

Pre-LGM Northern Hemisphere paleo-ice sheet topography

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Pre-LGM Northern Hemisphere paleo-ice sheet topography

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Abstract

We here reconstruct the paleotopography of Northern Hemisphere ice sheets during the glacial maxima of marine isotope stages (MIS) 5b and 4. We employ two approaches, geologically based reconstruction and numerical modeling, in mutually supportive roles to arrive at probable ice sheet extents and topographies for each of these two time slices. For a physically based 3-D calculation based on geologically derived 2-D constraints, we use the University of Maine Ice Sheet Model (UMISM) to calculate ice-sheet thickness and topography. The approach and ice-sheet modeling strategy is designed to provide robust data sets of sufficient resolution for atmospheric circulation experiments for these previously elusive time periods. Two tunable parameters, a temperature scaling function applied to a spliced Vostok-GRIP record, and spatial adjustment of climatic pole position, were employed iteratively to achieve a good fit to geological constraints where such were available. The model credibly reproduces the first-order pattern of size and location of geologically indicated ice sheets during marine isotope stages (MIS) 5b (86.2 kyr model age) and 4 (64 kyr model age). From the interglacial state of two north-south obstacles to atmospheric circulation (Rocky Mountains and Greenland), by MIS 5b combined Quebec-Central Arctic and Scandinavian-Barents/Kara ice sheets had effectively increased the number of such highland obstacles to four. This number remained constant through MIS 4, but at the last glacial maximum (LGM) dropped to three, through the merging of the Cordilleran and the proto-Laurentide Ice Sheet to a single continent-wide North American ice sheet.

1 Introduction

We here address a fundamental information gap in climate science, Northern Hemisphere paleotopography during the last interglacial to glacial transition. At present, a good understanding of Northern Hemisphere paleotopography exists for the last glacial maximum (LGM) and the deglacial phase of the last glacial cycle (Peltier, 2004).

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However, there is a lack of geologically constrained data sets for defined time intervals during the 95 kyr build-up phase towards the last glacial maximum. Considerable progress is made in understanding ice sheet dynamics in this elusive time interval (Stokes et al., 2012), but the scarcity of geological and geomorphological data (Kleman et al., 2010) that can constrain numerical models is still an impediment to our understanding of the time periods of most rapid ice sheet build-up. This situation hampers research regarding ice-sheet topographical feedbacks on atmospheric circulation (Cook and Held, 1998; Roe and Lindzén, 2001; Langen and Vinther, 2009; Liakka and Nilsson, 2010), and is an obstacle to our understanding of Northern Hemisphere atmospheric circulation during periods which in terms of number, location and size of ice sheets were different from both interglacial and full-glacial conditions. Indeed, idealized coupled atmosphere–ice-sheet models have demonstrated that the interaction between ice-sheet topography and atmospheric circulation can strongly influence the spatial distribution of continental-scale ice sheets (Roe and Lindzen, 2001; Liakka et al., 2011). Thus, a number of important climate questions can only be addressed from a foundation of improved paleotopographical reconstructions; can atmospheric circulation changes have caused mass-balance patterns that left Alaska largely ice-free throughout the last glacial cycle, the high-elevation Cordilleran region only ephemerally glaciated (Clague, 1989), but the almost entirely lowland Laurentide area (Fig. 1) persistently glaciated? Did downstream influence of the North American ice-sheet topography on the atmospheric circulation contribute to southwestward migration of the mass center of the Eurasian ice sheet? Amplification of the effects of modest insolation changes appears necessary to explain ice sheet build-up in the Laurentide area (Otieno and Bromwich, 2009). In addition, the topographical and albedo changes caused by ice sheets are likely part of the non-linear response of the climate system to the insolation changes that pace glacial cycles. Considerable progress is being made on establishing the general response characteristics of the climate system by statistical analysis of the global ice volume response to insolation forcing (Lourens et al., 2010; Imbrie et al., 2011). However, for understanding the cause of non-linear responses related to the

growth of ice sheets, also the location, extent and height of individual ice sheets must be understood.

Improved understanding of Northern Hemisphere paleotopography of intermediate sized ice sheets can, if combined with the relevant array of paleoclimatic proxy data, yield a valuable calibration data set for tuning and comparison of climate models at global ice volumes smaller than the LGM-case normally used (Kageyama et al., 1999; Otto-Bliesner et al., 2006). The present situation, in which research focus has been on two climatic end members, largely precludes research on non-linear responses to ice build-up. Over the last decades there has accumulated evidence that ice sheets extent and volume varied more drastically over shorter time-scales than previously thought, and a new paradigm of highly dynamic ice sheets has emerged. Of particular importance is the evidence for ice-free core areas in Fennoscandia (Wohlfarth, 2010; Helmens and Engels, 2010), the consequence of which is that the Fennoscandian ice sheet had to be fully rebuilt from zero ice-volume during each of the cold phases (MIS 5d, 4 and 2). From this new paradigm, *available time* and *precipitation* emerge as important limiting factors for the glacial development, and there is a need to explore the atmospheric circulation and conditions that allowed ice sheets to be fully rebuilt during the relatively short intervals of declining Northern Hemisphere high-latitude summer insolation.

Reconstructions of the LIS at the LGM based on numerical ice sheet modelling have ranged from a thick mono-dome ice sheet (Denton and Hughes, 1981) to a relatively thin multidomed ice sheet (Clark et al., 1996). More recent ice sheet models (Marshall et al., 2000; Tarasov and Peltier, 2004) converge on the LIS at the LGM exhibiting (i) a multidomed shape with a complex configuration of ice streams and inter-stream ridges in peripheral areas, (ii) frozen-bed core areas under the main domes, in line with was inferred from geomorphological evidence by Kleman and Hättestrand (1999), and (iii) ice thicknesses substantially smaller than the early monodome reconstructions. Few numerical modelling exercises focus on, or include, the growth phase of the ice sheet (Huybrechts and T'siobbel, 1995; Charbit et al., 2007; Bintanja et al., 2008;

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Bonelli et al., 2009). Many of those who do, commonly conflict with the geological evidence for ice sheet extent at the LGM, which is currently the prime available “target” period for assessing the validity and credibility of models, by showing too much ice cover in Alaska and the Mackenzie mountains (NW sector), areas that were largely ice-free at the LGM, and also showing continuous and massive glaciation of British Columbia instead of the geologically indicated ephemeral ice cover (Clague, 1989). These common problems are discussed at length in Bonelli et al. (2009). Most models are also unable to during ice sheet build-up generate a southward extent of the Quebec dome sufficiently large to match the geological evidence (Stea, 2004; Kleman et al., 2010). The recent ensemble-based reconstruction by Stokes et al. (2012) is focused on the build-up phase and represents a major step forward in understanding the dynamics of proto-Laurentide ice sheets. It was not constrained by geological data, but the output is thoroughly discussed in relation to geological data. One of the currently leading ice-sheet models (ICE-5G, Peltier, 2004) is based on isostatic recovery rather than mass-balance driven ice physics, and is unable to reconstruct pre-LGM conditions because of the time constants involved in isostatic recovery. An early attempt at a geologically and geomorphologically constrained pre-LGM LIS model was presented by us some years ago (Kleman et al., 2002), and we here build on experience gained during that work.

Ice sheet reconstructions for the 115–21 kyr build-up phase involve additional difficulties compared to reconstructions for the LGM and deglacial time, because the amount of stratigraphic control for this time interval is much smaller than for the deglaciation period 21–8 kyr. The time interval is much longer, and can rarely be addressed with radiocarbon dates. In addition, almost all landforms and deposits created in this time interval have been subjected to overriding ice during the LGM and later stages, which in most cases has led to erosion or burial and therefore full or partial destruction and reshaping, and to severe spatial fragmentation of the record compared to the records from the LGM and deglacial periods (Kleman et al., 2010). These difficulties add up to a major methodological challenge, and require a focus on only the *first-order* patterns

in the evolution of ice sheets in this time interval. For the pre-LGM period it is not possible to achieve the spatial and temporal precision of post-LGM reconstructions (Dyke et al., 2002; Ehlers and Gibbard, 2003).

For the ice volume maxima of marine isotope stages 5b and 4, the two youngest build-up phases preceding the LGM, we here present first-order paleotopographical data sets tailored for use as the topographic input for atmospheric circulation experiments. We have used the best available geological constraints and through numerical modeling strived for first-order reconstructions that represent our current best understanding of location, extent and thickness in this time interval. We have not tried to capture details in ice sheet outline or internal dynamics because the intended use, atmospheric circulation modeling, is not critically dependent on such information.

Our approach is not puristic. We do not regard geological data as an indisputable “truth” against which model output is evaluated. Instead, in line with Kleman et al. (2010), we argue that discrepancies between “control” data and numerical modeling output can point to flaws in interpretation of data as well as deficiencies in the modelling step. We here employ both approaches, geologically based reconstruction and numerical modeling, in mutually supportive roles to arrive at the most probable ice sheet extent and topography for each time slice.

2 Spatial extent of ice sheets

Our strategy is governed by the intended end use, paleo-atmospheric circulation modeling. For that purpose a high precision ice sheet outline is not crucial, but the existence or non-existence of an ice sheet in a certain location is. It can be expected that small and thin ice sheets will not significantly affect the large-scale atmospheric circulation. Hence we are primarily concerned with capturing the main ice sheet features and ice sheets with a length of at least 1000 km.

The situation regarding data that spatially constrain pre-LGM ice sheets is very uneven for the respective glaciation centers, with a relatively good understanding in

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western Eurasia, some effective constraints, albeit with poor dating control, for the Laurentide Ice Sheet, and very little solid knowledge for the Cordilleran and Inuitian Ice sheets in these time intervals.

We have here chosen to base our synthesis primarily on previous spatial reconstructions (Dredge and Thorleifson, 1987; Clague, 1989; Clark et al., 1993; Kleman et al., 1997, 2010; Lundqvist, 2004; Svendsen et al., 2004; Mangerud, 2004; Lambeck et al., 2006) which cover this time interval. A full consideration of primary morphological and stratigraphic data is offered in several of the key source publications. In compiling the set of 2-D modeling targets (Fig. 2), each corresponding to a certain time interval, we have used the following approach: (i) where approximate consensus on ice sheet outline exists, we have drawn an averaging and somewhat simplified outline, (ii) where clearly dissenting views exist, we have weighed the evidence and decided on what we consider the most likely alternative, (iii) for an outline that is considered spatially reliable, but for which alternative age assignments exist, we have weighed the age options, and chosen that which we consider to be the most probable one.

The generally poor understanding of pre-LGM extent of particularly the Cordilleran and Inuitian ice sheets constitutes a “missing ice sheets” problem. There is not sufficient, nor reliable enough, data to allow us to show ice sheets outlines in these areas on any of the panels in Fig. 2, yet embryonic or smaller ice sheets are very likely to have existed in these areas during the time intervals in question. To use the outlines shown in Fig. 2 as “hard” modeling targets (in which ice in areas marked as ice-free in Fig. 2 is regarded as indicating a poor model fit) would clearly be erroneous, and would have caused us to use inappropriate forcing of the model to keep ice out of these areas.

For both Eurasia and North America there are indications that the ice sheet extent was approximately similar during stages 5d and 5b. In view of the scarce data and dating difficulties in this time interval it is futile to separate spatial evidence from the two with any degree of certainty. Instead we have chosen to focus on stage 5b, which is closest in time to the two other major stadials MIS 4 and the MIS 2 (LGM). Given the similarity in global ice volume (Lisiecki and Raymo, 2005) and available sea-level data

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(Lambeck et al., 2002) between the two stages we consider it reasonable to regard this pattern as archetypical also for stage 5d.

2.1 North America

For the extent of North American ice sheets before the LGM, we have based our 2-D outlines primarily on the recent review by Kleman et al. (2010), and reference contained therein. For the Keewatin sector of the Laurentide ice sheet there is good evidence for one or two, probably closely related, stages of a pre-LGM ice sheet with a dome center in the northern Keewatin or south central Arctic region (Kleman et al., 2002, 2010; McMartin and Henderson, 2004). However, the east-west extents, maximum southerly extents, and ages are poorly constrained. Inferences about their pre-LGM age can be made on the basis of flow pattern and relation to other flow systems, but we know of no firm dating constraints. The flow patterns do indicate that both were dynamically independent of the Quebec Dome. The lineation evidence for this ancestral Northern Keewatin dome, highlights the “missing ice sheet problem”. The lineation evidence for flow out of an ancestral Northern Keewatin/Central Arctic Dome is solid but covers only a restricted sector or corridor which lacks any topographic funneling. Glaciological considerations therefore require this lineation system to be only a fragment of a much wider flow pattern emanating from northern Keewatin or the south-central Arctic islands. Despite its limited area it thus indicates the existence of a full ice sheet of which, unfortunately not enough is known about its east-west extent to warrant its full inclusion in Fig. 2. However, its minimum southward extent into the area of preserved traces is a viable modeling target.

In the Quebec sector, Kleman et al. assigned the “Atlantic” flow system, which indicates an ice divide closer to the Atlantic coast than any subsequent stage, to stage 5b or 5d. We consider it likely that this pattern characterized both stages and have therefore included it as a constraint for stage 5b. Regarding the Cordilleran Ice Sheet, Kleman et al. (2010) concluded that there was flow-trace evidence for “old” configurations with more easterly dispersal centers (Stumpf et al., 2000) than the ones that

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prevailed in near-LGM time, but constraints on age and outline are lacking. Hence, the Cordilleran Ice Sheet is one element of the “missing ice sheet” problem, and it appears very likely that this region was at least partly glaciated during the stadials 5d, 5b, and 4, but there is simply not enough geological and geomorphological data to suggest an outline. Similarly, it appears likely that a proto-Inuitian Ice Sheet existed on the Queen Elisabeth Island during these stadials, but again there is not enough data to warrant the suggestion of a specific outline.

2.2 Eurasia

For Northern Eurasia we mainly follow the spatial reconstructions of Svendsen et al. (2004) and Lambeck (2006). For Fennoscandia, the main sources we have used are Mangerud (2004), Lundqvist (2004), Kleman et al. (1997) and Lambeck et al. (2006). For stage 5b there is agreement that the Bothnian Bay and Gulf were glaciated, but not the Baltic. This indicates a width of the Fennoscandian Ice Sheet of less than half its LGM width. Southern Sweden was not glaciated at this time, nor were the British Isles. According to Svendsen et al. (2004), a unified ice sheet existed from southern Norway all the way to the Putorana Plateau. Lambeck et al. (2006) instead indicate a narrow gap between the Scandinavian and Barents-Kara components. We here follow the Svendsen et al. (2004) reconstruction, but regard this suggested separation as a minor issue since the overall outline of the two reconstructions are quite similar. During stage 4, there was apparently an overall shift of mass towards Fennoscandia, and a reduction of ice mass in the Kara sector. The Putorana Plateau was unglaciated, or alternatively, had an independent ice cap. The Fennoscandian Ice Sheet filled the entire Baltic depression and touched upon mainland Europe. The eastern margin of the Scandinavian Ice Sheet at this stage was approximately halfway between its stage 5b and LGM positions. The British Isles were largely glaciated during the LGM, but we acknowledge great uncertainty regarding the ice extent and possible confluence with Scandinavian ice during stage 4. Although volumetrically small, the

southwestward lengthening of the Eurasian ice sheet that a British Ice Sheet may have provided was potentially important for atmospheric circulation changes.

3 Model and model setup

We employ the University of Maine Ice Sheet Model (UMISM), which consists of a time-dependent finite-element solution of the coupled mass, momentum, and energy conservation equations (Fastook and Chapman, 1989; Fastook, 1993) with a broad range of applications (Holmlund and Fastook, 1995; Johnson and Fastook, 2002; Kleman et al., 2002). The primary input to the model is present bedrock topography, the surface mean annual temperature, the geothermal heat flux, and the net mass balance, all defined as functions of position. The solution consists of ice thicknesses, surface elevations, column-integrated ice velocities, the temperature field within the ice sheet, the amount and distribution of water at the bed resulting from basal melting, and the amount of bed depression resulting from the ice load.

The primary solution is of the mass conservation equation. The required column-integrated ice velocities at each point in the map-plane are obtained by numerically integrating the momentum equation through the depth of the ice. Stress is a function of depth and related to strain rates through the flow law of ice. The temperature, on which the flow law ice hardness depends, is obtained from a 1-D finite-element solution of the energy conservation equation. Internal heat generation produced by shear with depth and sliding at the bed is also included. A flow enhancement factor allows “tuning” of the model to field data. Areas of basal sliding can be specified if known, or internally determined by the model for regions where the bed reaches the pressure melting point and lubricating water is available. A 2-D solution for conservation of water at the bed allows for movement of the basal water down the hydrostatic pressure gradient (Johnson and Fastook, 2002).

With surface temperature and mass balance obtained from a simple climatological parameterization (Fastook and Prentice, 1994), the model can be run with no more

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input than the specified bedrock topography, in what can be called a “free-running mode”. However, it can also be tightly constrained by climatological, geological and glaciological data when such data are available. Ice margin positions can be specified and areas of basal sliding can be derived from the distribution of basal temperatures or they can be specified by the presence of erosional features on exposed landscapes.

We used a relatively coarse grid spacing of 95 km, suitable for the broad picture we are attempting to produce for use in atmospheric circulation modeling. A regular rectangular grid with 10 000 (100 × 100) nodal points spanning the Northern Hemisphere was used. We divided this into a Western Hemisphere grid, and an Eastern Hemisphere grid so that climatic adjustment could be applied separately to the two hemispheres to obtain the best fit. The time step required at this resolution was 10 yr.

Ice hardness is calculated from modeled internal temperatures and is based on ice hardness parameterizations from Paterson (1994). Sliding at the bed follows Weertman (1964, 1969, 1982) with a lubricating factor coupled to the presence of water at the bed generated by the internal temperature calculation. While this is a shallow-ice approximation model where only the basal shear stress is considered, we do include a longitudinal thinning strain rate at the grounding line that models ice drawn into a buttressing ice shelf that is not explicitly modeled. A parameter we call the “Weertman” ranges from 0 (full buttressing, no thinning into the ice shelf) to 1 (no buttressing, thinning at a rate commensurate with an unconfined floating ice slab (Weertman, 1957).

As forcing function, we used a spliced GRIP (Johnsen et al., 1997) and Vostok (Petit et al., 2001) record (shown in Fig. 3a as a red line labeled “Core”) as our global temperature proxy. The latter part of the GRIP record is reliably dated but reliability deteriorates towards the last interglacial. This record also contains two extreme but shortlived cooling events at 70–75 kyr which are suspected to not reflect regional climate in North America and Eurasia, and which in previous modeling experiments created brief and unrealistically large ice sheet area increases (Kleman et al., 2002). We therefore employed a spliced Vostok-GRIP record to avoid these problems in the GRIP data. The use of a global temperature such as our spliced GRIP-Vostok record is common in

reconstruction of ice sheet glaciation cycles, and indeed provides an important timing and amplitude to whatever climate parameterization is used. Since we are fitting our modeled ice sheets to geologically-constrained areal footprints, we regard it as justified to adjust this global temperature proxy with a temporally variable and regional scaling factor, shown in Fig. 4a, in order to recreate ice sheet sizes compatible with the constraints. Deviations from the core proxy do not exceed 1 °C for Eurasia and 1.5 °C for North America. For both North America and Eurasia it was necessary to increase amplitudes by a factor of 1.2 (a decrease in temperature by approximately 1 °C) in the early part of the glacial cycle in order to match the target ice sheet areal footprints. The scaling amplitudes were reduced for both ice sheets during and after stage 4, in order not to overshoot documented stage 2 ice margins (1.4 °C warmer briefly for North America and 0.25 °C warmer for Eurasia almost to the LGM).

The model uses a radially symmetrical mass balance function, with mass balance being a function of elevation and the latitude relative to the climatic “pole” (Fastook and Prentice, 1992). In the mass balance scheme we have employed, the ablation is based on the positive degree day method (Braithwait and Olsen, 1990) and the accumulation is given as a function of the saturation specific humidity and the local ice-sheet slope. In the model, the temperature field has a seasonal cycle and a simple spatial structure: it increases linearly with distance from the climatic pole and decreases linearly with height, i.e. a constant lapse rate is assumed. The two key control parameters of the mass balance scheme are the position of the climatic pole and a constant background temperature that sets general climatic state.

Moving the climatic pole to East Central Greenland is instrumental in achieving a good fit to the glacial constraints for both the Eurasian and the North American ice sheets. Figure 4b and c shows the changing position of this pole throughout the modeled time interval. Green (Eurasia) and red (North America) lines show the adjustments in the pole position that was necessary to obtain the fits. The Eurasian pole starts in the north and moves to the south and west, whereas the North American pole begins further south and west and stays stationary except for a slight shift south and east

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calving parameter. The only hard ice-cutoff function applied was the -200 m bathymetric contour. Several hundred iterations were run, before selecting the ones reported here.

4 Results

4.1 Fit to the constraints

Figure 5 shows ice thickness in relation to the constraints, and Fig. 6 shows ice elevations. The main deviations from a “perfect” fit to the geological constraints for stage 5b are the following. There is too little ice in Newfoundland, making the Quebec Dome shorter in the NW–SE direction than the constraints indicate, and too much ice on the western flank of the Quebec Dome. The overall size of the Quebec Dome is in line with the constraints. The model slightly undershoots the minimum size for the proto-Keewatin Dome. In Eurasia, there is too little ice in southern Norway, and the Scandinavian Ice Sheet reaches a somewhat larger easterly extent than indicated by the constraints. The Barents-Kara Dome is well centered within the constraints but slightly too small, particularly in the sector of the Putorana Plateau. In line with the situation in North America, over- and undershoots of the ice margin largely cancel out.

For stage 4, the fit is in general very good. The only significant deviations from the constraints are the somewhat too small ice extent in the Canadian Maritimes, a slightly too small extent of the proto-Keewatin Dome, and some excess ice in the easternmost part of the Cordilleran Ice Sheet. The model generates a plausible Cordilleran Ice with restricted extent and an easterly located ice divide.

4.2 Paleotopographic evolution

In the paleotopographic plots (Fig. 6) we see that the interglacial topography is characterized by two major highlands of a size large enough to significantly affect atmospheric circulation; Rocky Mountains – Coast Range and Greenland. In contrast to

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the situation in North America and Eurasia, glaciation south of 60° N is impossible in Greenland. During stage 5b (Fig. 6a), emerging ice sheets constitute four additional highlands large enough to potentially influence atmospheric circulation. These are arranged in pairs in close proximity to each other, the Barents-Kara Ice Sheet and the Scandinavian Ice Sheet in Eurasia, and the Quebec and Central Arctic Domes in North America. These pairs constitute two highlands, 3500 and 3000 km long, respectively, each with only a minor gap in the central parts.

In contrast to the interglacial situation with two significant and separated N–S oriented highlands located across the westerlies, by stage 5b the number has thus increased to four. In North America, a 1500–2500 km separation exists between the Rocky Mountains and the two ice sheets on the easternmost part of the continent.

By Stage 4 (Fig. 6b), the number and location of ice sheets is still the same as during stage 5b, but the Quebec Dome has grown radically, constituting the largest topographic feature in the Northern Hemisphere with the exception of the Tibetan Plateau. The North American and Eurasian ice sheets are by this stage both over 3500 km long, each with a high saddle connecting two domes. The separation between the North American Ice Sheet and the Rocky Mountains has shrunk to few hundred kilometers in the north, and approximately 1500 km further south.

The time separation between stage 4 and the LGM is large, more than 40 kyr, and there is conflicting evidence (Ukkonen, 2003; Helmens and Engels, 2010; Wohlfarth, 2010) regarding the ice extent in both Eurasia and North America during stage 3. Our model indicates a significant shrinkage of ice in Eurasia, but a slow growth in North America. In stage 2 (Fig. 8c) the total ice sheet size and elevation in Eurasia is approximately similar to stage 4, but its footprint has moved 700 km to the southwest. The North American evolution leading into stage 2 is far more dramatic. Here, the two eastern and northern domes have merged with the Cordilleran Ice Sheet, to form a unified Laurentide and Cordilleran ice sheet reaching from coast to coast. There is no longer a midcontinent gap, and consequently the *number* of obstacles across the westerlies is down to three, although one of them has a very large east–west extent.

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4.3 Evolution of mass

A surprising observation regarding the evolution of mass for Eurasian and North American ice sheets is the contrast between the behaviors of the two ice sheets. The ice volume curve (Fig. 3b) shows ice volume variation until approximately 70 kyr as being in concert on the two continents, and being of comparable magnitude. However, during stage 4 and stage 2 the volume buildup is far more dramatic in North America than in Eurasia. During stage 5b ice volume in North America is larger than in Europe by a factor of 1.3. During stage 4 this number increases to 2 and by stage 2, North America has four times the Eurasian amount of ice.

5 Discussion

5.1 Forcing of the model

There are two main ways to force a stand-alone ice-sheet model, either mass-balance parametrizations based on first-order physics, or by interpolating measured or modeled climate parameter fields representing end-member climates, typically the LGM and the present. The strong zonal misfits (too much ice in the Alaska–E. Siberia sector, and too little ice in Quebec, southern Scandinavia and the British Isles) that commonly affect Northern Hemisphere modeling experiments are an inherent effect if simple north pole-centered mass balance-schemes are employed even when present-day zonal climate anomalies are built in (Abe-Ouchi et al., 2007). Still, in order to tune the model to the available constraints, we have chosen the first approach because it can in a simple way be spatially tuned to account for first-order spatial misfits by moving a climatic pole. No similar straightforward spatial-tuning possibility exist for fixed climate fields, and it is unclear what realism as representing real climates they retain if spatially manipulated.

When comparing with the recent ensemble modeling study by Stokes et al. (2012) we notice that they agree on a first-order pattern of ice build-up on the northeastern

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continental rim, but that the two models differ regarding the presence or absence of a robust Quebec Dome during stage 5b. The Stokes et al. model also indicates an early (MIS 4) full confluence between Cordilleran and Laurentide ice sheet and complete incursion of the Canadian prairies already at that time. The volume evolution of the studies show considerable similarity, both agreeing on small and comparable ice sheet volumes through stage 5 and a radical build-up of mass during stage 4. However, they differ in that the Stokes et al. study show more dramatic stadial-interstadial swings in ice volume. This is may be related to their inclusion of surficial and extramarginal (proglacial lakes) hydrology in their model, the effect of which will be that calving in such lakes is an additional fast process for ablation during interstadials.

5.2 Overall glaciation patterns during the three stadials 5b, 4 and 2

When comparing the evolution of the Eurasian and North American ice sheets through the three stadials 5b, 4 and 2, some striking patterns emerge. In northeastern Eurasia, it appears that the ice sheet extent became successively smaller through the three stadials, closely mimicking the diminishing amplitude of the corresponding insolation minima. In contrast, the southwestern extent of the ice sheet increased successively through these three events. Hence, the whole footprint of the ice sheet moved to the southwest. In contrast to the North American situation, a clear highland centre of glaciation existed in the Scandinavian mountains throughout the glaciation cycle. Only during the LGM, did the centre of mass move eastwards, to eastern Sweden and the Baltic depression (Ljungner, 1949; Kleman et al., 1997).

The evolution in North America was quite different. Ice cover during stage 5b was likely to have already been established over the entire NE rim of the ice sheet, from Newfoundland to the Queen Elisabeth Islands. The evolution thereafter was mainly an expansion over the interior continental areas, rapid in the east with a full expansion of the Quebec dome to the vicinity of the LGM margin during stage 4. The expansion in the west is poorly constrained but was apparently much slower. The southern and central prairies were not infilled before stage 2. This pattern indicates that buildup of

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Because of the relatively coarse grid we used, the model does not well reproduce ice streams in peripheral areas, which may lead to an overestimate of marginal ice thickness, the effect of which will lessen inland. We note that the Eurasian Ice Sheet, despite its larger spatial extent, never attains elevations as high as the Greenland Ice Sheet, probably because the Greenland Ice Sheet over much of its perimeter has the margin on its relatively high coastal rim. The same holds true for both proto-Laurentide ice sheets in North America, neither of them is in the model as high as the Greenland Ice Sheet. Only at the LGM does our model show a Laurentide Ice Sheet that reaches above 3200 m, with one central dome. We will not here reiterate the long-standing debate over monodome vs. multidome configuration, but refer to Kleman et al. (2010) for a discussion on this topic. Our model underestimates the LGM ice sheet extent in westernmost Canada, probably due to the long poleward distance in the model domain when the climatic pole is located in east central Greenland. In terms of area and volume this is approximately compensated by an overshoot of the ice margin on the western prairies. The LGM ice sheet reaches the geologically documented maximum southerly extent.

The total modeled Northern Hemisphere ice volume reaches approximately 100 m sea-level equivalent at LGM, thus allowing for an 20 m contribution from Southern Hemisphere. Assuming the constraining footprints of Northern Hemisphere ice sheets to be correct, this constitutes a test on the modeled ice sheet height; any serious over- or underhoot of heights would have shown up as implausible LGM volumes.

6 Conclusions

Using geologically and geomorphologically constrained numerical ice sheet modeling we have explored the emergence of additional Northern Hemisphere ice sheets during the build-up phase of the last glacial cycle. The model credibly reproduces the first-order pattern of size and location of geologically indicated ice sheets during marine isotope stages (MIS) 5b and 4. The differences in ice margin outline between

constraints and model are considered to be small enough that they will not adversely affect the intended use, atmospheric circulation modeling based on the two paleotopographical data sets, for stages 5b and 4 respectively.

The results show that from an interglacial state of two north-south oriented obstacles to atmospheric circulation (Rocky Mountains and Greenland), by MIS 5b the emergence of combined Quebec-Central Arctic and Scandinavian-Barents/Kara ice sheets had effectively increased the number of such highland obstacles to four. This number remained constant through MIS 4, with the most significant change being a large increase in area and southerly extent of the Quebec Dome. By MIS 2, the confluence of the Laurentide and Cordilleran ice sheets had reduced the number to three, albeit one of them with a very large east-west extent.

The controls on growth pattern appears to have been distinctively dissimilar in Eurasia and North America; in Eurasia, the whole footprint of the ice sheet over the three stadials moved in a southwesterly direction, towards the main precipitation source, with only modest and short-lived expansion into the continental interior at the LGM. In North America, the entire northeastern continental rim was glaciated early on, and expansion thereafter was by successive infilling of the continental interior, more rapid in the east than in the west.

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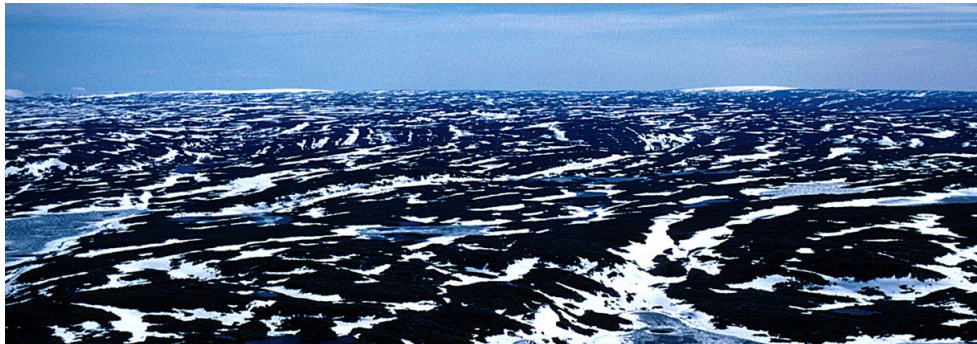


Fig. 1. The Terra Nivea and Grinnell ice caps (on the skyline) rise only about 200 m above the level 600 m elevation plateau of Meta Incognita Peninsula, Baffin Island. The setting is typical of the inception and build-up areas of North American ice sheets in the mainly lowland Laurentide area during the early part of the last glacial cycle.

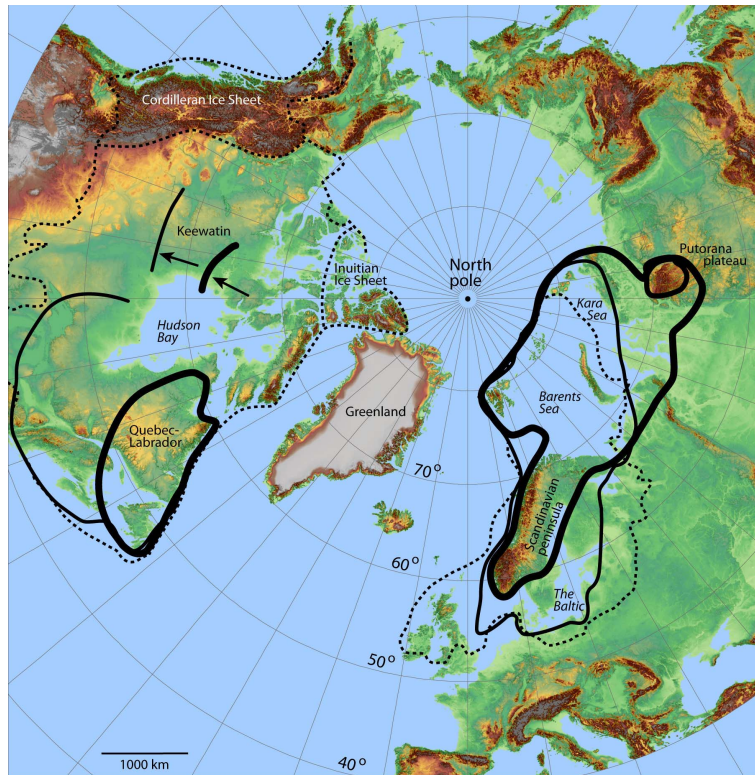


Fig. 2. The geological and geomorphological constraints (ice margin outlines) that were used in modeling paleotopography for glacial maxima during MIS 5b (heavy solid line) and 4 (thin solid line). LGM ice extent is indicated with stippled line. For source data and construction of outlines, see text.

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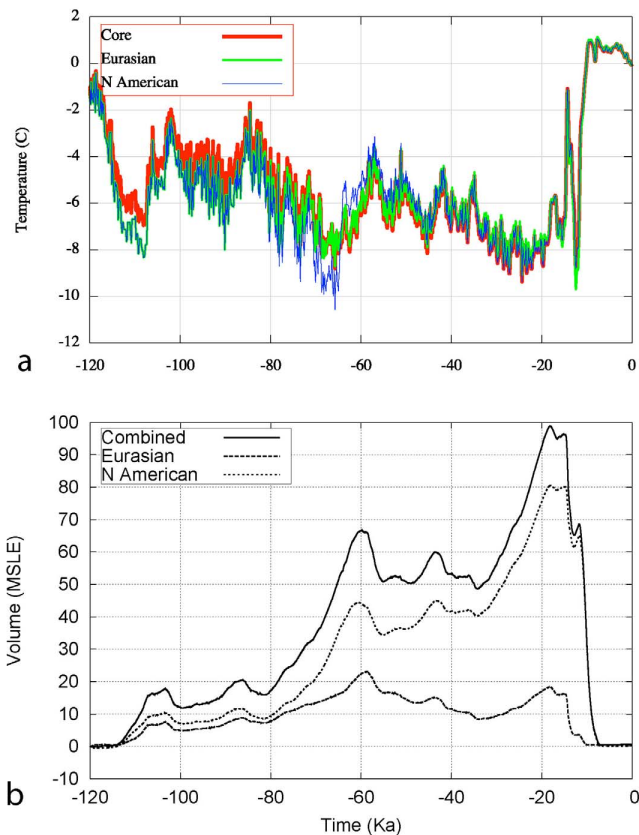


Fig. 3. (a) The temperature forcing, spliced Vostok and GRIP records, used for the modeling. A time-dependent scaling was used to fit ice-sheet size to geological constraints. See text and Fig. 4 for full explanation. **(b)** Modeled ice volumes for the North American and Eurasian domains of the model, respectively.

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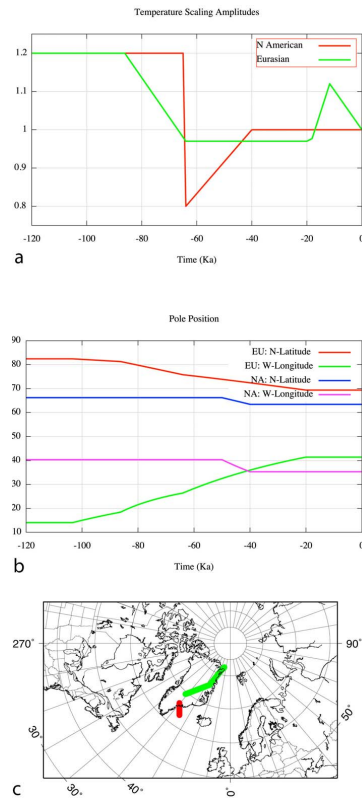


Fig. 4. Primary tuning parameters, **(a)** temperature scaling applied to the spliced Vostok–GRIP record. **(b)** and **(c)**, position of climatic pole. These two parameters were the primary controls used to achieve fit in terms of ice-sheet size and location to the geologically and geomorphologically based constraints in Fig. 3. See text for a full explanation.

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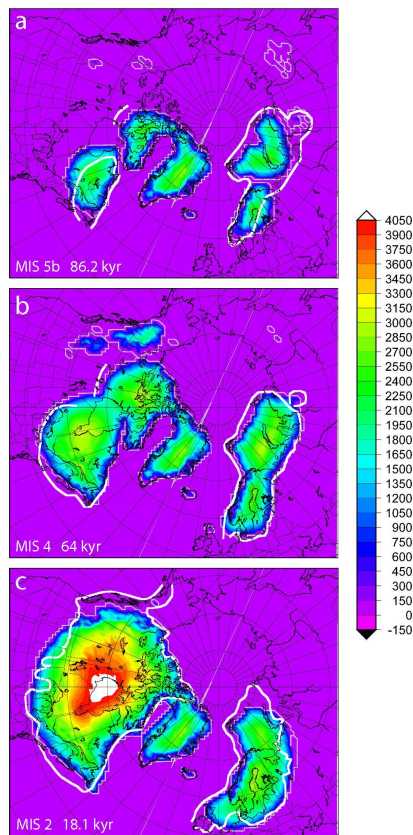


Fig. 5. Modeled ice sheet thickness for MIS 5b (a), MIS 4 (b) and MIS 2 (c), respectively. Geological and geomorphological constraints from Fig. 3 are shown as thick white lines.

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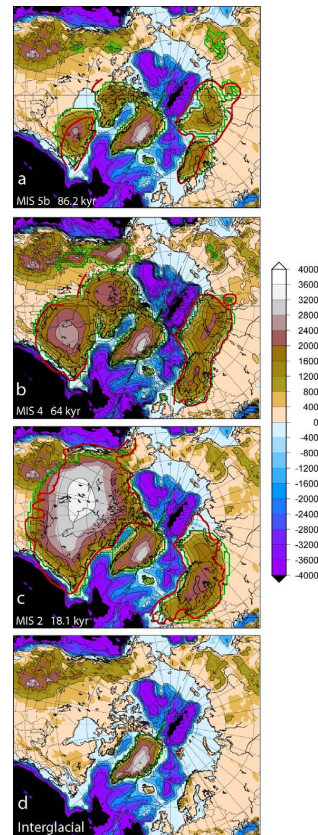


Fig. 6. Terrain topography and modeled ice sheet surface topography for MIS 5b (a), MIS 4 (b), and MIS 2 (c), respectively. (d) Holocene topography shown for reference.

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