De-Tuning Albedo parameters in a coupled climate ice sheet model to simulate the North American ice sheet at the last glacial maximum

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Key Points:

- Simulating the Last Glacial Maximum North American Ice Sheets successfully requires de-tuning the coupled Climate-Ice Sheet model FAMOUS-ice
- After running a large ensemble, we identify multiple acceptable simulations of the LGM North American Ice Sheet
- The LGM North American cloud cover revealed overtuned albedo parameters to compensate for modern Greenland cloud biases

Supporting Information:

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

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De-Tuning Albedo Parameters in a Coupled Climate Ice Sheet Model to Simulate the North American Ice Sheet at the Last Glacial Maximum

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Abstract The Last Glacial Maximum extent of the North American Ice Sheets is well constrained empirically but has proven to be challenging to simulate with coupled Climate-Ice Sheet models. Coupled Climate-Ice Sheet models are often too computationally expensive to sufficiently explore uncertainty in input parameters, and it is unlikely that values calibrated to reproduce modern ice sheets will reproduce the known extent of the ice at the Last Glacial Maximum. To address this, we run an ensemble with a coupled Climate-Ice Sheet model (FAMOUS-ice), simulating the final stages of growth of the last North American Ice Sheets' maximum extent. Using this large ensemble approach, we explore the influence of numerous uncertain ice sheet albedo, ice sheet dynamics, atmospheric, and oceanic parameters on the ice sheet extent. We find that ice sheet albedo parameters determine the majority of uncertainty when simulating the Last Glacial Maximum North American Ice Sheets. Importantly, different albedo parameters are needed to produce a good match to the Last Glacial Maximum North American Ice Sheets than have previously been used to model the contemporary Greenland Ice Sheet due to differences in cloud cover over ablation zones. Thus, calibrating coupled climate-ice sheet models on one ice sheet may produce strong biases when the model is applied to a new domain.

Plain Language Summary At the peak of the last ice age, an ice sheet covered much of North America. The extent of this ice sheet is well-understood after decades of intensive data collection, but producing a computer simulation of the ice sheet which matches our observations has been a challenge. This is partly because of uncertainty about the "correct" model set-up to create the best simulation, and partly because the computer models used in the simulations require large computing resources. In this paper, we present a series of simulations of the North American Ice Sheet at the peak of the last ice age using a fast-running computer model in which the atmosphere and ice sheets interact. We run hundreds of simulations to tackle the uncertainty about the optimum values for unknown input parameters. We find that the model's representation of how reflective the ice sheet surface is has the most impact on the size and shape of the simulated ice sheet. Importantly, the parameter values that produce the best simulations of modern-day Greenland produce poor simulations of the North American Ice Sheets during the last ice age, calling into question whether the parameters chosen for modern Greenland are appropriate from simulating ice sheets under different conditions.

1. Introduction

Accurately estimating future changes in ice sheets is crucial for producing meaningful projections of future sea level rise (IPCC, 2021). Ice sheets interact with the atmosphere and ocean, and are vulnerable to instabilities in their growth and retreat (e.g., Gregory et al., 2012; Shepherd et al., 2012). These instabilities, along with the uncertainties in the processes of climate and ice sheet evolution, make future projections using numerical models difficult, and the accuracy of future simulations is a challenge to assess. For the Greenland Ice Sheet, one of the main sources of uncertainty is the future changes in Surface Mass Balance (SMB, the balance of accumulation and total runoff) (Fettweis et al., 2011). SMB is highly dependent on both the climate and the ice sheet topography, as well as the strong interactions between the two. Thus, projections of future Greenland evolution need to account for climate-ice sheet interactions (Goelzer et al., 2017). However, there are major challenges in representing SMB and climate ice sheet interactions in models: (a) climate models often have large biases in ice sheet

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regions (Davy & Outten, 2020), (b) these regions are difficult environments to work in, limiting the observations we have of the climate and SMB (Vernon et al., 2013), and (c) surface melt occurs in narrow steep regions at the edge of the ice sheets that are difficult to represent in global climate models.

Major progress has been made to tackle these challenges through a combination of model development and increased computational power, and there are now several Earth System Models that include interactive ice sheets in Greenland and/or Antarctica (e.g., Muntjewerf et al., 2021; R. S. Smith et al., 2021). However, there are numerous challenges in simulating climate-ice sheet interactions. Perhaps most acutely, there is a mismatch of spatial and temporal scales between typical ice sheet and global climate models. Spatially, kilometer (or sub-kilometer) processes are important for accurate simulation of ice sheet processes, such as grounding line migration, and margin precipitation gradients (Cornford et al., 2013; Franco et al., 2012), so recent ice sheet models have been developed to simulate ice sheets at this scale, for example, through adaptive mesh refinement, and they require climate (or SMB inputs) at that scale. On the other hand, Atmosphere Ocean General Circulation Models (AOGCMs) have a grid-box size 10–1,000× larger than the scale of modeled ice sheet processes and inputs (Danabasoglu et al., 2020; Sellar et al., 2019). This conundrum of mismatching scales is flipped in the time domain. Temporally, AOGCMs require sub-daily timesteps to accurately simulate the climate system, while ice sheet changes occur over several centuries or millennia. The spatial-temporal mismatch of scales creates a problem for computational efficiency, since high resolution is needed for both and is not easily compromised for one in favor of the other. Consequently, most AOGCMs cannot practically simulate interactive ice sheet change, instead prescribing the ice sheet extent as a boundary condition (e.g., Ivanovic et al., 2015; Kageyama et al., 2017; Menviel et al., 2019) and updating the ice sheet periodically for palaeo runs where significant ice sheet change occurs. Similarly, ice sheet simulations often rely on prescribed climate or SMB fields (e.g., Gandy et al., 2021; Gregoire et al., 2016; Patton et al., 2013). Uncoupled run like this does not fully capture the complex and potentially important climate-ice sheet interactions. These interactions may be important to correctly simulating deglaciation of the past and future, so it is important that progress on these simulations is pursued.

The spatial mismatch between ice sheet and climate model grids can be addressed by calculating ice-sheet relevant processes, such as albedo calculations, with sub-gridscale parameterizations (Ganopolski et al., 2010; Sellevold et al., 2019; R. S. Smith et al., 2020, 2021; Vizcaíno et al., 2013; F. A. Ziemen et al., 2014). For example, in FAMOUS-ice the spatial mismatch between the ice sheet and climate model is accounted for by calculating ice fractions on coarse resolution tiles (R. S. Smith et al., 2021). The model also uses a number of parameterizations to simulate cloud formation, including thresholds for humidity leading to cloud formation, and cloud liquid water leading to precipitation (Murphy et al., 2004). One way to solve the problem of time scale (sub-daily timesteps for multimillennial integrations) is to couple the ice sheet to climate models which are typically computationally efficient, in part due to having relatively low spatial resolution, meaning that the simulations can be run for the length of time needed to spin-up and simulate the co-evolution of ice sheets and climate.

A remaining challenge is how coupled climate-ice sheet models should be calibrated and tested. For many models, uncertain parameters are hand-tuned to produce stable modern ice sheets of the right shape and size compared to observations. However, this only represents one point in time. The recent past, for which we have direct observations of ice sheets, has seen relatively small changes compared to those expected in the next centuries. These observations thus provide poor constraints on the strength of climate-ice sheet feedbacks and there is a danger of over-fitting the model to modern conditions and compensating for biases in the simulated climate. To have confidence in the ability of coupled climate-ice sheet models, we need to test them under conditions different from today where we have sufficient observational constraints on the climate and ice sheets.

We propose to use the Last Glacial Maximum as a benchmark for coupled climate-ice sheet models. This period, which occurred around 21,000 years ago, has been a focus of the Palaeoclimate Model Intercomparison Project since the 1990s (Kageyama et al., 2021a) because it was a period of relatively stable and well-documented climate, with CO_2 concentrations much lower than today (around 180 ppm). During this time period, the North American Ice Sheet (the Laurentide, Cordilleran, and Innuitian Ice Sheets collectively) is thought to have reached a relatively stable maximum extent, that is very well reconstructed (e.g., Dyke et al., 2002; Gowan et al., 2021; Peltier et al., 2015). It is thus possible to run equilibrium simulations under LGM conditions with an interactive North American Ice Sheet until a stable maximum ice extent is reached, which can be meaningfully compared to reconstructions. It has been proposed that the North American Ice Sheets were not at steady state at the LGM (F. Ziemen et al., 2014), in which case the LGM reconstruction can be used as a minimum target extent for a simulation to achieve, rather than an absolute target. During the LGM, the ice sheet extent reached much lower latitudes than today, providing a way to test the ability of models to represent SMB and climate-ice interactions under energy balance conditions different than modern Greenland.

We use FAMOUS-ice (R. S. Smith et al., 2020), a coarse resolution, fast running AOGCM, which has been used in long-palaeo simulations (Dentith et al., 2019; Gregoire et al., 2012; Gregory et al., 2012; Roberts et al., 2014; R. S. Smith, 2012) and uncertainty quantification (Gregoire et al., 2010). Rather than previous work using a PDD SMB scheme (Gregory et al., 2012), FAMOUS-ice is coupled to the Glimmer Ice Sheet model by downscaling SMB calculations (R. S. Smith et al., 2020). This coupled model has been used to simulate present and future Greenland Ice Sheet evolution (Gregory et al., 2020). We start the manuscript by presenting the first attempt at simulating the LGM with the FAMOUS-ice model with interactive ice sheets in Greenland and North America. Here, we use an atmosphere-only version of FAMOUS, prescribing Sea Surface Temperatures (SSTs) and Sea Ice Concentrations (SICs) in order to minimize biases in surface climate.

2. Model Description and Setup

FAMOUS-ice is a fast climate model coupled to the Glimmer ice sheet model (R. S. Smith et al., 2020). The atmospheric component is the same as in FAMOUS, a fast low resolution general circulation model designed for running simulations of the climate on multi-millennial timescales (e.g., Dentith et al., 2019; Gregoire et al., 2012) and large ensembles for calibration or uncertainty quantification purposes (Gregoire et al., 2010). The atmosphere in FAMOUS is a configuration of version 4.5 of the UK Met Office Unified Model (Gordon et al., 2000). Based on the primitive equations in contains a full suite of prognostic atmospheric physics parameterizations appropriate to a state-of-the-art climate model of that era, and runs without flux adjustments. Our configuration has a horizontal resolution of $7.5° \times 5°$, with 11 vertical levels that extend just above the tropical tropopause. A full description of UM4.5 and FAMOUS is available in previous work (Gordon et al., 2000; Jones et al., 2005; Pope et al., 2000; R. S. Smith et al., 2008). The Glimmer ice sheet model is a fast simplified 3D dynamical ice sheet model based on the shallow ice approximation that is used to simulate continental ice sheets over glacial-interglacial cycles (Gregoire et al., 2015; Rutt et al., 2009). FAMOUS can also be used with a dynamical ocean (e.g., Dentith et al., 2019) and the Glimmer ice sheet component can be replaced with the more complex and computationally expensive BISICLES ice sheet model (Cornford et al., 2013; Gandy et al., 2018; Matero et al., 2020).

Ice is one of the nine surface types simulated by FAMOUS. This is for the purpose of simulating climate and does not include dynamic ice sheet processes such as deformation or calving. Every individual grid cell can host a combination of these types, with each type being assigned a fraction of the total grid cell area (R. S. Smith et al., 2020). In practice, this means that the atmosphere at the ice sheet edge is simulated considering a mix of ice and non-ice land types. A multilayer snow scheme is used in the land model to calculate SMB at 10 different elevation levels within each grid cell that contains part of an ice sheet. This scheme simulates the density and temperature gradients within the snowpack. The SMB calculated by the land model is regridded from the coarse FAMOUS-ice grid (7.5°longitude by 5°latitude) onto the surface of the Glimmer ice sheet model (in this case 40×40 km) each model year. The use of 10 elevation levels on which SMB is calculated in every FAMOUS grid cell provides a mean to effectively downscale the SMB from the coarse atmospheric grid to the finer ice sheet grid. The albedo of ice sheet surfaces in FAMOUS-ice is a prognostic quantity calculated using the age, density, and temperature of the snow or ice at the surface, with a range of simplified regimes appropriate for scattering from fresh snow grains through to melt ponding on bare ice. Full details of the ice coupling and albedo approach are described in R. S. Smith et al. (2020). In summary, coupled simulations are completed by repeating the following steps;

1. The SMB field is simulated within FAMOUS
2. The SMB field is interpolated onto the Glimmer grid
3. The ice sheet evolution is simulated by Glimmer
4. Glimmer passes updated orography and ice cover fields back to FAMOUS

Coupled climate-ice sheet simulations would usually be too computationally expensive to run as part of large multi-millennial ensembles, but the speed of both FAMOUS and Glimmer make these experiments possible, with 500 years simulated in around 150 core hours. During the development of FAMOUS-ice, albedo parameters were manually tuned to simulate a stable Greenland Ice Sheet at present day (R. S. Smith et al., 2020). The model has been applied to evaluate the long-term future decline of the Greenland Ice Sheet (Gregory et al., 2020). In both cases, SSTs and SICs in FAMOUS were prescribed from the output of higher resolution and complexity climate models to allow some control of the climate evolution and reduce the impact of biases resulting from atmosphere-ocean and sea ice interactions.

We setup FAMOUS-ice to simulate the climate of the Last Glacial Maximum with interactive North American and Greenland Ice Sheets. We follow the PMIP4 LGM protocol (Kageyama et al., 2017) to set up most of the climate boundary conditions, including the CO_2, CH_4, and N_2O concentrations. Global orography and the land-sea mask were taken from the 21 ka BP Glac-1D reconstruction (Tarasov et al., 2012). We do not include dynamical vegetation in our model. Instead, we prescribe a preindustrial (PI) vegetation distribution which sets the land surface properties where there is no ice. Thus, our simulation neglects the effects of climate-ice-vegetation feedbacks that can affect ice sheet evolution (Stone & Lunt, 2013), and previous simulations have shown that alterations to the prescribed vegetation distribution can have significant climate and ice sheet impacts (Sommers et al., 2021). However, as ice advances, the fraction of land ice within the grid cell increases and the fractions of other land surface types decrease proportionally, thus increasing the mean albedo of the grid cell (and vice versa when ice retreats). The orbital configuration is set to 23 ka BP, rather than 21 ka BP as in the PMIP4 protocol. This is 2,000 years prior to the ice sheet maximum extent to represent an orbit closer to the point of maximum volume for the North American Ice Sheet (Peltier et al., 2015; Tarasov et al., 2012). This results in a slightly higher (2 Wm^2) summer insolation at 65°N, slightly inhibiting ice sheet growth as shown by a sensitivity experiment included in Section S2 in Supporting Information S1. We use the Glimmer Isostasy model, simulating an elastic lithosphere floating on viscous asthenosphere. A lower mantle relaxation time value results in a less viscous asthenosphere, and hence a quicker topographic response to ice sheet growth and decay.

SSTs and SICs are taken from the statistical reconstruction of Astfalck et al. (2021), combining information from the PMIP LGM multi-model ensemble (Kageyama et al., 2021b) and compilations of proxy data (Kucera et al., 2005), and their associated uncertainties. The method is able to generate ensembles of plausible SST and SIC pairs that can be used to drive atmosphere models. The simulated SSTs are in good agreement with the proxy-based reconstruction of Paul et al. (2021) with significantly warmer tropical SSTs than the data assimilation product of Tierney et al. (2020).

The interactive ice sheet model domain is set to cover North America, Greenland, and Iceland. All other ice sheets are fixed to match the Glac-1D reconstruction. The Glimmer initial condition is taken from a previous ensemble of North American Ice Sheet deglaciations (Gregoire et al., 2016); specifically, ensemble member *Cano3-022* at 18.2 ka BP. This was chosen to represent an intermediate sized ice sheet resembling the likely extent during Marine Isotope Stage 3 (Gowan et al., 2021), from which to grow the North American Ice Sheet to an equilibrium ice sheet volume. More details on the initial ice sheet extent and the sensitivity of the model to this are provided in Sections S1 and S5 in Supporting Information S1. In FAMOUS-ice, ice is able to grow by flowing onto a gridcell not previously covered in ice, but ice is not able to form from the accumulation of snow into an unglaciated gridcell. We thus chose an initial condition with a Cordilleran Ice Sheet. We also chose to start with ice already covering the Hudson Bay as the Glimmer ice sheet model does not represent the complex processes of grounding line migration and may not simulate the necessary marine instabilities of advance and retreat (Pattyn et al., 2012). If we were simulating the ice sheet retreat phase, or a largely marine-grounded ice sheet (such as the Eurasian Ice Sheet or the West Antarctic Ice Sheet), it would be important to use an ice sheet model with advanced physics required to simulate grounding line migration, such as BISICLES (Cornford et al., 2013). Furthermore, Glimmer is setup to calve ice when the water depth exceeds 200 m. This simplified representation of calving and the absence of grounding line migration processes in Glimmer could lead to errors in the position of the marine margins simulated to the North and East of the North American Ice Sheets. However, in this study, we focus on constraining processes of SMB that control the position of the well-documented land-terminating margins (Dyke et al., 2002).

Simulations are run with 10× ice sheet acceleration; that is, for every climate year simulated with FAMOUS, the resulting SMB is used to force the ice sheet for 10 years with Glimmer. After this, the new Glimmer ice sheet surface elevation is passed back to the climate model, regridded and processed to update its orography and the land ice and vegetation fraction fields. Running the simulations with 10× ice sheet acceleration significantly reduces computational cost. Gregory et al. (2020) tested the effect of ice sheet acceleration using FAMOUS-ice and found no significant difference in ice sheet evolution under a stable climate forcing.

Table 1
Values of Key Climate and Ice Sheet Parameters in the Standard Setup Alongside the Ranges Sampled in Our Ensembles of Simulations Described in Section 4

Parameter	Standard value	Ensemble range	Units	Notes
Daice	−0.35	−0.4–0	K^{-1}	Sensitivity of bare-ice albedo to surface air temperatures when it is above a melt threshold to represent darkening of the surface due to summer melt ponds
AVGR	0.007	0–0.01	μm^{-1}	Sensitivity of the surface snow albedo to variation in grain size
Fsnow	600	350–800	$kg\ m^{-3}$	The threshold in snow density above which albedo darkens toward bare ice values
Alpham	0.2152	0.2–0.65		The sea ice low albedo (Crossley & Roberts, 1995).
Flow factor	3	1–10		Tuneable enhancement factor in Glen's flow law (Rutt et al., 2009). Increasing the factor makes the ice softer and more deformable
Mantle relaxation time	3,000	500–9,000	yr	The relaxation time of the mantle, a lower value essentially makes the mantle less viscous, thus allowing a quicker topographic rebound
Basal sliding	10	0.5–20	$mm\ yr^{-1}\ Pa^{-1}$	The basal sliding rate. A higher value allows increased ice velocity
RHCRIT	0.85	0.6–0.9		The threshold of relative humidity above which clouds form (R. N. B. Smith, 1990)
VF1	1.882	1–2	$m\ s^{-1}$	The precipitating ice fall-out speed (Heymsfield, 1977)
CT	0.000302	5×10^{-5}–4×10^{-4}	s^{-1}	The conversion rate of cloud liquid water droplets to precipitation (R. N. B. Smith, 1990)
CW	0.001688	0.0001–0.002	$kg\ m^{-3}$	The threshold values of cloud liquid water for formation of precipitation (R. N. B. Smith, 1990). Only the value for the land is varied
Entrainment rate	3	1.5–6		Entrainment rate coefficient Convection Scales rate of mixing between environmental air and convective plume
Lapse rate	6	2–10	$K\ km^{-1}$	Prescribed lapse rate for air temperature used to downscale FAMOUS near-surface ice sheet climate onto surface elevation tiles. Downwelling longwave radiation is also adjusted consistently

Note. The parameter values for all simulations are available in Figure S4 in Supporting Information S1.

We first run an initial standard experiment, using the atmosphere model parameters from simulations that produce a stable contemporary Greenland Ice Sheet (Gregory et al., 2020) (apart from boundary conditions altered according to the PMIP4 LGM protocol), and ice sheet model parameters from previous simulations of the North American Ice Sheet with Glimmer (Gregoire et al., 2016), also using the same spatially variable bed softness map (see standard parameter values in Table 1).

3. Collapse of the LGM Ice Sheet With a Standard Setup

In a simulation with the standard parameter values from (R. S. Smith et al., 2020), instead of growing from a mid-glacial ice sheet size, the North American Ice Sheet rapidly deglaciates in our standard experiment, losing half its volume in 2,500 ice sheet years, and continuing to deglaciate when the simulation finishes at 4,000 ice sheet years. This eventually results in a simulation with LGM climate conditions, but no North American Ice Sheet (Figure 1). The deglaciation is driven by ablation across the North American Ice Sheet, including the ice

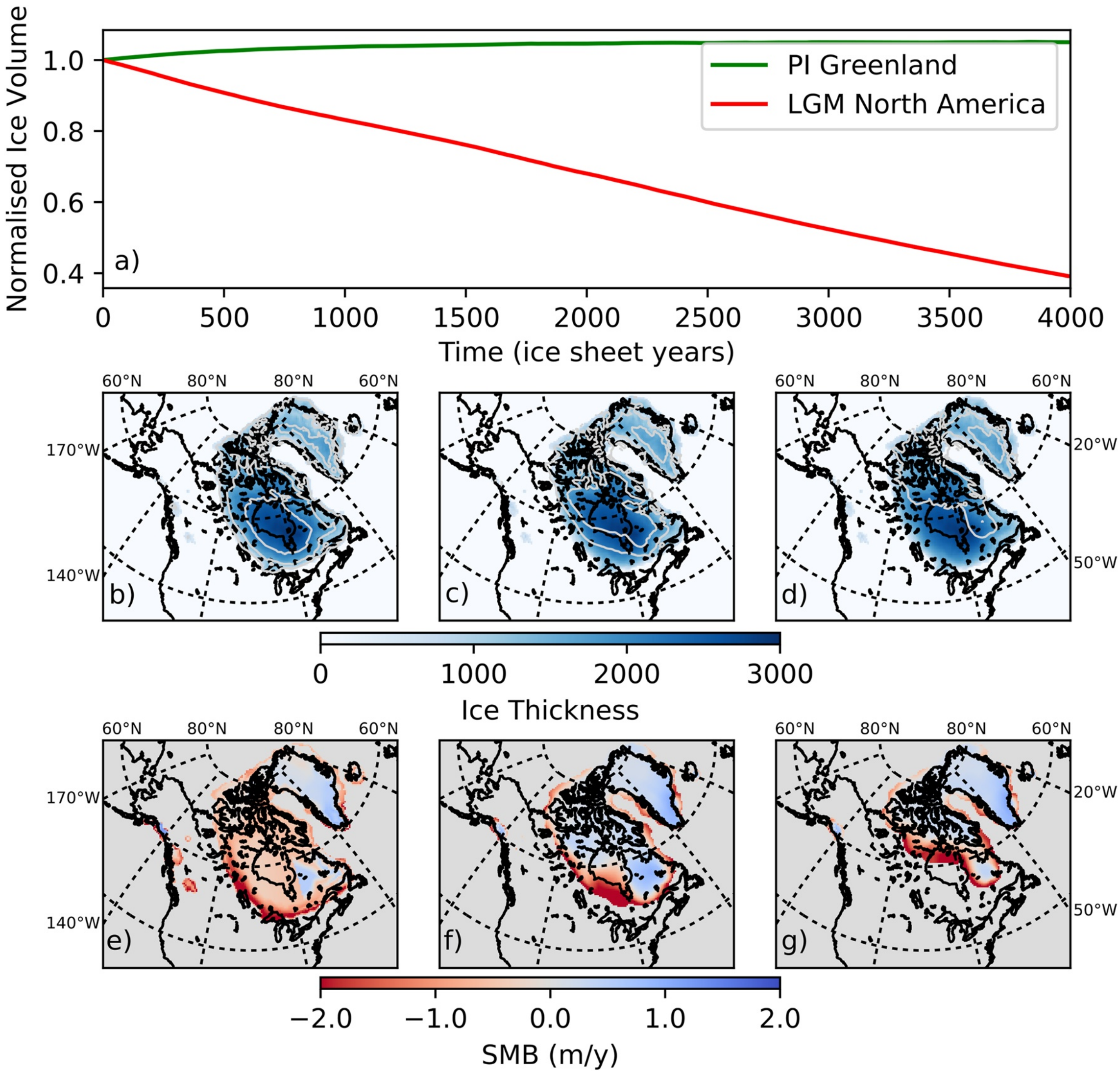

Figure 1. LGM North American Ice Sheet evolution in the standard setup. (a) The ice sheet volume, normalized by the initial volume (green line) compared to the PI Greenland simulations (red line). (b–d) Ice Thickness at 0, 2,000, and 4,000 years into the run, with gray contours at 1,000, 2,000, and 3,000 m. (e–g) The Surface Mass Balance at 0, 2,000, and 4,000 years into the run.

sheet interior (Figure 1e), which occur from the start of the simulation and causes a rapid retreat of the southern margin northward through Hudson Bay (Figures 1b–1d). The Greenland Ice Sheet on the other hand maintains its initial extent, which corresponds to full glacial conditions, in good agreement with observations (Simpson et al., 2009).

We know from geologic constraints (Dyke et al., 2002) that the ice sheet should be considerably larger than simulated here (i.e., it should cover the whole of Canada). We therefore conclude that parameters previously tuned to simulate the present day Greenland Ice Sheet well (as in R. S. Smith et al. (2020)) are not suitable for the LGM North American Ice Sheet.

4. The Ensemble Approach

To produce a reasonable simulation of the Last Glacial Maximum, we thus need to fully explore the uncertainty in model input parameters controlling the SMB, ice sheet dynamics and climatic conditions over the ice sheets in FAMOUS-ice as described below, in essence to "de-tune" the model and find parameter combinations that produce good representations of the LGM North American Ice Sheet. We identified 13 parameters (detailed in

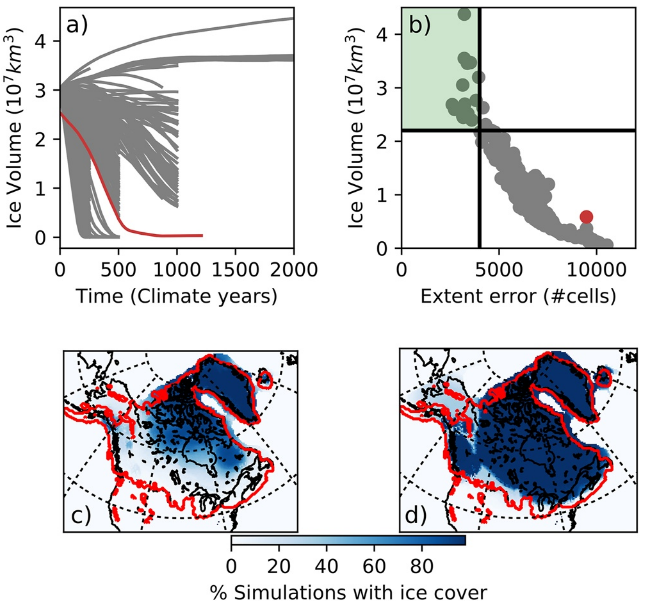

Figure 2. (a) North American and Greenland Ice volume evolution for each ensemble member. (b) The final ice volume and extent error (compared to the Dyke et al. (2002) margin) for each ensemble member. The accepted criteria are set at over $2.1 \times 10^7 \text{km}^3$ ice volume (58.2 m SLE), and less than 4,000 grid cells of error (6,400,000 km^2). (c) The % of ensemble simulations with ice cover compared to the Dyke et al. (2002) margin shown in red. (d) The % of Not Ruled Out simulations with ice cover compared to the Dyke et al. (2002) margin shown in red. The red line and point on panels (a, b) show the control run, as shown in Figure 1.

Table 1) that are the most likely to influence ice sheet size. Some of these parameters directly control the ice sheet SMB in the FAMOUS-ice coupling scheme with three ice albedo parameters and one lapse rate parameter for downscaling. A further six climate parameters were chosen for their influence on energy balance and precipitation. And finally three ice parameters were chosen for their control on ice sheet flow and isostatic rebound. The range of values for the climate and ice sheet parameters were derived from previous uncertainty quantification work with FAMOUS (Gregoire et al., 2010) and Glimmer (Gregoire et al., 2016), with the addition of the Entrainment rate coefficient varied in Murphy et al. (2004). We purposefully chose wide but plausible ranges (Table 1) with the aim to identify a region of the parameter space that would produce a reasonable North American ice extent at the LGM.

We started by running a large 280-member ensemble of simulations, sampling the full range of parameter values from Table 1. These values were sampled using Latin Hypercube sampling (Williamson, 2015), which varies all parameter values in tandem, to efficiently explore the parameter space by avoiding repeated parameter values. The number of simulations was constrained by the computational budget; the Latin Hypercube achieves optimal spacing across the parameter space, with the expectation that only a limited number of parameters will be influential. We expected that similar to the standard simulation (Section 3), many runs would fully deglaciate. Thus, to optimize our use of computing resources, simulations were stopped at 2,500, 5,000, and 10,000 ice sheet years if they lost more than 25% of the initial ice sheet area.

In the majority of the 280 ensemble members, the whole domain volume (including the North American and Greenland Ice Sheets) reduces dramatically and thus all but 50 simulations were terminated at or before 5,000 ice sheet years (Figure 2a). However, five simulations remain relatively stable at the initial volume and area, and ice volume grows in 13 simulations. Of these 18 simulations, there is a variety of ice configurations but some consistent model-data mismatch. All the ice sheets that grow from the initial extent include extensive ice in Alaska, which was mostly ice free at the LGM (Dyke et al., 2002). The remaining 32 simulations that continued past 5,000 ice sheet years were slowly deglaciating, and not yet reached the threshold for termination. These simulations were stopped at 10,000 ice sheet years. The deglaciating simulations could be caused by errors in the simulated climate. We thus compared the ensemble results to prior simulations from PMIP3 and PMIP4 (Figure 3). We find that global mean temperature is broadly in line with previous PMIP3 and PMIP4 simulations—on the warmer end around 284 K. Previous simulations have shown that warmer PMIP3 LGM models do not produce an extensive ice sheet when forcing a stand-alone ice sheet model (Niu et al., 2019), which could explain the deglaciations in the FAMOUS-ice ensemble. However, the ensemble mean North American summer surface temperature is similar to many PMIP3 simulations, accounting for the smaller initial ice sheet in the FAMOUS-ice simulations. A visual comparison of the mean summer surface temperatures of the ensemble and PMIP3 simulations is included in Section S7 in Supporting Information S1. The range of global mean temperatures in the ensemble is limited by the prescribed SSTs, which have variability in the temperatures spatial distribution, but the same global mean SST (within 0.5 K). The global precipitation pattern is also reasonable despite the cloud parameters varied (Figure S3 in Supporting Information S1).

To assess our ice sheet results, we compare them to the ice extent reconstructed by Dyke et al. (2002). We calculate ice extent error as in Gregoire et al. (2016) by summing up the number of gridcells where the ice does not match the reconstruction. The maximum allowed error extent was chosen to align with the maximum LGM extent error from the Not Ruled Out Yet (NROY) ensemble of Gregoire et al. (2016). Gregoire et al. (2016) applied a cumulative extent error over the whole deglaciation to identify their NROY set of simulations; we translated this into a maximum bound for our LGM extent error metric by applying our metric to their final NROY set and identifying the maximum value obtained. Constraints on North American ice volume are not as well known as ice extent, but can provide a useful metric for ruling out simulations. We chose to set a minimum threshold of $2.1 \times 10^7 \text{km}^3$

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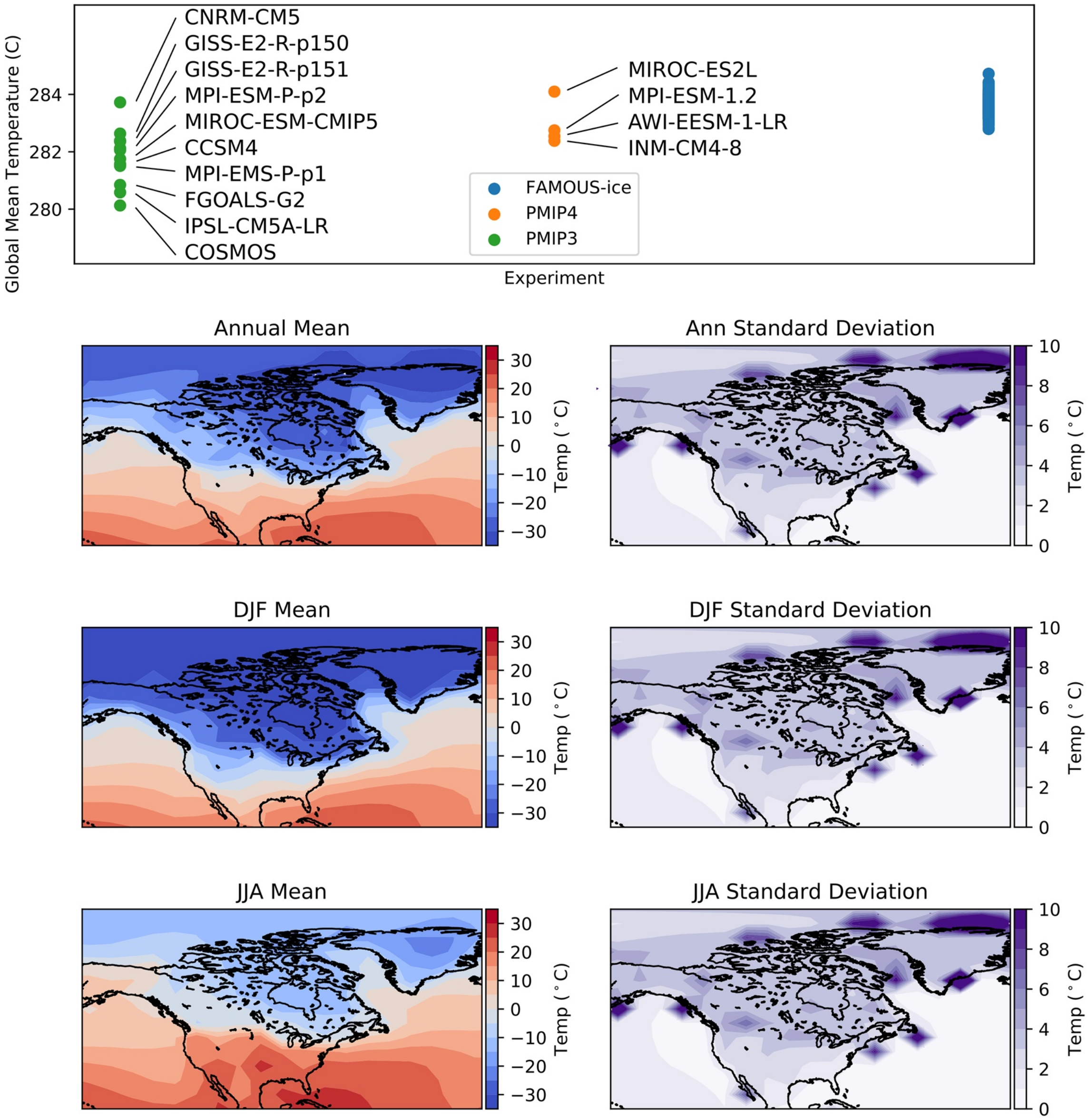

Figure 3. The surface temperature climatology of the LGM ensemble compared to previous simulations as part of PMIP3 and PMIP4. Standard deviation is calculated from the ensemble mean.

(58.2 m SLE) for ice volume as in Gregoire et al. (2016), based on a variety of individual reconstructions (Clark & Tarasov, 2014; Lambeck et al., 2014; Peltier et al., 2015; Tarasov et al., 2012). Only a small subset of the simulations (18 out of 280) terminate the simulation within this accepted criterion for volume and extent, highlighted by the green box in Figure 2b. We refer to the parameter values of these simulations as Not Ruled Out (NRO). The parameter values for these simulations are realistic in that the ensemble range for the parameters is wide but plausible (Table 1). However, the albedo parameter values are distinct from previous simulations of Greenland (Gregory et al., 2020), and further Greenland simulations would be required to identify a region of the parameter space that produces reasonable simulations for both LGM North American and Pre-industrial Greenland.

The mean ice sheet extent of the NRO simulations (Figure 2d) is close to the southern ice sheet margin. However, simulations that meet this margin still show some consistent model-data mismatch. All simulations with a low North American Ice Sheet extent error have ice that is too extensive in Alaska. This is also common in ice sheet-only simulations with Glimmer (Gregoire et al., 2016; Ji et al., 2021) driven by climate forcing from FAMOUS and from the higher resolution CCSM3 model, so is likely a systematic bias resulting from the climate

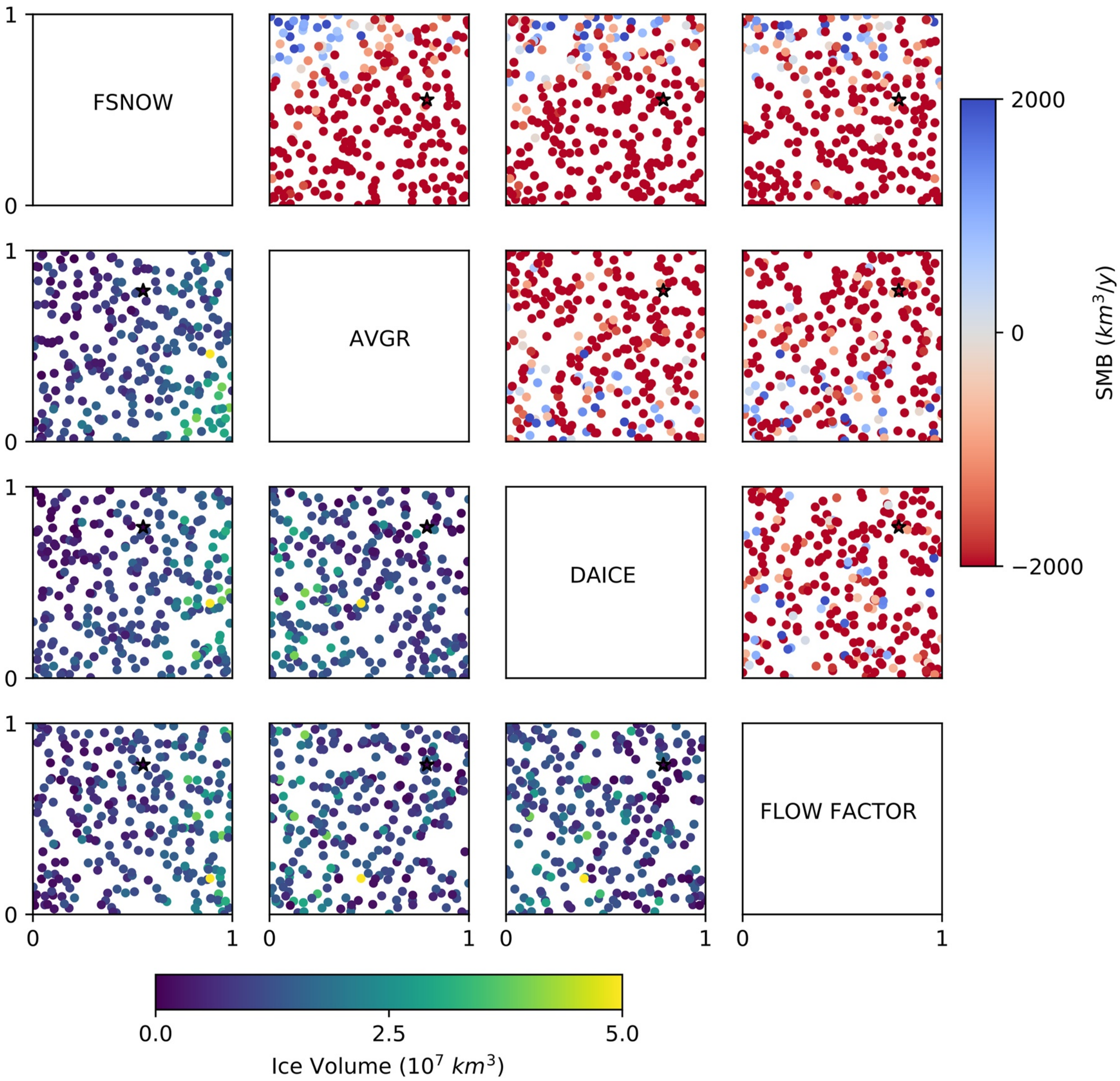

Figure 4. The influence of four parameters on the resulting initial 20-year Surface Mass Balance (top, red-blue color map) and equilibrium Ice Volume (bottom, purple-yellow color map). Parameter pairs can be found from the axis labels in the blank plots. All values are normalized. For reference, the control simulation is marked with a star.

model. Alaskan ice extent was limited by the wider ice sheet disrupting atmospheric circulations (Löfverström & Liakka, 2016; Tulenko et al., 2020). The likelihood of matching observations is not helped by the climate coupling; a low resolution of FAMOUS struggles to simulate the temperature and precipitation gradients caused by steep topography such as the Aleutian Range (Abe-Ouchi et al., 2007). Prescribing the SMB forcing in an uncoupled model nudges the simulated ice sheets toward the ice sheet prescribed (usually from reconstructions) in the climate-only model, encouraging a better match between modeled and reconstructed ice geometry, but not necessarily for predictive reasons. Instead, in this case, a coupled climate-ice sheet model essentially introduces additional freedom into the simulations to produce a greater variety of ice sheets.

5. Importance of Albedo Values

While 13 parameters are varied in the ensemble, only three to four of these parameters explain the majority of the variation in the model outputs. This is clearly demonstrated in Figure 4, where the parameter values for Ruled Out and NRO simulations are compared. NRO simulations are clustered in the FSNOW, DAICE, and AVGR parameter space, but this is not apparent for the other 10 parameters.

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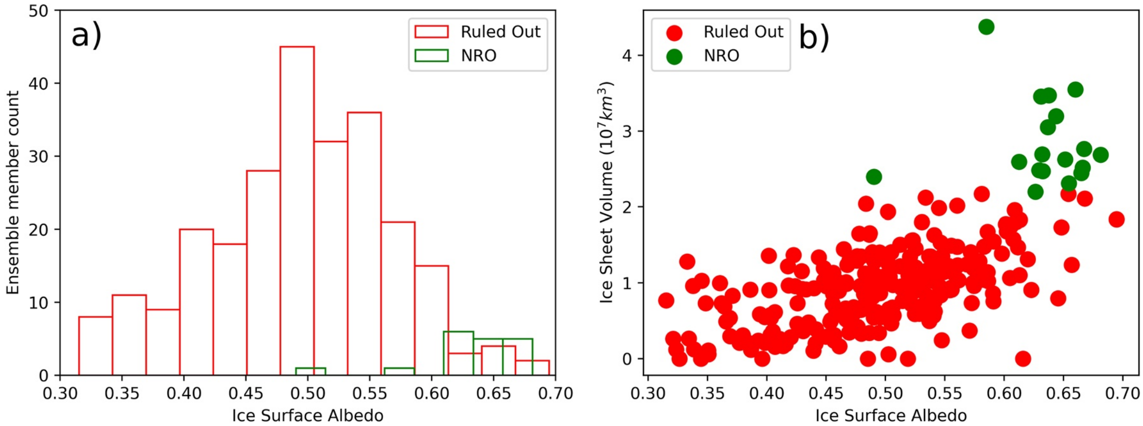

Figure 5. Average Ice Sheet surface albedos for Not Ruled Out (NRO) and Ruled Out (RO) ensemble members. (a) The distribution of ice surface albedos for NRO and RO ensemble members. (b) The relationship between the final ice sheet volume and the ice sheet surface albedo. All ice sheet surface albedos are taken as a 30-year average of the July albedo at the start of a simulations, so that ice sheet extents are comparable between ensemble members.

The parameters that are most influential on the simulated ice sheet volume all control the ice sheet surface albedo, these are FSNOW, DAICE, AVGR (Table 1, see R. S. Smith et al. (2020) for full details of how these terms are used in their respective parameterizations). We stress again that the simulations do not include a dynamic vegetation model and that including one or prescribing a different fixed vegetation distribution could affect the simulated ice sheet extent (Sommers et al., 2021). The type of land cover and vegetation surrounding the ice sheet could impact the surface energy balance and induce further feedbacks as the ice sheet and climate co-evolve. However, simulating the evolution of vegetation through ice advance or retreat with a dynamic vegetation model introduces more complexity and requires careful setup and analysis. This could be an interesting opportunity for future work.

Figure 5 shows that, as expected, the largest ice volumes result from having more reflective snow and bare ice (from AVGR to DAICE, respectively), and a high density threshold (FSNOW) to start considering snow to be ice (keeping a surface classed as more reflective snow, rather than ice, for longer).

5.1. Prior Overtuning

We have shown that albedo parameters that have been manually tuned to produce a reasonable contemporary Greenland Ice Sheet (Gregory et al., 2020; R. S. Smith et al., 2020) produce a collapsed North American Ice Sheet at the LGM. Here, we explore the mechanism causing the parameter discrepancy between the ice sheet configurations, and why the present day Greenland Ice Sheet may not be a sufficient target for calibrating a coupled climate-ice sheet model. The contrasting behavior between present day Greenland and North American LGM Ice Sheets is the FAMOUS-ice simulations associated with differences in the magnitude of downward shortwave radiation at surface and cloud cover between the two ice sheets. In the Pre-Industrial simulation, both the margins of the Greenland ice sheet and the northern North America are often covered by thick clouds, reducing downward shortwave radiation (Figure 6). In response, the ice sheet surface albedo was tuned lower than observations when initially configuring the model for the PI Greenland Ice Sheet (R. S. Smith et al., 2020). In contrast, in the LGM simulations, there is slightly reduced cloud cover over the Greenland ice sheet, and significantly reduced cover over the North American Ice Sheets. This means that the downward shortwave radiation is disproportionately increased over North America compared to Greenland.

Simply, the relationship between cloud and ice sheet surface albedo can be split into three regimes: "sunny cold," "cloudy warm," and "sunny warm" (Figure 7).

- *Sunny cold*. At high surface elevations, there are lower temperatures and cloud cover. This regime can sustain an ice sheet.
- *Cloudy warm*. At lower elevations, where temperatures are warming, there can be sufficient cloud cover limiting ice sheet melt by reducing downward shortwave radiation. In this regime, the ice sheet is insensitive to its surface albedo. This regime can sustain an ice sheet.
- *Sunny warm*. At lower elevations with limited cloud cover downward shortwave radiation causes surface melting. In this regime, the ice sheet is highly sensitive to surface albedo parameters.

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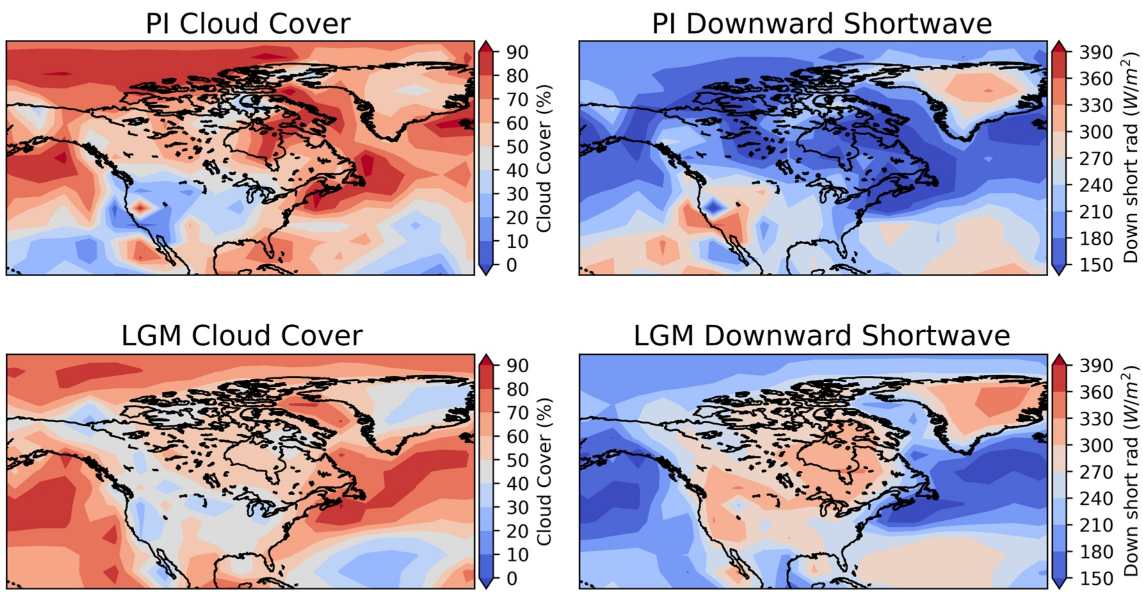

Figure 6. July cloud cover and downward shortwave radiation for the PI simulation, and the LGM ensemble.

Figure 7 shows downward shortwave radiation and ice surface albedo values for each ice sheet surface grid cell in the control runs for PI Greenland and LGM North America. While both runs occupy the "sunny cold" and "cloudy warm" regimes, the "sunny warm" regime is dominated by the LGM run. In this manner, the North American Ice Sheet becomes very sensitive to low ice albedo parameters.

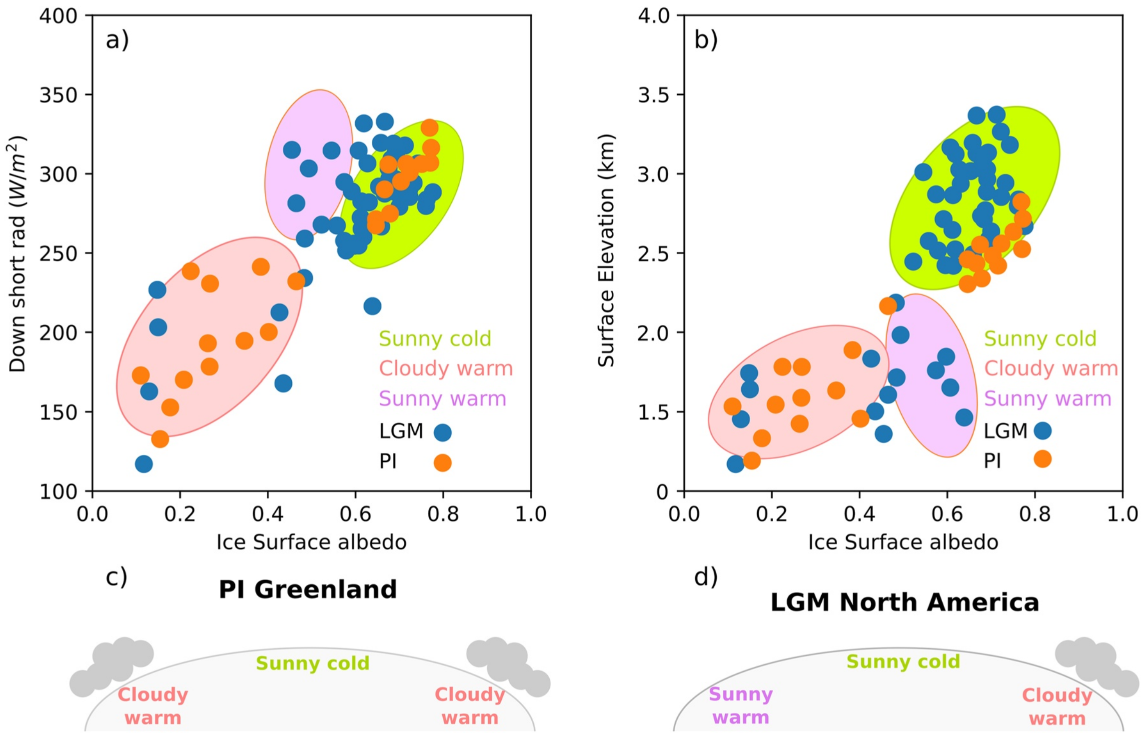

Figure 7. Comparison of energy balance between LGM North American and modern Greenland Ice Sheets: (a) Downward Shortwave radiation versus ice surface albedo in each ice sheet gridcell for Greenland in the "standard" pre-industrial simulation (orange) and for the North American Ice Sheet in the "standard" LGM simulation (blue). Shaded ovals show the proposed surface albedo and radiation behavior groups for the Greenland and North American Ice Sheets. (b) Panel (a) repeated for ice surface albedo and downward shortwave radiation (c, d) A schematic of the different radiative effects of clouds on the PI Greenland and LGM North American Ice Sheets.

Table 2
Albedo and Radiative Characteristics of PMIP3 Models and FAMOUS-Ice

PMIP3 model	Surface albedo	Sfc down s/w radiation	Absorbed s/w at sfc
FAMOUS-ice	0.68 (0.56–0.75)	296 (265–324)	96 (71–126)
CCSM4	0.70	295.18	87.40
CNRM	0.34	307.67	204.47
GISS	0.66	341.52	116.34
IPSL	0.75	343.80	86.58
MIROC	0.75	351.76	86.51
MPI	0.83	348.92	58.07
MRI	0.57	318.19	138.25

Note. FAMOUS-ice values are taken from the LGM ensemble, showing the mean value and ensemble range (in brackets).

5.2. Comparison to PMIP3 Models

We further examine outputs from PMIP3 models (Adloff et al., 2018; Brady et al., 2013; Sueyoshi et al., 2013; Ullman et al., 2014; Voldoire et al., 2013) to verify whether the very strong downward shortwave radiation over the North American Ice Sheet observed in FAMOUS-ice is a common feature among other models. The area-averaged values over the North American Ice Sheet are summarized in Table 2. A visual comparison of the simulated LGM and PI downward shortwave radiation in the FAMOUS-ice and PMIP3 models is included in Section S6 in Supporting Information S1. The results show that other PMIP models simulate stronger downward shortwave radiation at the southern margin of the ice sheet, and that FAMOUS-ice shows the smallest value for this together with CCSM4 (Table 2). In other words, other models may impose an even stronger melt on the ice sheet than in our simulations. This is associated with the strongest cloud radiative effect in shortwave radiation in FAMOUS-ice and CCSM4. These results show that the North American Ice Sheet would also be sensitive to low albedo values in other PMIP models; if they allow a very small minimum bare ice albedo, a very large amount of solar energy would be absorbed. Therefore, the larger sensitivity of the LGM North American Ice Sheet to low albedo parameters is likely not a unique feature of FAMOUS-ice but seems to be a common feature among other climate models.

There is a large uncertainty and variety in minimum ice sheet surface albedo among PMIP models ranging from 0.2 in MRI and FAMOUS-ice to 0.7 in MPI (Alder & Hostetler, 2019). These differences in albedo values are induced by the combined effects of discrepancies in the physics of the albedo scheme and biases in AGCMs. For the latter, biases in cloud radiative effects (R. S. Smith et al., 2020) and horizontal resolution (Kapsch et al., 2021) can affect the choice of albedo values. Importantly, the albedo values selected are often strongly tuned to reproduce the modern SMB. This is sensible when the focus is on future change of Greenland Ice Sheet in the next few decades, since changes in SMB are the dominant driver on this time scale. However, ice sheets evolve (e.g., in response to climate) over much longer timeframes, and on a longer time-scale, when the ice sheet is subject to larger instabilities and more pronounced climate interactions, tuning the albedo parameters may cause an unrealistic relationship between surface albedo and cloud cover. Moreover, changes in the clouds over the next century could have pronounced effects on SMB with unrealistic surface albedo. Thus, we suggest that it is important to tune climate-ice sheet models less tightly on a wider variety of climatic conditions. The Last Glacial Maximum, with its good observational constraints on climate and ice extent, offers an ideal target for testing and tuning coupled climate-ice sheet models.

In the case of FAMOUS-ice, the surface albedo parameters used for the contemporary Greenland Ice Sheet were originally tuned to a low value to compensate for an excessive reflection of shortwave radiation by clouds (R. S. Smith et al., 2020). This was performed to better simulate a stable and realistic Greenland Ice Sheet geometry under modern day climate. However, the resulting ice albedo parameter sets are too low to produce a realistic North American Ice Sheet at the LGM due to the different cloud cover and downward shortwave radiation over the ablation zone. Therefore, we are in an undesirable situation where a model with good SMB in modern climate is unable to simulate the LGM North American Ice Sheets and could also be unable to simulate other ice sheets and time periods. This result suggests that overtuning the albedo to compensate for biases in other components under modern climate may cause degraded simulation results under different climate states, when the cloud properties and downward shortwave radiation over ablation zones are different from modern.

6. Conclusions

We have applied a new coupled Climate-Ice Sheet model (FAMOUS-ice) to simulate the maximum extent of the last North American Ice Sheets. The standard model setup manually tuned for modern day Greenland resulted in a collapsed ice sheet at the LGM. We underwent a process of "detuning" the model, running hundreds of simulations to produce a range of reasonable equilibrium ice sheets, enabling us to explore the influence and importance of key uncertain parameters. Large parts of the parameter space produced collapsed ice sheets at the

LGM, but a selection of simulated larger ice sheets show that FAMOUS-ice is able to simulate the maximum extent of the LGM North American Ice Sheet. These simulations show a particularly good match to the southern Laurentide limits, but some systematic ice overgrowth remains in Alaska.

From our results, we are able to identify that the parameters controlling ice sheet surface albedo dominate the simulated variability in ice sheet geometry. The results demonstrate the potential importance of uncertain ice sheet albedo parmeters. Importantly, combinations of albedo parameter values that produced a reasonable contemporary Greenland Ice Sheet do not necessarily produce a reasonable LGM North American Ice Sheet. This is because albedo parameter can be overtuned to compensate for biases in modern clouds over Greenland. The different cloud distributions over the Southern Laurentide Ice Sheet at the Last Glacial Maximum provide a useful "stress test" for coupled climate-ice sheet models. This highlights the potential problems of relying solely on contemporary observations for model tuning. Efforts to find a region of the parameter space that produces reasonable simulations of contemporary and glacial ice sheets are important and could lead to improved confidence in future ice sheet projections.

Data Availability Statement

Output from an ensemble simulation is archived online (Gandy et al., 2023) (https://doi.org/10.17632/8kswwpnjyz.1).

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