to 0.80 and 0.55 in the visible and near-infrared, respectively (Lipscomb & Sacks, 2012).

The iTraCE experimental design includes four experiments that additively apply the four major changing boundary conditions during the deglaciation. The first experiment (hereafter ICE) only involves changing ice sheets and ocean bathymetry, holding orbital conditions and greenhouse gas concentrations at LGM values. Next, insolation forcing from changing orbital conditions was added (hereafter ICE + ORB). Then, greenhouse gases (hereafter ICE + GHG + ORB) and meltwater fluxes were prescribed (referred to below as ICE + GHG + ORB + MWF or simply iTRACE). Following He et al. (2021), we use “iTRACE” to refer to the fully forced simulation and “iTraCE” to reference the entire simulation suite. The iTRACE run is branched from the ICE + GHG + ORB run starting at 19 ka. This experimental setup is similar to that of Gregoire et al. (2015) who also investigated the contribution of greenhouse gases and evolving orbital parameters during the last deglaciation using a dynamical ice sheet model (GLIMMER-CISM; Rutt et al., 2009) forced by a GCM. Previous studies have shown that iTraCE captures key large-scale features of deglacial climate variability (e.g., Du et al., 2021; Fu, 2023; He et al., 2021; C. Zhu et al., 2022), supporting its use as a framework for evaluating SMB evolution.

2.2. Melt Rate Calculation

To calculate the SMB of the LIS, we account for snowfall, sublimation, melting calculated via the positive degree day scheme as described in Section 2.2, and refreezing of liquid precipitation and meltwater. We do not consider the contribution to SMB from snow blowing/redistribution as it is not a model output of iTraCE. Over ice sheets, the land component of iCESM1.3, the Community Land Model or CLM4.5 (Oleson et al., 2013), which is used in the iTraCE runs, assumes that sublimation occurs rather than evaporation, which we denote here as E . Snowfall is denoted by P_S and rainfall by P_L . We focus on the SMB and the evolution of the LIS, not including the Eurasian, Greenland, and Antarctic ice sheets. In the iTraCE archive, the post-12 ka portion of the simulation was saved with a different temporal resolution and file-handling convention than the output for earlier periods, resulting in missing years/months of data. To ensure a consistent monthly chronology across the full deglaciation interval, we end our analysis at 12 ka. We smooth the SMB time series in some cases, as indicated below, using the Savitzky-Golay filter (Python package `scipy.signal.savgol`) with a window length of 100 years and a linear polynomial fit.

Surface melt is normally calculated using one of two methods: the positive degree days method (hereafter PDD, e.g., Braithwaite, 1995; Hock, 2003; van de Wal, 1996) or the surface energy balance method (SEB; e.g., Lipscomb et al., 2013; Vizcaíno et al., 2014). The SEB of an ice sheet is the small residual of large terms (specifically net shortwave and longwave radiation) and any small errors/perturbations in any of the input variables could produce large deviations in the melting calculation. Since iTraCE uses a 2° grid and only monthly output is available, small errors in any of these input variables are likely (e.g., due to cloud parameterizations), thus potentially causing large discrepancies in the SEB. In addition, our results using the SEB method reveal some unphysical patterns, for instance ablation zones that extend too far into the interior of the ice sheet (see Section 3.3 and Figure 5 for details). Therefore, we chose to use the robust and widely used PDD method to calculate melt for our main analysis. The methods and results using the SEB method can be found in Section 3.3 and in Supporting Information S1.

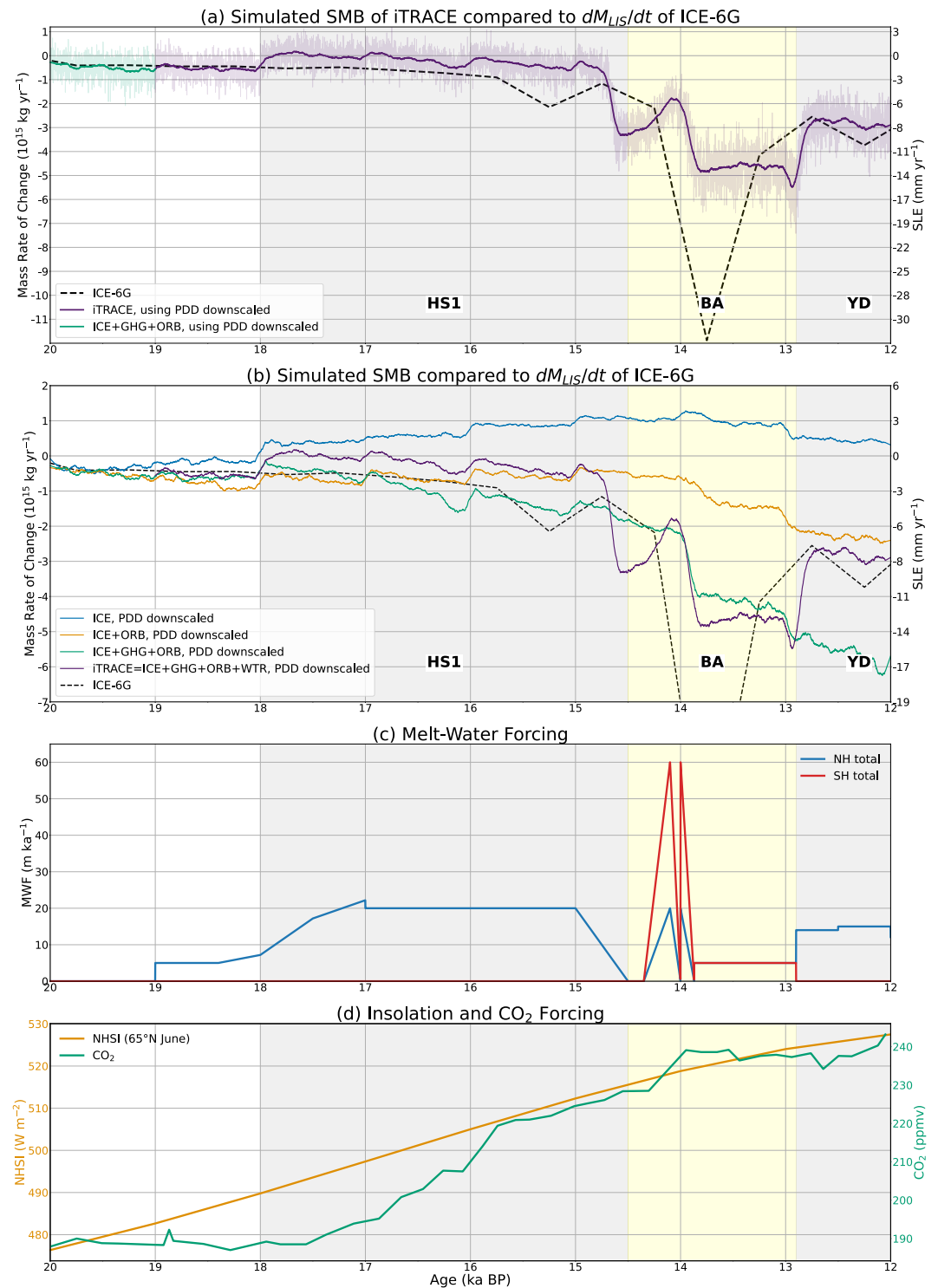


Figure 1. (a) Downscaled surface mass balance (SMB) integrated over the Laurentide Ice Sheet (LIS) from the fully forced iTRACE experiment (purple), compared to the LIS ice mass rate of change from ICE-6G (black; Equation 4). The dark purple line shows the smoothed annual SMB, with light shading indicating annual variability. Gray, yellow and gray shading denote Heinrich Stadial 1 (18–14.5 ka), the BA (14.5–12.9 ka), and the Younger Dryas (12.9–11.7 ka), respectively. (b) Time-smoothed SMB from all iTRACE experiments (blue: ICE, orange: ICE + ORB, green: ICE + GHG + ORB, purple: iTRACE). (c) MWF in the Northern Hemispheric in blue and Southern Hemispheric total in red. (d) June 21st insolation at 65°N in orange (Berger & Loutre, 1991) as well as the CO_2 in green (Lüthi et al., 2008).

The PDD method parameterizes melt by assuming that the snow or ice melt rate is proportional to the number of days above the freezing temperature. The physical basis for this melting scheme is that radiative and turbulent fluxes (primarily net longwave, sensible heat flux and latent heat flux) over a melting ice sheet are correlated with the 2-m air temperature (T_{2m} , Ohmura, 2001). Since this method requires daily T_{2m} data and iTraCE provides only monthly averages, we make the following assumptions. First, we assume that the daily temperatures can be represented by assuming a Gaussian distribution around the monthly average with a 5°C standard deviation (in the range of typical values found for the GrIS, Calov & Greve, 2005; Wake & Marshall, 2015). The total monthly melt at a given grid cell, for either snow (s) or bare-ice (i) surfaces can be found by,

$$M = \sum_d DDF_{s,i} H(T_{2m}(d) - T_0) \Delta t, \quad (1)$$

where d is the day during the month, Δt is taken to be 1 day, and H is the Heaviside function. We chose a degree-day factor, DDF , for snow-covered cells to be $DDF_s = 2.7$ mm/°C/day, and for bare-ice covered cells to be $DDF_i = 7.2$ mm/°C/day (Janssens & Huybrechts, 2000). We set $T_0 = 0^\circ\text{C}$. Although van den Broeke et al. (2010) found that a threshold of $T_0 = -5^\circ\text{C}$ better matches observations for present-day surface melt over the GrIS, this offset primarily reflects near-surface energy balance processes that are explicitly resolved differently in iTraCE. The relevance of this modified reference temperature to our setup that represents large-scale LIS boundary conditions is not obvious, and we follow common practice and adopt $T_0 = 0^\circ\text{C}$.

We downscale the variables used to calculate the SMB to a higher resolution of 10 arc min using the topography given by Argus et al. (2014) and Peltier et al. (2015) following Sellevold et al. (2019). The 2-m temperature is downscaled using a lapse rate of 7°C/km (Thompson & Pollard, 1997). Previous studies have chosen a lapse rate between 6.5°C/km (e.g., Löfverström & Liakka, 2018) and 8°C/km (Berends et al., 2018); we find that our results are not sensitive to this choice (Figure S9b in Supporting Information S1). The sublimation is downscaled using a vertical gradient of 47 mm/yr/km. For our downscaling method, we use 10 elevation classes where the boundaries are fixed at 0, 200, 400, 700, 1,000, 1,300, 1,600, 2,000, 2,500, 3,000, and 10,000 m elevation (Sellevold et al., 2019) and downscale to a 10 min resolution (Argus et al., 2014; Peltier et al., 2015). Then, we apply different lapse rates/rates of change with height to each of the input variables (Sellevold et al., 2019). The downscaling is applied to temperature, snowfall, rainfall, and sublimation to calculate a downscaled SMB (see Section 2.4). Snowfall and rainfall rates are downscaled through simple interpolation without phase corrections due to changing elevations, whereas the sublimation is downscaled using an assumed vertical gradient of 47 mm/yr/km (Sellevold et al., 2019).

To demonstrate the role of downscaling of the relatively coarse resolution of iTraCE, we analyzed the SMB using the original 2° grid. The results (Figure S9 in Supporting Information S1) further confirm the known importance of downscaling as the coarser grid increases ablation by about a factor of 5.

2.3. Refreeze Parameterization

A fraction, f , of rain and meltwater is refrozen into the near-surface permeable layer of snow and firn rather than running off the surface of the ice sheet. Based on Pfeffer et al. (1991), Thompson and Pollard (1997) proposed a parameterization that links f to the ratio of annual snow or ice melt rate ($\overline{M}$) to annual snowfall rate ($\overline{P}_s$). This parameterization is then translated to a fraction of water that undergoes refreeze following Colleoni et al. (2009),

$$f = 1 - \min\left(1, \max\left(0, \frac{\overline{M}/\overline{P}_s - 0.7}{0.3}\right)\right). \quad (2)$$

The refreeze typically occurs at the start and end of the melt season, when the temperature of the thermally active snow layer is colder. That means that meltwater does not necessarily freeze in the same month in which ice melted, necessitating the use of annual quantities for the formulation of the parameterization. This refreeze parameterization was originally used and developed for the GrIS. We adopt the GrIS refreeze parameterization for our purposes as it encodes process-level physics (meltwater percolation, firn thermal capacity, and pore-space availability) that are not unique to GrIS. Although the southern margin of the LIS lay at lower latitudes, the climate during the deglaciation over much of the LIS was cooler than modern mid-latitudes. Firn temperatures and retention capacity were therefore plausibly within the regime where the GrIS formulation remains applicable.

Note that by including refreezing or “internal accumulation rate”, we are formally modeling the “climatic mass balance” (Cogley et al., 2011) and are therefore choosing to deviate from the formal definition of SMB. As shown later, the inclusion of this specific refreeze parameterization has a very small effect on overall ice sheet mass balance in this case.

2.4. SMB Calculation and the LIS Rate of Change From ICE-6G

The annual SMB is calculated offline from the model outputs of iTraCE at each grid point in m/s water equivalent (w.e.) as in Vizcaíno et al. (2014),

$$SMB = \bar{P}_s + f\bar{P}_L - \bar{E} - (1 - f)\bar{M}. \quad (3)$$

Overbars denote annual averages again, and the SMB terms, therefore, include the snowfall, refrozen fraction of liquid precipitation, minus sublimation, and minus the part of the melt rate that is not refrozen. Finally, the SMB at each grid point is multiplied by the number of seconds in a year to convert to units of m (w.e.)/yr and multiplied by the density of water to convert to a mass change unit of kg/m²/yr.

The ICE-6G rate of change of the LIS ice mass, shown in the figures below, is calculated at times t midway between given LIS ICE-6G ice volume estimates $V(t)$ using

$$\left. \frac{dM_{LIS}(t)}{dt} \right|_t = \rho_{ice} \frac{V(t + 250 \text{ yr}) - V(t - 250 \text{ yr})}{500 \text{ yr}}, \quad (4)$$

where $V(t)$ is calculated by integrating the ICE-6G ice thickness over the LIS area, and it is available from 26 ka to 0 ka in 500-year increments. We multiply the finite difference volume rate of change by the assumed value for density of ice, ρ_{ice} , to convert to an ice mass rate of change.

We assess the sensitivity of SMB to plausible variations in key PDD parameters. While the SMB exhibits some dependence on these choices, the overall temporal structure and magnitude of inferred mass-loss trends are preserved across the tested parameter range. The degree-day factors and temperature variability are constrained within relatively narrow, observation-based ranges (Braithwaite, 2008; Reech, 1991; Zheng et al., 2023), and perturbations within these bounds do not qualitatively alter the evolution of SMB.

3. Results

In Section 3.1, we first present time series of the calculated SMB for each of the iTraCE experiments throughout the deglaciation, calculated as explained in Section 2. We compare this to the rate of change of the LIS ice mass deduced from ICE-6G using Equation 4. We also assess the statistical significance of differences between the different climate model experiments and ICE-6G (see Figures S3–S6 in Supporting Information S1). We then analyze the different components contributing to the SMB in Section 3.1. Next, we discuss the sensitivity to, and importance of downscaling the 2° results onto a 10-min grid. We then analyze the spatial structure of the SMB and its components in Section 3.2. Finally, we briefly compare our PDD-based results to ones using a surface energy balance to calculate melt in Section 3.3 and Figure 5.

To quantify the sensitivity to the PDD parameters, we perform a targeted set of experiments in which the standard deviation (σ) and degree-day factor DDF (for both snow- and ice-covered surfaces together) are perturbed by $\pm 10\%$ relative to the baseline iTRACE experiment (Figures S7a–S7b in Supporting Information S1). Integrated over the entire simulation period, we find that a $+10\%$ (-10%) perturbation in σ results in a -18.4% ($+16.6\%$) change in total mass change, whereas a $+10\%$ perturbation in $DDF_{s,i}$ yields a -33.5% ($+33.8\%$) change. Thus, SMB is more sensitive to DDFs than to temperature variability within the tested range.

We also note that the temporal response is not uniform: SMB anomalies relative to the iTRACE baseline (Figure S7b in Supporting Information S1) indicate enhanced sensitivity during the BA, when the small parameter perturbations produce amplified deviations compared to earlier and later intervals. This reflects the increased importance of melt processes during this interval, although the timing and overall pattern of SMB change remain consistent across the experiments.

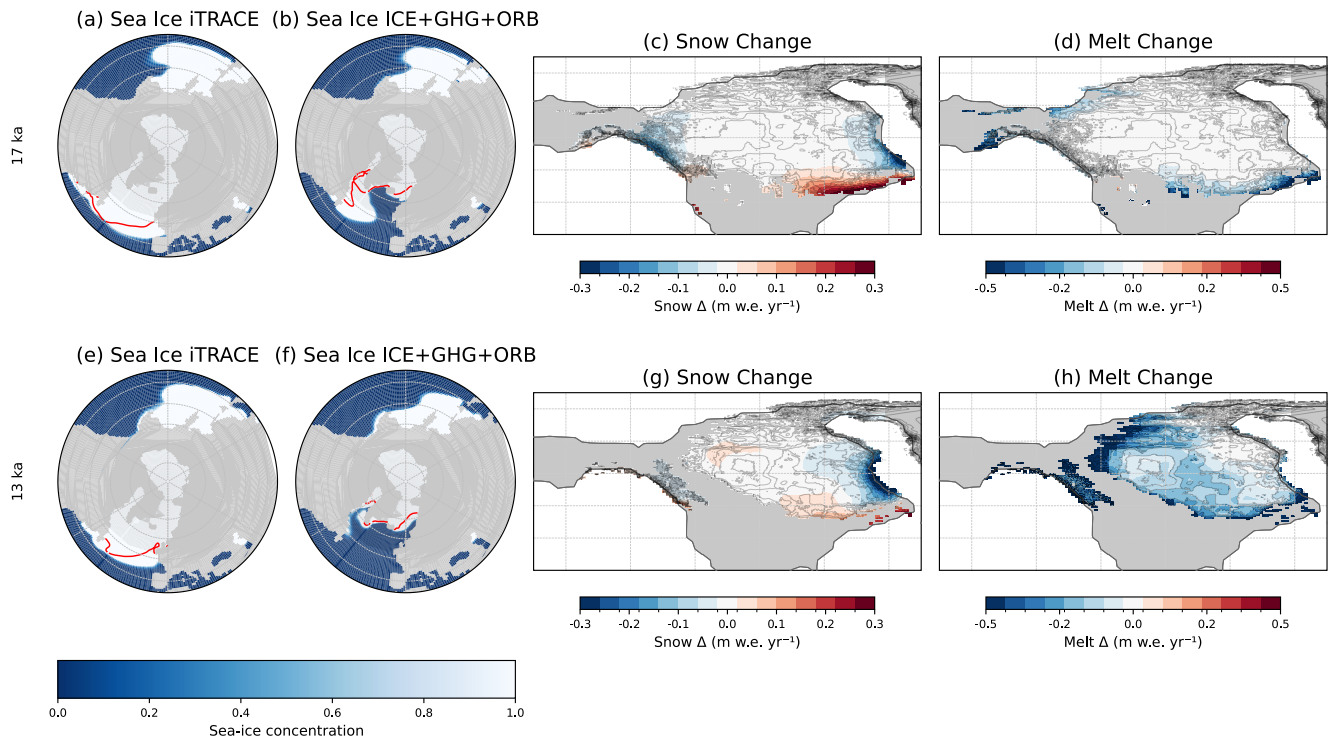


Figure 2. The interaction between meltwater, sea ice, and SMB. The upper row shows the 17–16 ka time average, and the lower 13–12 ka. (a) March sea ice concentration (blue to white shading) and the 85% September concentration line in red for the iTRACE experiment. (b) Same as (a) but for ICE + GHG + ORB. (c) The color shading shows the difference between snowfall (averaged between 17 and 16 ka) in iTRACE and ICE + GHG + ORB experiments. The difference is, therefore, a result of the effects of meltwater. The effects are mixed, with some areas showing enhancement (red) and some a decrease in snowfall (blue) due to the introduction of meltwater forcing. (d) Same as (c) but for melt rate. The blue regions indicate a lower melt rate for the iTRACE experiment as compared to the ICE + GHG + ORB. The lower panels (e) to (h) are the same as (a) to (d) respectively, but instead averaged between 13 and 12 ka. The gray contours in c/d and g/h represent the ICE-6G downscaled volume elevation classes (see Figure S8 in Supporting Information S1).

3.1. The Overall Laurentide Ice Sheet SMB During Deglaciation

The dark purple line in Figure 1a is the smoothed yearly SMB calculated using downscaling for the fully forced iTRACE experiment. The shaded, light purple depicts the unsmoothed yearly SMB, which fluctuates with a range of about $\pm 0.5 \times 10^{15}$ kg/yr around the smoothed SMB.

Between 20 ka and 18 ka, the SMB parameterization given by Equation 3 indicates a rate of ice mass decline that is faster than that of ICE-6G (black line, Equation 4). Integrating the smoothed SMB prediction over this interval yields a total mass loss of $\sim 1.02 \times 10^{18}$ kg, compared to $\sim 6.36 \times 10^{17}$ kg from ICE-6G, corresponding to a difference of about 60%. The SMB and ICE-6G time series are statistically distinguishable over this period (Figure S3 in Supporting Information S1), indicating that the difference is robust relative to internal variability. This discrepancy indicates a difference between SMB-derived mass loss and the ICE-6G reconstruction during this interval.

From 18 to ~ 14.5 ka and between 13 and 12 ka, the SMB-derived mass loss rate becomes smaller than that inferred from ICE-6G, and remains comparable magnitude to ICE-6G through much of the subsequent deglaciation. Some notable deviations occur during brief intervals, including around 14.5 ka and 13 ka, when SMB indicates more negative mass balance than ICE-6G.

3.1.1. The Effect of the Individual Forcing Experiments on SMB

Figure 1b includes the other experiments in the iTraCE simulation suite (blue: ICE, orange: ICE + ORB, and green: ICE + GHG + ORB). These factorial experiments isolate the effects of individual forcings on the SMB evolution.

The ICE experiment shown by the blue curve in Figure 1b indicates that *the ice sheet albedo and melt-elevation feedback by itself is not sufficient to change the SMB to negative during deglaciation*. In this experiment, simulated SMB becomes positive after 18 ka and remains so for the rest of the experiment. This experiment represents the internal ice-climate feedbacks, including the roles of ice albedo, atmospheric circulation, the atmospheric lapse rate, and the elevation-desert effect. This indicates that additional external forcings are required to drive sustained negative SMB.

Results from the ICE + ORB experiment (orange curve in Figure 1b) indicate that overall, *orbital conditions cause the biggest impact on the SMB trajectory*, enough to switch it from positive to negative, primarily after 18 ka. Orbital conditions both increase the magnitude of SMB and affect the overall trajectory of the SMB-induced mass loss rate. Though the ICE + ORB SMB curves in Figure 1b exhibit the same substantial annual variability as seen in Figure 1a, the ICE curve is distinguishable from the ICE + ORB (see Figure S4 in Supporting Information S1), and one can robustly say that orbital forcing drives this change in SMB trajectory.

In the ICE + GHG + ORB experiment (green curve in Figure 1b), *greenhouse gases forcing amplifies surface mass loss*, particularly during and after BA. This is seen after about 17 ka, when SMB becomes more negative relative to the ICE + ORB case. Additionally, ICE + ORB and ICE + GHG + ORB simulations are also statistically distinguishable (see Figure S5 in Supporting Information S1).

The addition of meltwater forcing reduces the magnitude of surface mass loss and brings the SMB into closer agreement with ICE-6G after 18 ka as shown by the purple curve in Figures 1a and 1b. A key feature of this experiment is an abrupt shift in SMB at ~14.5 ka that is absent in the other experiments (Figure 1b), indicating that *this transition is directly driven by the inclusion of meltwater forcing*. During the YD, the experiment without meltwater forcing (ICE + GHG + ORB) continues to overestimate mass loss, whereas the iTRACE simulation shows reduced melt and improved agreement with ICE-6G. The SMB simulated by iTRACE and ICE + GHG + ORB are statistically distinguishable (Figure S6 in Supporting Information S1), confirming that these differences arise from the imposed meltwater forcing.

Finally, all experiments except ICE show too much surface mass loss between 20 and 18 ka and are inconsistent with the loss rate deduced from ICE-6G. The only experiments that are consistent with ICE-6G in the period of 18 ka to 16 ka are the one with the meltwater added (purple curve in Figure 1b) and the ICE experiment (blue). All these experiments are statistically distinguishable from each other (Figures S4–S6 and Table S2 in Supporting Information S1). We have examined the sensitivity of the above results to the downscaling by calculating the SMB using the coarse iTrace output, and the results are shown in Section S3 and Figures S9–S10 in Supporting Information S1.

3.1.2. The Role of Sea Ice in Meltwater Effects on the SMB

As shown above, meltwater forcing plays a significant role. This effect appears to operate, at least in part, through changes in sea ice. Specifically, *the expansion of sea ice during the YD appears to dramatically decrease surface melting* and has less of an effect on snowfall rates.

Figure 2 shows the response of March sea ice to meltwater forcing (Figures 2a and 2b, for 17 ka and Figures 2c and 2f for 13 ka), and the resulting effects on snow accumulation and surface melt (Figures 2c, 2d, 2g and 2h). A substantial expansion of sea ice is observed due to a weakened AMOC (Figures 2a–2e, Gildor & Tziperman, 2000; McManus et al., 2004; Obase et al., 2021; Pearce et al., 2013; Pöppelmeier et al., 2023). This expansion is associated with reduced snowfall on the eastern margin of the ice sheet and modest increases along the southeastern inland margins (Figures 2c–2g). At the same time, melt rates dramatically decrease across the entire ice sheet in Figures 2d–2h.

During 18–16 ka, the combined effect of reduced melt and localized increases in snowfall leads to SMB values consistent with ICE-6G. Although meltwater forcing still modifies sea ice extent during this interval (Figures 2a and 2b), the resulting reduction in surface melting for the iTRACE (blue regions in Figure 2d) is more spatially limited, primarily affecting the southeastern margin and Alaskan border margins in the northwest. Additionally, there is generally more snowfall occurring in the same southeastern margin (red regions in Figure 2a) and less melt in the iTRACE experiment.

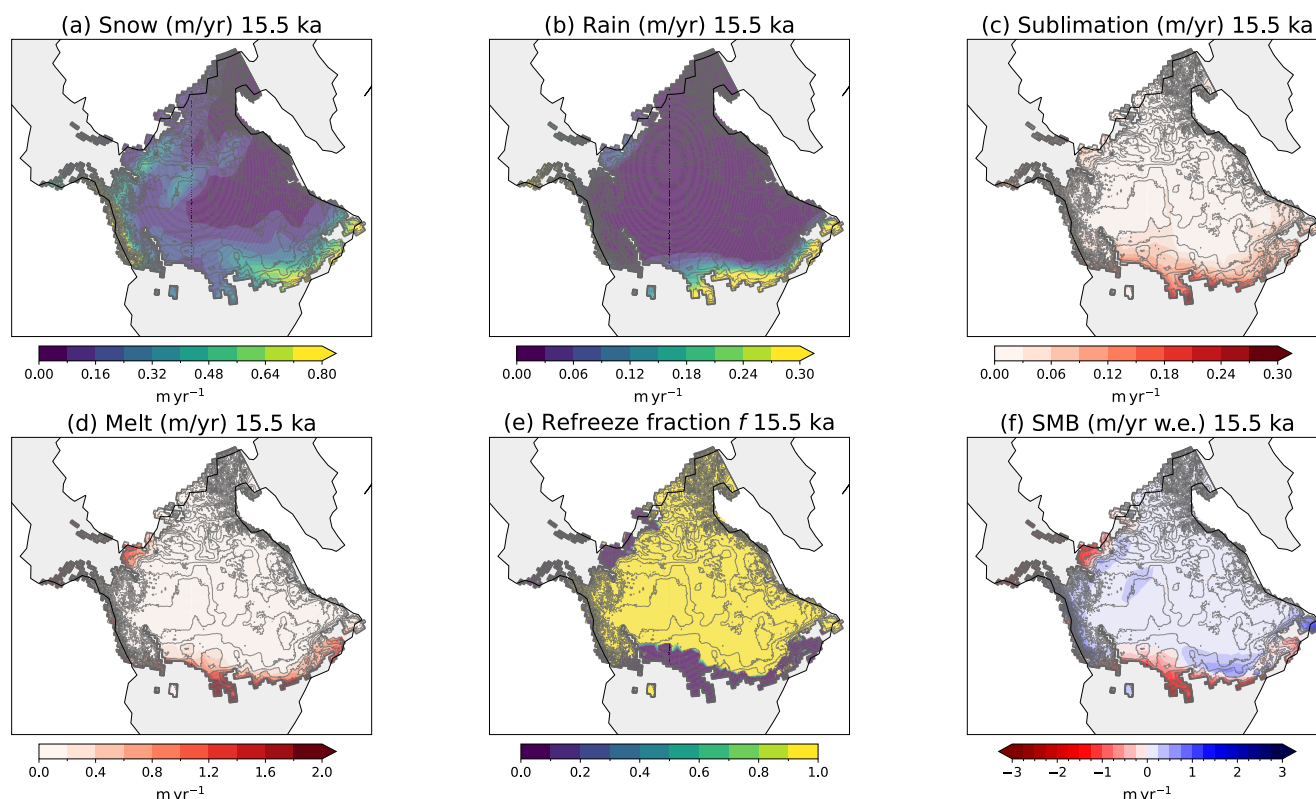


Figure 3. The surface mass balance (SMB) components in the iTRACE experiment are shown by color shading, and the ICE-6G ice volume is shown as contour lines (contour interval of 1,000 m). Each subplot is the century average, centered at 15.5 ka, of the respective variable at each grid cell containing ice. Panel (a) shows snowfall rate in m/yr w.e. , (b) rainfall, (c) sublimation, (d) melt as calculated by Equation 1, and (e) refreeze fraction, f , as calculated by Equation 2, and (f) SMB as calculated by Equation 3. Gray contours indicate the elevation classes (see Figure S8 in Supporting Information S1).

3.2. The Spatial Distribution of the SMB and Its Components

We next examine the spatial distribution of the different SMB components. Figure 3 shows the century-mean fields centered at 15.5 ka, for snowfall, rainfall, sublimation, melt, refreeze fraction, and SMB.

Snowfall is primarily concentrated along the western margin of the ice sheet due to orographically-induced precipitation, with additional accumulation along the southeastern margin (Figure 3a). In contrast, the interior of the ice sheet receives little snowfall, consistent with the elevation-desert effect, which forces air to rise releasing precipitation on the western (due to the jet stream) side of the ice sheet. Rainfall is limited to the southeastern margin of the ice sheet (Figure 3b), where the refreeze fraction is mostly zero (Figure 3e) due to the small snowfall rate there Equation 2.

Melt is restricted to a narrow margin in the southern margin of the ice sheet, as expected (Figure 3d), while sublimation remains relatively small across most regions (Figure 3c). The resulting SMB (Figure 3f) reflects the balance between strong marginal ablation and interior accumulation.

Note that for downscaled grid points falling within the highest elevation bin (3,000 to 10,000 m), we use the average elevation of these grid points rather than the bin value, to account for the high topography of the LIS which reaches 4.8 km.

The spatial evolution of the SMB calculated for the iTrace is shown in Figure 4, where each panel represents the century mean centered on the indicated time slice. Net ablation is consistently strongest along the southwestern margin of the LIS, while net accumulation dominates on the western margin until the split from the Cordillera. This pattern reflects persistent orographically induced precipitation sourced from the Pacific Ocean. The ICE, ICE + ORB, and ICE + GHG + ORB spatial SMB similar to Figure 4 are shown in Figures S12–S14 in Supporting Information S1 respectively.

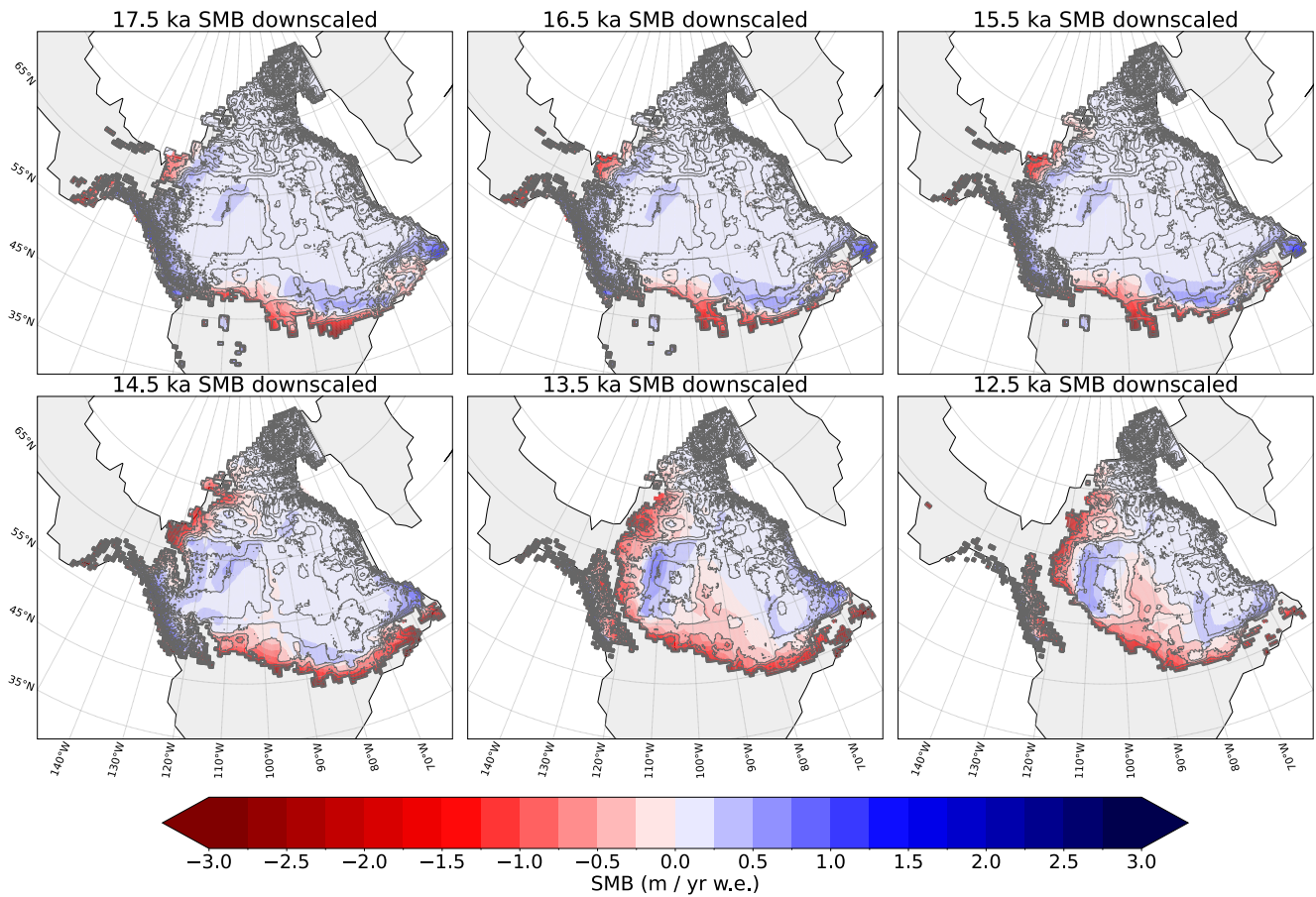


Figure 4. Contour plots of the downscaled surface mass balance (SMB) of the iTRACE experiment with paleo coastlines (gray; 100% land cells) at different time slices during deglaciation. Each subplot is the century average, centered at the time slice noted in the respective title, of SMB of each grid cell containing ice. The colored grid-cell contours describe the SMB, where red marks net ablation and blue marks regions of net accumulation rate. The same contour plots of the SMB for ICE, ICE + ORB, and ICE + GHG + ORB, and of the coarse resolution iTraCE results are shown in the Figures S12–S14 in Supporting Information S1. Gray contours indicate the elevation classes (see Figure S8 in Supporting Information S1).

The overall magnitude and spatial distribution shown in Figure 4 are comparable to previous simulations (e.g., Kapsch et al., 2021), although here we focus on the time-continuous evolution of the LIS and the comparison to the ICE-6G rather than on selected time slices.

3.3. Comparison to SEB Results

Finally, we briefly compare the SMB derived using a PDD-based melt scheme to that obtained using an SEB-based approach. In the SEB method, melt is calculated from the sum of the monthly surface radiative (net shortwave and longwave radiation) and turbulent fluxes (sensible, latent), and ground heat fluxes. These fluxes are then converted to a melt rate (mm w.e./s.) and then downscaled using the same elevation-class method applied to the PDD inputs. The vertical gradient used for downscaling the melt rate calculated from the SEB is of -425 mm/yr/km (Selleveid et al., 2019), and full details are provided in Supporting Information S1.

The SEB-based method produces melt in regions of the ice sheet where temperatures are well below freezing. This is illustrated in Figure 5a which shows that there is significant energy available for melt occurring at temperatures roughly between -20°C and -5°C at a represented interior grid point. As a result, the SEB-derived melt extends unrealistically far into the interior of the ice sheet (Figure 5b).

In contrast, along the ice sheet margins, the SEB-based method underestimates melt relative to the PDD scheme. This likely reflects the use of monthly-averaged fluxes (model outputs), which smooth short-lived but intense melt events and reduce peak melt rates. These two biases (spurious interior melt and underestimated marginal

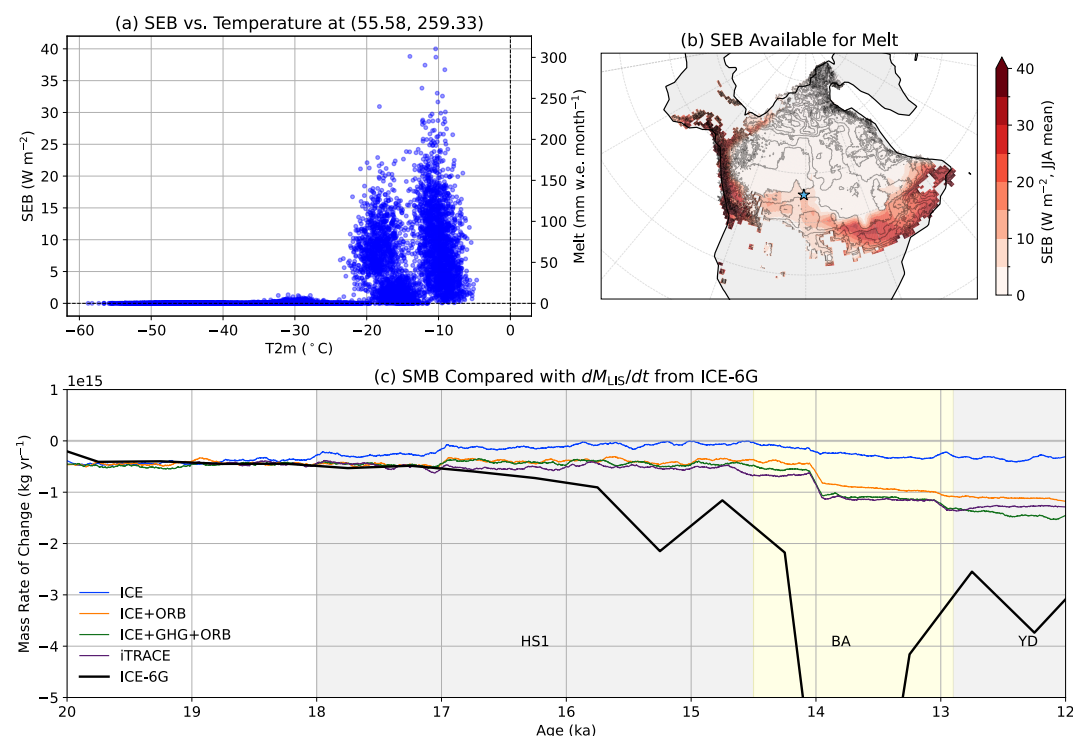


Figure 5. (a) Scatter plot of downscaled SEB versus downscaled 2-m temperature at (55.6, 259.3). (b) Contour plot of the SEB available for melting ($SEB > 0$) in W/m^2 . The blue star indicates the selected point of predicted interior melt that is shown in detail in (a) with gray contour lines indicating the elevation classes. (c) A similar plot to Figures 1a and 1b. Annual surface mass balance (SMB), downscaled to a 10 min grid, for the Laurentide Ice Sheet (LIS), integrated over the LIS surface and calculated using the SEB method. The integrated SMB is compared to the ice mass rate of change of the LIS of ICE-6G (black curve). The dark purple line depicts the smoothed annual SMB of the fully forced iTraCE experiment, blue: ICE, orange: ICE + ORB, green: ICE + GHG + ORB. The shaded regions are the same as in Figures 1a and 1b showing Heinrich Stadial 1, BA, and the Younger Dryas periods.

melt) arise from different aspects of the SEB implementation and together contribute to the discrepancies between the SEB- and PDD-derived SMB fields (see also Figure S11 in Supporting Information S1).

The integrated SMB time series derived using the SEB method (Figure 5c) differs substantially from that obtained using the PDD-scheme (Figures 1a and 1b), both in magnitude and temporal structure. In particular, the SEB-based SMB does not produce the same acceleration of mass loss during the BA and shows reduced sensitivity to the imposed forcings. These differences highlight limitations of the SEB approach in this application, particularly when applied to coarse-resolution, monthly averaged climate model outputs.

For these reasons, we emphasize the PDD-based results in the main analysis, while using the SEB calculations primarily for comparison.

4. Discussion

The comparison between SMB-derived mass loss and the ICE-6G reconstruction provides a test of whether the climate simulated by iTraCE is dynamically consistent with the prescribed ice sheet evolution. Overall, the SMB captures both the magnitude and temporal evolution of ice mass loss through much of the deglaciation, although notable discrepancies occur during specific intervals.

While the iTraCE simulation suite does not include a coupled dynamic ice model, published analyses demonstrate that it captures the principal isotopic signature and hydrological signals of the last deglaciation (e.g., Du et al., 2021; He et al., 2021; C. Zhu et al., 2022; Fu, 2023). Nevertheless, the TraCE/iTraCE model lineage is known to smooth abrupt regional transitions relative to proxy evidence, warranting caution when interpreting absolute SMB values derived from it. In addition, because iTraCE prescribes ice sheet geometries rather than

coupling it dynamically, any biases in atmospheric or oceanic processes do not feed back onto the ice sheet. Discrepancies between iTraCE and ICE-6G may therefore reflect known limitations in either the climate model or the ice reconstruction.

During 20–18 ka, the mass loss predicted by the simulated SMB exceeds that inferred from ICE-6G. Although the difference visually appears small in Figure 1a, summing the mass loss through time yields a cumulative SMB-derived loss of $\sim 1.02 \times 10^{18}$ kg, compared with $\sim 6.36 \times 10^{17}$ kg inferred from ICE-6G—an overestimate of approximately 60%. Thus, modest but persistent differences in instantaneous rates can produce substantial differences in cumulative ice loss. This highlights the importance of considering integrated mass loss, rather than instantaneous rates alone, when comparing SMB to reconstructed ice volume changes.

A key result of this study is the role of meltwater forcing in modulating SMB. The inclusion of meltwater forcing produces abrupt shifts in SMB, particularly at ~ 14.5 ka, and improves agreement with ICE-6G during both HS1 and the YD. This is consistent with a link to the expansion of sea ice and its role in reducing melt rates (Figure 2h), although further targeted experiments would be required to isolate the mechanism. When meltwater is included, in particular between 18 and 14.5 ka, the simulated SMB becomes consistent with ICE-6G, with remaining differences attributable to ice flow and calving not represented in SMB. On the one hand, this important role of sea ice is consistent with the previously proposed sea ice switch mechanism (Gildor & Tziperman, 2000) where sea ice switches the ice sheet from net accumulation to net ablation and vice versa. On the other, we find here that sea ice mostly affects the melt rate, while the SIS mechanism assumed its main effect to be on accumulation. This may provide an important input to further developments of a complete and self-consistent glacial cycle mechanism, which is still out of reach after decades of research.

Additionally, aside from brief periods that appear to respond to the initial ice sheet change in the iTraCE experimental design, the SMB remains negative over the entire deglacial period, in contrast to Ullman et al. (2015), who found a positive SMB until about 9 ka using a different model and reconstruction. Our results are more consistent with Bradley et al. (2024) who also found negative SMB at the LGM.

The period from 14 to 13.5 ka corresponds to the separation of the LIS from the Cordilleran Ice Sheet (Figure S8 in Supporting Information S1) and the onset of the BA (14.5–12.9 ka). This abrupt warming may have triggered enhanced ice flow and calving, marking a transition from SMB-dominated to dynamically dominated mass loss. While the negative SMB is consistent with overall ice sheet decay, it does not uniquely determine total mass loss. For example, the present-day GrIS exhibits a positive SMB alongside net mass loss (Mouginot et al., 2019; Shepherd et al., 2020). Enhanced surface melting often results in increased water injection into the base of the ice sheet via moulins (Banwell et al., 2016; Colgan & Steffen, 2009). This can accelerate the ice flow and calving, consistent with the fact that our calculation indicates stronger effects of ice flow and calving later in the deglaciation, when the SMB indicates more melting. While increased surface melting associated with negative SMB during a deglaciation may accelerate ice flow and enhance calving, this relationship is complex as at sufficiently high meltwater production, channelization of drainage can increase basal friction and ice flow deceleration (Schoof, 2010), with further different effects on a seasonal time scale in the ice sheet interior versus margins (Hewitt, 2013). These processes, along with internal ice dynamics and subglacial conditions (Golledge et al., 2009; Schoof, 2010; Williams et al., 2020) limit the ability of SMB alone to predict total mass loss.

There are several important caveats to consider. Ice sheet margins, where slopes and melt rates are highest, remain difficult to resolve in global climate models such as iTraCE. Although we downscale to a 10-min resolution, this limitation persists, as higher resolution was shown to improve agreement with regional climate models (Kapsch et al., 2021). As a result, the coarse resolution required for long simulations likely under-represents narrow ablation zones and steep SMB gradients, introducing biases in melt rates and warm-air advection through mis-represented gravity wave drag and the effects of surface roughness (Löffverström et al., 2020). Additionally, the SEB method uses monthly fluxes which may miss short-lived and strong melt events, and the PDD uses assumed daily downscaling which may be state dependent and depend on the temperature and configuration of the ice sheet (Sections 3.3). More broadly, SMB depends on prescribed ice geometry, and reconstructions such as ICE-6G carry significant uncertainties, especially during periods of rapid ice retreat (or advance), that propagate into climate-model-based SMB estimates (Kapsch et al., 2021). While an ELA diagnostic may provide a more geometry-independent metric for SMB, it is not available in ICE-6G and its use in an offline calculation as is described in this manuscript would require additional assumptions.

Additional uncertainties arise from process representation and boundary conditions. The refreeze parameterization (Colleoni et al., 2009), developed for the GrIS, may not fully apply to the LIS and does not explicitly resolve firm processes. Similarly, the land component used in iTraCE (Oleson et al., 2013, or CLM4.5) imposes a snow depth cap of 10,000 kg/m² that can limit more realistic refreeze schemes. Discrepancies with ICE-6G reflect biases in the climate model, the ICE-6G reconstruction, or both, and could be further evaluated using alternative ice histories (e.g., Lambeck et al., 2014; Tarasov et al., 2012) and coupled or offline ice sheet modeling.

5. Conclusion

In this study, we estimated the surface mass balance (SMB) of the Laurentide Ice Sheet (LIS) during the last deglaciation using downscaled climate model outputs from the iTraCE simulation suite (He et al., 2021) and compared it to the ice mass loss rate inferred from the ICE-6G reconstruction (Peltier et al., 2015).

We find that the SMB-derived mass loss exceeds that inferred from ICE-6G during 20–18 ka, corresponding to an approximately 60% overestimate in cumulative mass loss over this interval. After 18 ka, the SMB captures the evolution of ice mass loss more closely, with remaining differences attributable to processes such as ice flow and calving that are not represented in SMB alone. The inconsistency between 20 and 18 ka could be related to any number of errors, such as climate model biases, errors in ICE-6G, or choice of DDF or standard deviation of daily temperatures. After 18 ka, the SMB is less than the ICE-6G rate of mass loss and hence consistent with it, although the partition between ice flow and SMB in the total mass loss requires verification using explicit modeling of the ice flow.

A unique aspect of the iTraCE simulation suite is its factorial design, which isolates the incremental climate effects of evolving ice-sheet geometry, orbital parameters, greenhouse gases, and meltwater forcing. Among these forcings, orbital changes exert the largest influence on the SMB-derived trajectory of LIS mass loss during the last deglaciation (Figure 1b), consistent with Gregoire et al. (2015) (see also Figure S15 in Supporting Information S1). Increasing greenhouse gases further amplify mass loss, particularly after 17 ka (Abe-Ouchi et al., 2013; Gildor & Tziperman, 2000; Koepnick & Tziperman, 2024). Meltwater forcing, in contrast, introduces abrupt SMB variability by altering sea ice and, consequently, surface melt. Its inclusion improves agreement with ICE-6G during HS1 and the YD, highlighting the importance of sea-ice feedbacks in modulating deglacial SMB (Gildor & Tziperman, 2000). During the BA, strongly negative SMB indicates enhanced surface melting that may also have contributed to ice-flow acceleration and dynamic mass loss.

Despite these limitations, we find that the iTRACE SMB broadly captures the magnitude of deglacial ice loss estimated by ICE-6G. The enhanced melt during the BA may promote basal lubrication and increased ice flow, although sufficient meltwater penetration causes channelization which can slow down the ice flow (Hewitt, 2013; Schoof, 2010). Overall, our results demonstrate that SMB derived from transient climate simulations can reproduce key features of the deglacial ice sheet evolution when meltwater forcing is included. These findings underscore the importance of both surface processes and dynamic ice sheet responses in shaping past ice sheet change, with implications for understanding how ice sheets may respond to a warming climate.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Availability Statement

The iTraCE output model data is available on the NCAR Casper at /glade/campaign/cesm/collections/iTRACE and from the Climate Data Gateway at NCAR at Otto-Bliesner et al. (2021). The data was downloaded December 2024. The ICE-6G data is also available online at <https://www.atmosphysics.utoronto.ca/~peltier/data.php>.

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