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The influence of glacial landscape evolution on Scandinavian Ice Sheet dynamics and dimensions

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Abstract

The Scandinavian topography and bathymetry have been shaped by ice through numerous glacial cycles in the Quaternary. In this study, we investigate how the changing morphology has influenced the Scandinavian ice sheet (SIS) in return. We use a higher-order ice-sheet model to simulate the SIS through a glacial period on three different topographies, representing different stages of glacial landscape evolution in the Quaternary. By forcing the three experiments with the same climate conditions, we isolate the effects of a changing landscape morphology on the evolution and dynamics of the ice sheet. We find that early Quaternary glaciations in Scandinavia were limited in extent and volume by the pre-glacial bathymetry until glacial deposits filled depressions in the North Sea and build out the Norwegian shelf. From middle/late Quaternary (~0.5 Ma) the bathymetry was sufficiently filled to allow for a faster southward expansion of the ice sheet causing a relative increase in ice-sheet volume and extent. Furthermore, we show that the formation of The Norwegian Channel during recent glacial periods restricted southward ice-sheet expansion, only allowing for the ice sheet to advance into the southern North Sea close to glacial maxima. Finally, our experiments indicate that different stretches of The Norwegian Channel may have formed in distinct stages during glacial periods since ~0.5 Ma. These results highlight the importance of accounting for changes in landscape morphology through time when inferring ice-sheet history from ice-volume proxies and when interpreting climate variability from past ice-sheet extents.

1 Introduction

Ice holds the power to transform landscapes and constituted a major geomorphological agent in northern Europe during the Quaternary (last 2.6 Ma) where recurring glacial cycles shaped the present-day landscape. Indeed, the topography and bathymetry in and around northern Europe reveal the extensive impact of its rich glacial history, with deep fjords and U-shaped valleys attesting to the accumulated effect of widespread glacial erosion and terminal moraines indicating the extent of past ice sheets (Hughes at al., 2016). The Eurasian ice sheet complex covered much of the British Isles, all of Scandinavia, and much of northern Europe including parts of Germany, Poland, Russia, and the Baltic through multiple glacial cycles since 1 Ma (Batchelor et al., 2019). During the Last Glacial Maximum (LGM), the complex consisting of the Scandinavian ice sheet (SIS), the Barents Sea ice sheet (BSIS), and the British-Irish ice sheet (BIIS), contained an ice volume corresponding

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to ~18.4±4.9 m sea-level equivalent (Simms et al., 2019). On a global scale, the pace of these glacial cycles results from solar insulation variations combined with feedback mechanisms and internal dynamic effects in the climate system, in part caused by the ice sheets themselves (Hughes and Gibbard, 2018). Differences in ice volume and extent of ice sheets between glacial cycles (Fig. 1) can also be attributed to variations in moisture supply through complex global atmosphere-ocean-ice interactions (e.g., Batchelor et al., 2019; Hughes and Gibbard, 2018), with topography and proximity to the ocean being key factors determining the spatial distribution of moisture to an ice sheet. Studies on glacial landscape evolution have indicated that glacial erosion and deposition can also influence ice-sheet dynamics, ice volumes, and extent (e.g., Kessler et al., 2008; Kaplan et al., 2009; MacGregor et al., 2009; Egholm et al., 2009, 2012a,b, 2017; Anderson et al., 2012; Pedersen and Egholm, 2013; Pedersen et al., 2014; Claque et al., 2020; Mas e Braga et al., 2023). But until now, these studies have been limited to synthetic landscapes and/or limited spatial scales (smaller glaciers and ice caps). A few ice-sheet scale models are starting to consider glacial erosion (e.g., Patton et al., 2022), but the effects of long-term Quaternary landscape evolution on ice-sheet dynamics are still to be explored on a large scales for realistic landscapes and ice-sheet configurations. Understanding the influence of landscape evolution on ice-sheet dynamics requires the reconstruction of landscapes that existed prior to or at earlier stages of glacial erosion, something that can be approached using source-to-sink studies, utilizing off-shore sediment volumes of a glacial origin (e.g., Steer et al., 2012; Paxman et al. 2019; Pedersen et al., 2021).

FIG. 1. Overview map of model domain. Maximum plausible extent of the Fennoscandian ice sheet complex during last glacial maximum (LGM, black line) and penultimate glacial maximum (MIS6, red dashed line) are overlaid (Batchelor et al., 2019) as well as the approximate location of the LGM ice divide position (Olsen et al., 2013).

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In this work, we focus on the well-studied Scandinavian region and investigate how the SIS may have changed its behaviour because of Quaternary landscape evolution. We use a higher-order ice-sheet model to investigate how large-scale glacial morphological features have influenced the development and dynamics of the SIS over a glacial cycle at two key times during the Quaternary: 1) before the inception of major glaciations in the beginning of the Quaternary (PREQ ~2.6 Ma) and 2) during the middle/late Quaternary (MLQ ~0.5 Ma) where major pre-glacial features in the bathymetry around Scandinavia had been filled with glacial deposits (Dowdeswell and Ottesen, 2013). Importantly, we do not intend to reconstruct realistic SIS configurations for these past time periods, but rather keep the climate forcing consistent between experiments, in order to isolate how changes in bed morphology has impacted SIS dynamics and extent. This allows us to i) explore how morphological changes can influence the dynamics, extent, and volume of the ice sheet, independent of the climatic forcing, and ii) gain insight into how ice-volume proxies could be influenced by glacial landscape evolution.

For the early Quaternar, we adopt the pre-glacial landscape reconstructions provided for the Scandinavian region by Pedersen et al. (2021) that include i) removal of glacially generated sediments offshore, ii) infilling of over-deepened fjords and glacial valleys onshore, iii) reconstruction of a wedge of older Mesozoic and Cenozoic sediments on the inner shelf that is assumed to have been eroded by glacial activity within the Quaternary (e.g., Hall et al., 2013), and finally, iv) adjustments of the landscape owing to erosion- and deposition-driven isostatic changes and dynamic topography (Pedersen et al., 2016).

In addition to this pre-glacial reconstruction, that explores an entirely different offshore bathymetry and onshore Scandinavian landscape, we also consider the more subtle effects of the large glacial trough that have been carved into the shelf bathymetry by ice streams since the middle/late Quaternary. One of the most notable of these glacio-morphological features offshore Scandinavia is the Norwegian Channel (Fig. 1). This channel is believed to have been formed by ice-stream activity sometime since 1.1 Ma (e.g., Sejrup et al. 2003), with studies suggesting that ~90 % of the deposits funneled through the channel and into the North Sea Fan were deposited within the last ~0.5 Ma (Hjelstuen et al., 2012). Recently, is has been suggested that the channel formed mostly within the last ~0.35 Ma (Løseth et al., 2022). An erosional unconformity at the base of the channel is draped by post-LGM sediments, suggesting that the channel experienced erosion within the last glacial cycle (Hjelstuen et al., 2012). For the last glacial cycle, it has been proposed that the Norwegian Channel Ice Stream (NCIS) was active in stages but mainly during the LGM (e.g., Sejrup et al., 1998; Sejrup et al., 2003). Ice streaming in the outer parts of the channel near the shelf break started close to the LGM with increased activity promoting ice retreat around 19 ka BP because of the increased ice mass loss (Sejrup et al., 2016). The retreat translated southwards over time as the SIS unzipped from the adjacent BIIS after which with ice streaming was mostly confined to the main trunk of the channel (Sejrup et al., 2016).

2 Methods

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For the numerical experiments presented in this study, we use the depth-integrated second-order shallow-ice approximation iSOSIA (Egholm et al., 2011, 2012a,b). We conduct our experiments by simulating a full glacial cycle of 120 ka on different topographies. In the following section we will present the numerical model, the model setup, and the experimental design.

2.1 Modelling the Scandinavian Ice Sheet

The ice flow in iSOSIA is governed by a second-order approximation of the equations for Stokes flow (e.g., Egholm et al., 2011). The velocities are depth integrated to yield a 2D one layer ice model, implemented here using a regular grid (e.g., Egholm and Nielsen, 2010). The second-order nature of the approximation ensures that ice velocities depend non-linearly on ice thickness, ice-surface gradients, as well as longitudinal and transversal horizontal stress gradients (Egholm et al., 2011, 2012b). Details on the iSOSIA model, including the importance of the higher order ice dynamics involved, have been described in depth elsewhere (Egholm and Nielsen, 2010; Egholm et al., 2011, 2012a,b).

The depth-integrated ice-creep velocity is calculated using temperature-dependent Glen's flow with a stress exponent, n, equal to 3:

$$\varepsilon_{ij} = A_{flow}\tau_e^{n-1}s_{ij},$$

where ε is the strain rate tensor, ij denoting the components of the tensor, A is the ice flow parameter, τ_e is the effective stress and s is the deviatoric stress tensor (Egholm et al., 2011). A simple Weertman sliding scheme is used to calculate the contribution of basal sliding to depth-integrated ice velocities:

$$u_b = A_{sliding}\frac{t_s^3}{N},$$

where u_b is the is the basal velocity, $A_{sliding}$ is an ice sliding coefficient, t_s is the bed parallel shear stress and N is the effective pressure at the base (Egholm et al. 2011). $A_{sliding}$ is chosen to give realistic sliding in the order of several hundred meters per year for example in fjords or near the shelf edge in the Norwegian Sea, similar to surface velocities in comparable areas of modern-day ice-bodies (e.g., Millan et al. 2022) . To allow for faster ice flow for soft bed subglacial conditions (e.g., Gladstone et al., 2020, Han et al., 2021), $A_{sliding}$ is enhanced by a factor of 5 in offshore regions and onshore in northern Europe where thick soft sediments cover the bed.

In this study, we focus on grounded ice only, as ice-shelf dynamics are computationally expensive to resolve on the timescales of our experiment and because constraints on ice shelf extent in middle or early Quaternary glaciations are sparse due to a lack of reliable dates on submarine landforms (e.g., Jakobsson et al. 2016). Some older studies suggest that an ice shelf was present during recent glaciations in the North Atlantic and

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Arctic regions (Hughes et al. 1977, Lindstrom et al. 1986). However, while ice shelf stability is sensitive to bathymetric configurations (Bart et al. 2016) and is a deciding factor in grounding line migration, we limit our focus here to large-scale morphological features, such as the Norwegian Channel, created by an ice stream in contact with the sea bed (Sejrup et al., 2016). Consequently, we do not consider floating ice in our simulations and remove floating ice by introducing a fast melt rate for ice that does not meet the grounding criterion:

$$H_{ice} > (SL + H_{ice})\frac{\rho_{water}}{\rho_{ice}},$$

where H_{ice} is ice thickness, SL is local sea level and ρ_{water} and ρ_{ice} are the densities of water and ice, respectively. Special boundary conditions are employed at the approximate locations where the SIS meets the BSIS and BIIS by introducing an 'ice wall' where the ice flux is zero to emulate divergent ice flow when these ice sheets merge during glacial maxima. At the edges of the model domain, we employ open boundary conditions to allow for ice to flow out of the domain. Common model parameters are presented in Table 1.

Parameter		**Value**	**Unit**
ρ_{ice}	Ice density	910	kg m^{-3}
q_b	Geothermal heat flux	0.045	W m^{-2}
L_i	Latent heat of ice	334	kJ kg^{-1}
A_{flow}	Ice flow parameter	$1.5 \cdot 10^{-15}$	Pa^{-3} y^{-1}
$A_{sliding}$	Ice sliding parameter	0.4	m Pa^{-2} y^{-1}
n	Ice flow exponent	3	
$f_{flow\ enhancement}$	Ice flow enhancement factor	100	
m	Ice sliding exponent	3	
$F_{sliding\ enhancement}$	Sliding enhancement factor offshore	5	
mPDD	PDD factor	0.005	m °C^{-1} d^{-1}
SL	Mean sea level	[0 : -130]	m
dT_h	Lapse rate	6.5	°C km^{-1}
$dT_{m,e}$	Easterly temperature gradient	$[-1.3 : -2.3] \cdot 10^{-6}$	°C m^{-1}
$dT_{m,n}$	Northerly temperature gradient	$[-3.5 : -10] \cdot 10^{-6}$	°C m^{-1}
$dA_{T,e}$	Easterly annual temperature variation gradient	$[7.8 : 0.11] \cdot 10^{-6}$	°C m^{-1}
$dA_{T,n}$	Northerly annual temperature variation gradient	$[2.0 : 1.4] \cdot 10^{-6}$	°C m^{-1}
dP/dT	Change in precipitation with change in temperature	0.029	°C^{-1}
D_L	Thickness of elastic lithosphere	50	km

TABLE 1. Common parameters in the ice sheet model and mass balance scheme. Numbers in brackets denote min and max values.

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2.1.1 Mass balance

In the simulations we present here, we assume that the mass balance ($\dot{M}_{ice}$) of the ice sheet can be approximated using three components:

$$\dot{M}_{ice} = \dot{m}_{acc} - \dot{m}_s - \dot{m}_b,$$

where $\dot{m}_{acc}$ is the rate of accumulation, $\dot{m}_s$ is the surface melt rate and $\dot{m}_b$ is the basal melt rate (Egholm et al. 2012b). We use a positive-degree-day (PDD) model to estimate accumulation rate and surface melt rate as a function of mean annual temperature, annual temperature variation, and mean annual precipitation at every point in our model domain for every time step (e.g., Magrani et al., 2022).

The yearly temperature variation in a given cell, is approximated by a sine function based on the mean annual temperature and annual temperature amplitude (see below). The melt rate in m/yr is calculated in the PDD model as:

$$\dot{m}_s = m_{PDD} \sum_{n=1}^{365} T_{positive},$$

where m_{PDD} is the positive-degree-day factor multiplied with the sum of positive degrees each year. Here, we consider a single melting degree factor for both ice and snow, since all precipitation is turned into ice after accumulation (based on yearly average rates). The accumulation rate is approximated by:

$$\dot{m}_{acc} = \frac{n_{frost}}{365} \cdot P,$$

where n_{frost} is the number of days with negative temperatures in a year and P is the annual precipitation. The temperature forcing that drives spatial and temporal changes in mass balance in our simulations is based on mean temperature, annual temperature amplitude, and lapse rate that vary across the model domain using spatial gradients that vary in time. Two climate states are chosen to represent the extremes of our model: a glacial maximum state and an interglacial state, and the spatial gradients of the full glacial cycle of our model simulations are subsequently defined to vary in between these extremes using a glacial index that resembles the normalized LR04 Benthic Stack (Lisiecki and Raymo, 2005) with glacial maximum in this climate forcing occurring at 18 ka BP. Here we define spatial (x, y, z) gradients at the glacial maximum using multiple linear regression on MPI-ESM climate model outputs (LGM experiment; Jungclaus et al. 2019). For the interglacial state we define spatial gradients using the ERA-interim reanalysis data for modern day (Dee et al. 2022). Finally, the lapse rate was found to be close to constant, so we keep this fixed at 6.5 °C km^{-1}. With this approach, the temporally-varying temperature forcing of the entire grid can be defined from a single grid cell in the lower left corner and still capture a coastal-continental (east-west) gradient, a polar gradient (south-north), and an altitudinal gradient (lapse rate) in temperature. However, we cannot capture local effects that arise from changes in complex atmospheric circulations patterns over time that might have important implications for glacial dynamics and ice extent (e.g., Liakka et al. 2016, Hughes and Gibbard, 2018).

To represent precipitation in our simulations, we use a climate-corrected modern-day mean precipitation field (Pendergrass et al., 2022), modulating the local precipitation in every grid cell using the following equation:

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$$P = P_0 \cdot e^{kTp \cdot \Delta T},$$

where P_0 is the local modern day (interglacial) precipitation, ΔT is the change in temperature in a cell from the previous time step, and kTp represent the rate of change in precipitation for a change in temperature with a value of 0.029 °C^{-1}. The value of kTp is found by optimization through a comparison between mean precipitation at LGM in MPI-ESM and mean precipitation in the modern-day ERA-interim data set. By scaling the precipitation with changes in temperature we can capture some of the effects an ice sheet will impose on moisture supply, by limiting snow fall in the central parts of the ice sheet (Fig. 3D).

Basal melting rate is calculated as the difference between of geothermal heat flux from the bed and the heat flux from the temperature gradient in the basal ice (Egholm et al., 2012a):

$$\dot{m}_b = \frac{q_b - q_c}{\rho_{ice} L_i},$$

where L_i is the latent heat for fusion of ice and ρ_{ice} is the density of ice (Tab. 1).

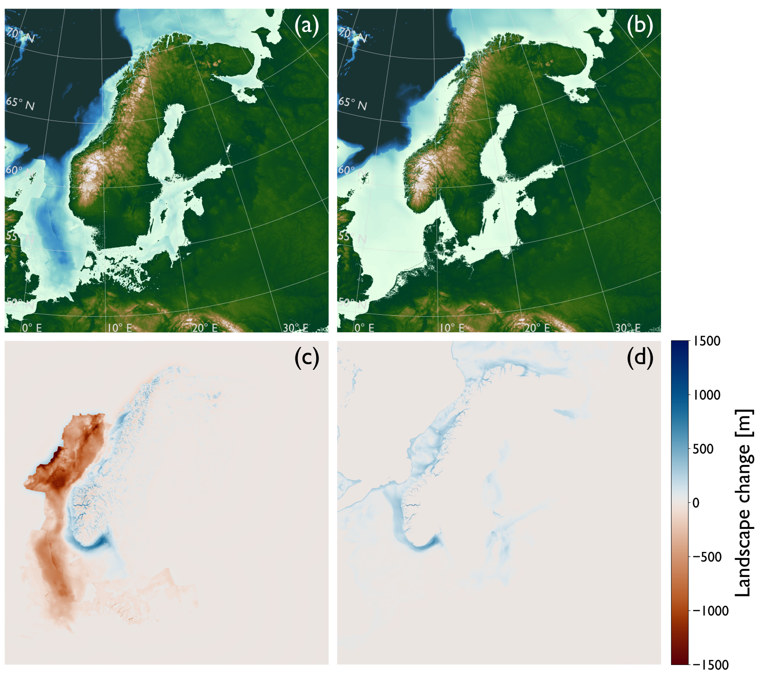

FIG. 2. Paleo-topographic and bathymetric reconstructions. a) the PREQ experiment, b) the MLQ experiment, c) and d) show the differences between the panel above and the modern-day topography and bathymetry.

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2.1.2 Topography and bathymetry

The focus of this study is to examine the influence of bed topography on ice sheet behaviour, exemplified by simulating the SIS on landscape configurations representing different periods in the Quaternary. For comparison, we simulate the SIS on modern-day topography and bathymetry over the last glacial cycle in a reference model. The reference experiment uses the global DEM GEBCO 2022 (GEBCO Bathymetric Compilation Group., 2022) global grid sampled at 10 km x 10 km for the ice model (the same grid resolution is used in all experiments). Because of computational limitations, a model resolution higher than 10 km is not feasible. Having a higher resolution would allow us to resolve glacial morphology in higher detail and could lead to interesting findings regarding the influence of fjord systems in western Norway on ice sheet dynamics. Here, we focus on larger features such as the Norwegian Channel where a 10 km resolution is sufficient. Throughout the model simulations, ice-driven isostasy is handled with a two-dimensional uniform thin elastic plate model (e.g., Pedersen et al. 2014).

The pre-glacial landscape is adopted from Pedersen et al. (2021) and reconstructed using a source-to-sink approach that also considers i) a component of glacial erosion that has taken place on the inner shelf, ii) erosion-driven isostasy, and iii) a component of dynamic topography (Pedersen et al., 2016). For further details on the approach see Pedersen et al. (2021). Here, we extend these previous reconstructions and remove the Quaternary sediment package from all sectors of the North Sea, to reconstruct a realistic pre-glacial bathymetry for the entire region (Binzer et al., 1994; Rise et al., 2005; Nielsen et al., 2008; Gołędowski et al., 2012; Lamb et al. 2018; The Southern Permian Basin Atlas). These additional sediment volumes, from outside of the Norwegian and Danish sectors, are not included in the landscape reconstruction onshore Scandinavia. The result is a landscape representing a pre-glacial state before any major glaciations in Scandinavia, featuring a large submarine depression in the North Sea and a much narrower continental shelf along the Norwegian margin than at present (Fig. 2a,c). In addition to the PREQ experiment two sub-experiments are presented: 'PREQ-onshore' where only the onshore fjord erosion has been reconstructed (material added compared to present-day) and 'PREQ-offshore' where only the offshore deposition has been reconstructed (material removed compared to present-day). Neither of these additional sub-experiments considers the offshore sediment wedge on the shelf. With the sub-experiments we can assess which processes control the behaviours and ice volume changes observed in the PREQ experiment.

For the middle/late Quaternary (MLQ) experiment, we reconstruct the bathymetry by estimating the volumes of erosion that have been carved into the modern day sea-bed by ice streams on the Norwegian shelf and in the Norwegian Channel (Fig. 1). This bathymetric erosion is estimated using the geophysical relief method (e.g., Steer et al., 2012; Pedersen et al., 2021) on the present-day GEBCO 2022 global DEM (GEBCO Bathymetric Compilation Group., 2022), using a grid resolution of 1 x 1 km and a sliding window radius of 35 km. The resulting filled bathymetry, that also fills fjords to sea level, is adjusted with the flexural isostatic response to loading using gFlex 1.1.1 (Wickert, 2016) with an effective elastic thickness of 15 km. This reconstruction of the Scandinavian morphology is meant to represent a state before the formation of the

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Norwegian Channel (Fig. 2b,d) and could represent an age of approximately ~0.5 Ma. This approximate age is supported by the presence of buried mega-scale glacial lineations and drumlins in stratigraphic sequences of the North Sea suggesting that grounded ice has been present since ~0.5 Ma, whereas the lack of these features in the older strata indicate that early Quaternary glaciations did not ground, but only supplied icebergs to the North Sea (Dowdeswell and Ottesen, 2013; Rea et al., 2018).

3 Results

In this section we start by presenting the results from our reference model simulating the evolution of the SIS on the present-day topography and bathymetry over the last glacial period. Then we present the results of our two experiments with reconstructed topography and bathymetry and how they differ from the reference model. Lastly, we present our findings regarding a possible timing of formation for the Norwegian Channel.

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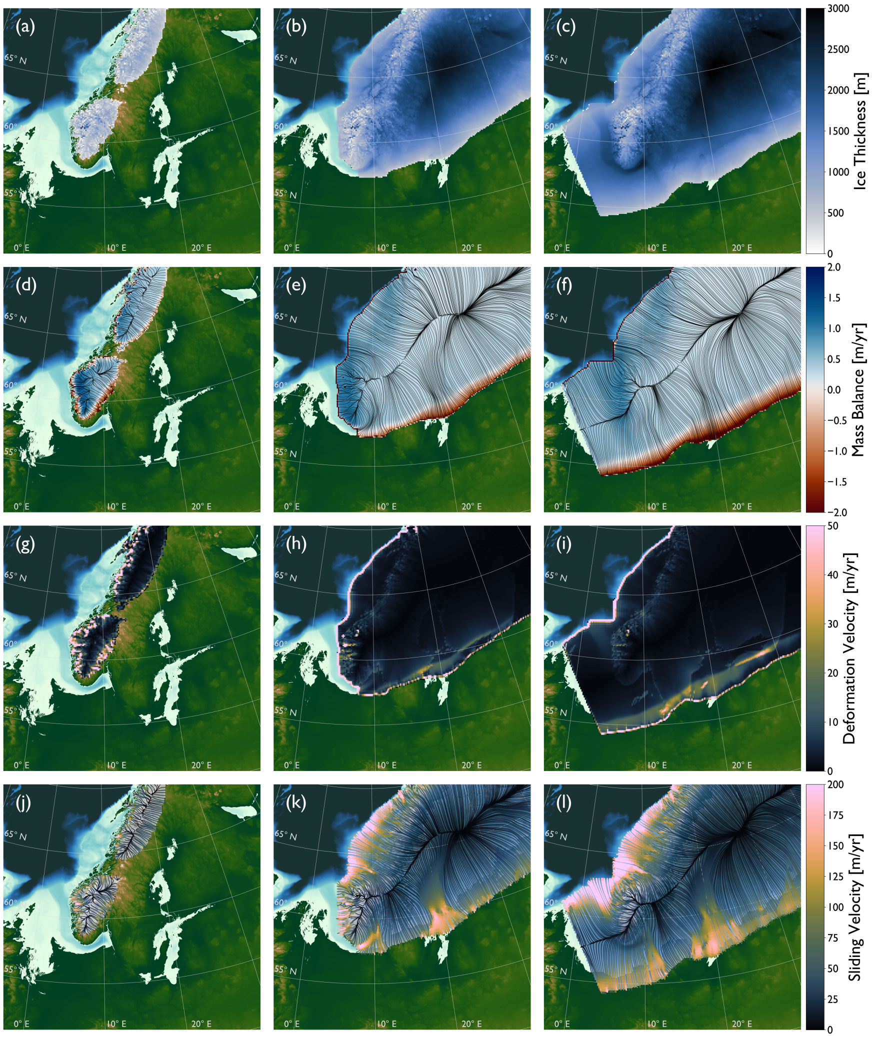

FIG. 3. Model output from three time slices of the reference experiment, left column: early glaciation (72 ka), middle column: late-intermediate glaciation (22 ka), right column: glacial maximum (17 ka). a-c) ice thickness, d-f) mass balance, g-i) depth averaged deformation velocity and j-l) sliding velocity.

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3.1 Reference model

To illustrate the spatial and temporal development of the SIS in our model simulations, we present model output from three snapshots in time (Fig. 3): minor ice build-up during early glaciation (72 ka), moderate glacial build-up during intermediate times of the glaciation (22 ka), and glacial maximum that happens in these simulations at 17 ka. We note that the delayed glacial maximum timing in our models compared to the timing of the reconstructed maximum extent in Scandinavia (~21-19 ka, Hughes et al., 2016) is a direct consequence of the chosen climate forcing, utilizing a glaciation index that peaks at 18 ka. We do not intend here to match the exact timing of the maximum extent (LGM). During our simulated early glaciation, ice extent is limited to mountain regions with high topography and high latitude regions in Norway and Sweden (Fig. 3a). Mass balance is positive ~1.5 m/yr in high altitude regions at the Norwegian coast where precipitation is high, and temperatures are low (Fig. 3d). Ice deformation and sliding is high up to >50 m/yr and >200 m/yr respectively, during early glaciation (Fig. 3g,j), where ice is thin and controlled by the underlying topography that includes mountainous regions dissected by fjords and valleys. During the intermediate glaciation, the ice sheet has advanced onto the shelf region, with grounded ice on the Norwegian margin, and the ice sheet has started to advance into the North Sea through the inner part of the Norwegian Channel (Fig. 3b). The mass balance reaches ~1 m/yr at the west coast of Norway (Fig. 3e), with values across most of the ice sheet <0.5 m/yr, and negative mass balance at the south/western margin reaching ~-2 m/yr where the ice is thin and velocities exceed ~200 m/yr (Fig. 3h,k). Along the coastal margin to the west, the mass balance is negative in a narrow zone where floating ice is melting fast. Sliding is notably high, reaching >200 m/yr in the inner parts of the Norwegian Channel (Fig. 3k). The ice flow is still steered by topography in the high regions of Southern Norway, and also in the Bothnic Bay, whereas the main divide in Northern Scandinavia has shifted east, being largely independent of the underlying topography (Fig. 3e). During glacial maximum, the ice sheet reaches a thickness of >3000 m in the central parts (Fig. 3c) with a relatively low positive mass balance along the west coast of Norway (<1 m/yr; Fig. 3f) with the same general spatial pattern in accumulation and ablation as the intermediate glaciation (Fig. 3e) across the ice sheet as during the intermediate glaciation. Sliding is high along the northeastern margin of the ice sheet (>200 m/yr) especially near the shelf break where ice is funnelled towards the deeper ocean (Fig. 3l). For a while (~5,000 yrs) during the maximum expansion, the ice sheet merges with the BIIS in the western part of the North Sea, simulated as an ice wall (Fig. 3f,l). At this time, the ice flow rearranges into a divergent pattern from the ice saddle that emerges between the BIIS and the SIS. Consequently, the ice flows across the Norwegian Channel during the maximum extent of the ice sheet instead of being focused in the channel itself, as the ice is diverged southward, driven by the surface slope of the ice sheet under this ice configuration (Fig. 3l). It is worth noting that the reference model captures a realistic placement of the LGM ice divide (Fig. 3f) in accordance with geological observations (Fig. 1; Olsen et al., 2013). The glacial maximum ice extent in our reference experiment is within the maximum LGM ice extent (Fig. 1; Hughes et al., 2016), albeit with less ice towards the southern margin and more ice in northeast.

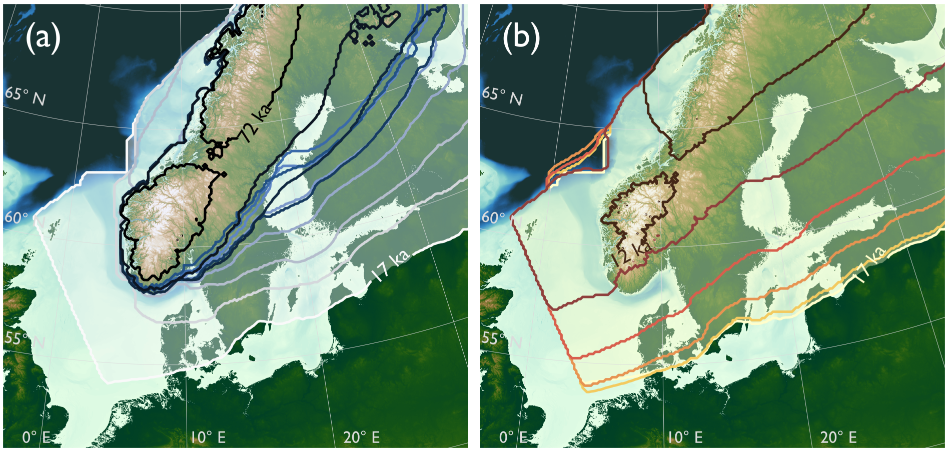

FIG. 4. Advance and retreat of the SIS in the reference experiment. A) ice advance in 5 kyr intervals between model years 72 ka and 17 ka. B) retreat in 1 kyr intervals from 17 to 12 ka.

Buildup of the SIS from early mountain glaciation to glacial maximum happens gradually with grounded ice on the Norwegian shelf forming 10.000 model years before glacial maximum, and ice advance in the North Sea occur over just 5000 model years approaching glacial maximum extent (Fig. 4A). In contrast, the ice retreat is rapid with ice mass loss from the glacial maximum back to a state similar to early glaciation happening over just 5000 model years (Fig. 4B).

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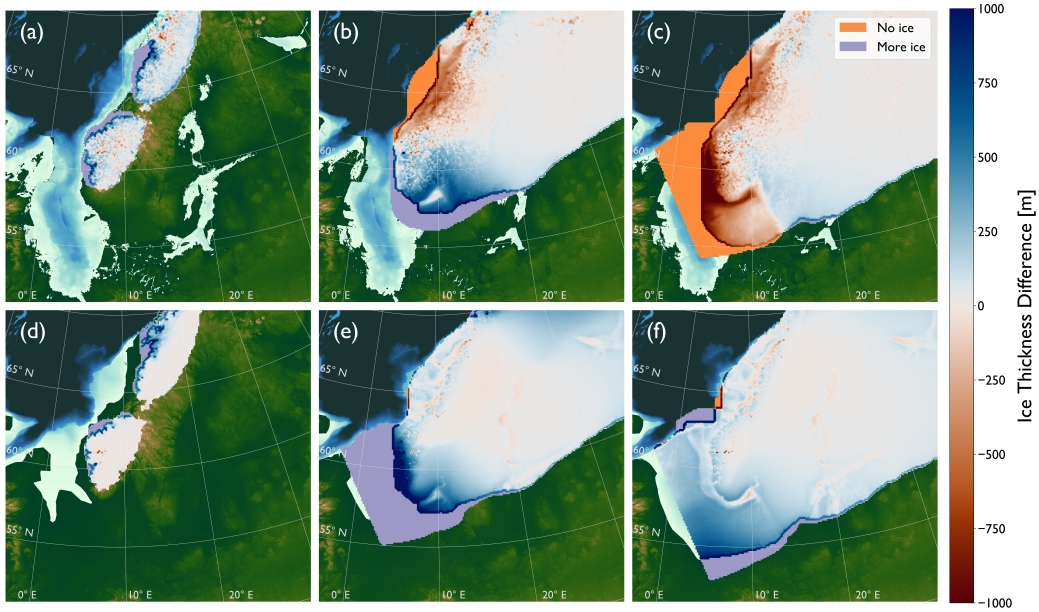

FIG. 5. Differences in ice thickness for the a,b,c) PREQ and d,e,f) MLQ experiments compared to the reference experiment for the same time slices as Fig. 3. Blue colors means more ice in this experiment than the reference experiment and red colors means less ice.

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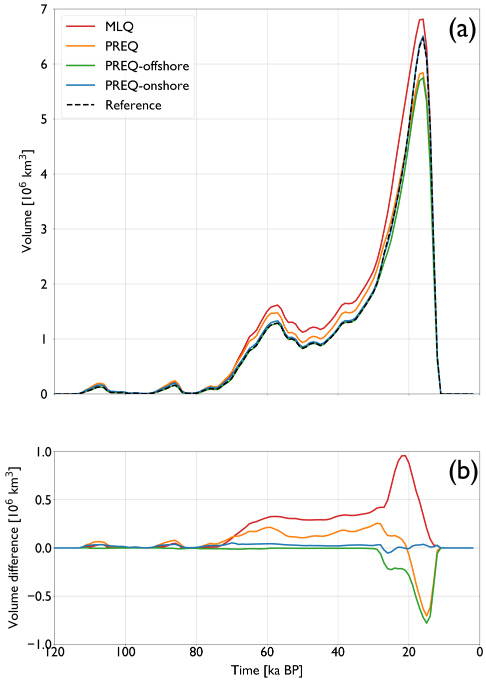

FIG. 6. a) ice volume over time for the different experiments, b) volume differences between the different experiments and the reference experiment.

3.2 Results from PREQ and MLQ

In the model simulation representing ice-sheet behavior on an early Quaternary landscape morphology (PREQ; Fig. 2a, Fig. 5a-c), the ice sheet initially extends further than the reference model (Fig. 5a, purple color), particularly towards the Norwegian coast. At the intermediate stage (Fig. 5b), the ice sheet shows a smaller extent and thickness towards the Norwegian margin (Fig. 5b, orange color), whereas the ice extends further towards the south (Fig. 5b, purple color) with an ice thickness increase of >500 m in some regions. The location of the present-day Norwegian Channel shows a much thinner ice since this bathymetric depression is not present in the PREQ landscape reconstruction (Fig. 5b). At the maximum extent, the ice sheet is smaller both along the western and the southwestern margins (Fig. 5c, orange color), with a general decrease in ice sheet thickness (Fig. 5c, red colors). The reduced extent and ice volume during the maximum extent result in ~10 % lower maximum ice volume than the reference model (Fig. 6, orange curve). The large difference in ice volume between the PREQ experiment and the reference experiment is largely driven by differences in bathymetry (PREQ-offshore; Fig. 6a, green curve) as changes in topography do not lead to significant differences in ice volume compared to the reference model (PREQ-onshore, Fig. 6a, blue line)

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For the MLQ simulation that represents ice flow on a landscape morphology that existed prior to extensive erosion of the bathymetry by ice streaming (Fig. 2b, Fig. 5d-f), the ice sheet also starts slightly larger (Fig. 5d, purple color) compared to the reference model. At the intermediate stage, the ice sheet has already extended all the way across the North Sea (Fig. 5e, purple color), showing also a significantly thicker ice sheet in the adjoining regions onshore Scandinavia. This trend is continued during the maximum extent, where the MLQ ice sheet extends even further, particularly towards the south (Fig. 5f, purple colors). In general, the extent of the MLQ ice sheet is not changed along the Norwegian margin, where the width of the shelf has not changed for this simulation (Fig. 5e-f). The increased ice extent and ice thickness in the MLQ simulation result in a maximum ice volume that is ~25 % more than the reference model during intermediate stage and ~5 % during the glacial maximum as a direct result of the changed bathymetry (Fig. 6, red line).

3.3 Sliding in the Norwegian Channel

The erosive power of ice is a product of ice flux over a region with grounded ice (Patton et al. 2022) and is strongly correlated with ice sliding velocity (Cook et al. 2020), which means sliding velocity can be considered a proxy for erosive potential. Here we explore whether our higher-order ice-sheet model can capture the erosive potential through sliding in the Norwegian Channel in the present-day bathymetry of the reference model and whether the model can predict erosion when the channel is not there in the PREQ and MLQ experiments. The ice dynamics in our reference simulation show significant sliding in the Norwegian Channel in four distinct phases (Fig. 7a-d). In the early glacial stage, the ice is sliding fast southeast of southern Norway along the deepest part of the channel (Fig. 7a). As the ice approaches maximum extent, the sliding pattern changes because of the different ice flow patterns that arise as an ice saddle emerges in the North Sea when the SIS merges with the BIIS (Fig. 7b). At this stage, ice flows south across the channel from the southern mountains of Norway, following the steepest surface gradient of the ice sheet. Instead, sliding is now mostly concentrated in the outer parts of the Norwegian Channel close to the North Sea Fan (Fig. 7b). During retreat, ice sliding continues in the outer parts of the channel, but also becomes prominent along the southern tip of Norway with ice sliding towards the southeast, and in the inner parts of the channel near Oslo Fjord (Fig. 7c). Finally, as the ice sheet retreats further, continued sliding toward the North Sea Fan is complemented by a phase of southward sliding in the channel along the south-western coast, a region that had not seen significant prior sliding (Fig. 7d). In figure 7e-h we show the same time slices for the MLQ experiment. Here, the ice extends further towards the west and has already formed a saddle between the SIS and the BIIS during the initial phase of the glacial cycle (Fig. 7e), and sliding is high towards the shelf break in the region that will later become the outermost part of the Norwegian Channel. Sliding velocities towards the shelf break are consistently high throughout the model simulation (Fig. 7f,h), whereas sliding accelerates in the inner parts of what will become the Norwegian Channel during ice retreat (Fig. 7g). In the last time slice, sliding velocity is lower than the reference experiment but has the same general pattern (Fig. 7h), with sliding in some regions along the west coast of Southern Norway. In figure 7i-l we present the time slices for the PREQ experiment. Across all four panels the patterns differ from the reference and MLQ experiments. Instead, we observe high sliding velocities

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towards the west across where the channel is today (Fig. 7i-k). In the last time slice we observe very little sliding as the ice has retreated mostly onshore at this time in the PREQ experiment (Fig. 7l).

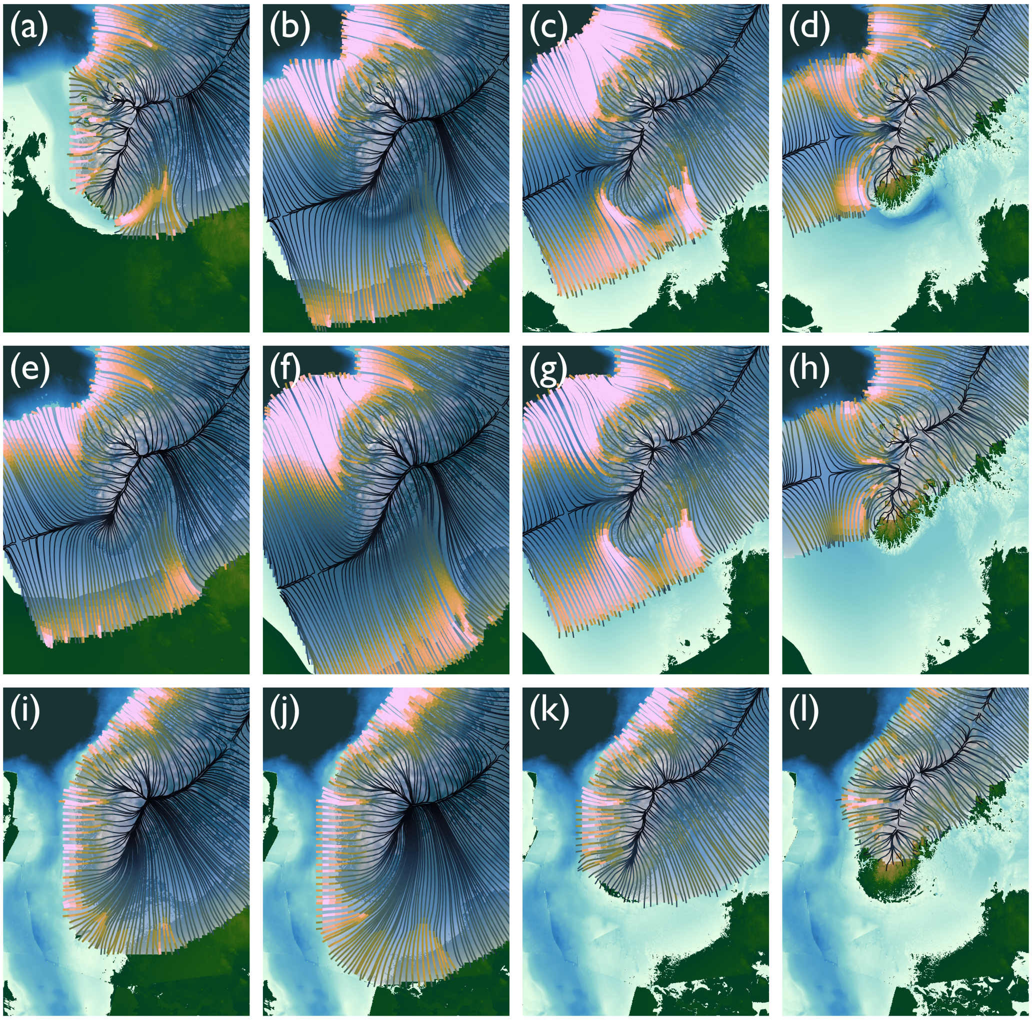

FIG. 7. Sliding velocity in southwestern Norway for reference model year a) 23 ka, b) 17 ka, c) 14 ka, and (d) 13 ka. Same color scale as Fig. 3j-l. Same for MLQ experiment e-h and PREQ experiment i-l.

4. Discussion

4.1 Ice extent and volume

The volume of ice contained in the Scandinavian ice sheet at LGM is estimated to be between 5.3-6.5 10^6 km^3 (Hughes et al. 2016). This is in good agreement with our reference experiment that reaches 6.5 M km^3 at glacial maximum. The ice divide of the SIS in the reference experiment is in good agreement with observations (Fig. 1, Fig. 3f) which also affirms that our model captures an adequate representation of the ice sheet during the last glacial period. The differences in maximum ice extent between our reference experiment and observations (Hughes et al., 2016; Fig. 1) can be attributed to the simple mass balance implemented in our model using linear gradients that does not capture the complex nature of the regional climate during the last glacial cycle but is an adequate approximation for our purposes. Geological observations suggest that the main ice advance in Denmark approaching glacial maximum between 20-22 ka came from the northeast bringing till deposits of Middle Swedish provenance (Houmark-Nielsen, 2004), whereas the main ice advance into Denmark in our reference experiment comes from the north (Fig. 4a). A possible reason that our model does not capture this dynamic in the southerly ice advance could be the lack of subglacial hydrology in the model which can increase sliding rates (Egholm et al. 2012a). It could also be the lack of a more complex stress dependent ice viscosity, where the Glen's flow law stress exponent can increase to $n \approx 4$ in some areas, which can increase the flow velocity by an order of magnitude (Millstein et al., 2022). These effects could be important especially in the southern parts of the ice sheet where the ice is thin and fast flowing during advance (Fig. 3c,i). Here, an even faster and thinner ice might be more sensitive to the low relief topography of southern Scandinavia leading to a more westerly ice flow from Sweden into Denmark in agreement with the observations.

We cannot directly compare the ice extents in our experiments with reconstructions of past SIS extent as we use the same climate forcing between experiments, but we can assess whether differences in past ice sheet extents follow the same trends as we see in this study that is based solely on differences in morphology. Batchelor et al. (2019) use empirical data to evaluate past northern hemisphere glacial extents, and suggest best-estimate maximum southern extents of the MIS 12 (429-477 ka), MIS 16 (622-677 ka), and MIS 20-24 (790-928 ka) ice sheets to be somewhere between the best-estimate maximum MIS 6 extent and the LGM extent (Fig. 1; dashed red line, black line, 132-190 ka), although the MIS 16 and MIS 20-24 maximum ice sheet extents are highly uncertain. These reconstructions are based on very limited observations and in some cases (e.g., MIS 12 and 16) the estimates are mostly based on similarities in the δ^{18}O curve (Batchelor et al., 2019). We show with this study that purely morphological differences in bathymetry between the last glacial period and ~0.5 Ma ka (MLQ experiment, similar in time to MIS 12/16) allow for larger ice-sheet extents simply owing to geomorphic changes during this time period. This suggests that both climatic and topographic forcing might have caused these (possibly) large ice extents of the mid-late Quaternary (MIS 12,16,20-24). Indeed, our results showcase that a smooth bathymetry in the North Sea region (i.e., lacking glacial morphology), such as before the inception of the Norwegian Channel, could lead to earlier and more extensive southerly ice advance within a glacial period (Fig. 5e). On the other hand, our simulation of early Quaternary glaciations suggests that ice buildup across the North Sea was not plausible at this early stage of glacial landscape evolution. Indeed, in the PREQ experiment we find that the SIS could extend no further than the continental shelf during the early Quaternary (Fig. 5b,c). This is consistent with a study of buried glacial landforms in the central North Sea documenting ice-berg plough marks in early Quaternary sediments (Dowdeswell et al., 2013). Our reconstructed early Quaternary ice sheet would have supplied icebergs that created these plough marks.

The differences we find in ice volume at the maximum glacial extent (~5 % higher for MLQ, ~10 % lower for PREQ), illustrate how differences in morphology affects ice volume independent of the climate forcing. This has implications for the proxies we use for ice volume history. Looking at the peak values in the LR04 Benthic Stack for LGM (5.02 ±0.03 ‰), MIS 6 (4.98 ±0.05 ‰), MIS 12 (5.08 ±0.05 ‰), MIS 16 (5.08 ±0.06 ‰), and MIS 20-24 (4.69 ±0.08 ‰), the proportional differences between these δ^{18}O peaks (~1-7 %) are less than the proportional differences of 5-10% in peak ice volume between the model simulations presented here, suggesting that landscape evolution can play an important role in controlling ice volume. Clearly the effect of glacial morphology is local whereas the LR04 Benthic Stack is a global proxy and in addition, local ice volume also depend on global atmospheric circulation patterns which can lead to asynchronous development of the ice sheets during a glacial period (e.g., Liakka et al., 2016) that will also influence ice-sheet volume between glacial cycles. Nevertheless, according to this study landscape morphology alone can account for ~10 % (~25 % during ice builp-up) difference in ice volume between glacial cycles for the Scandinavian region implying that glacial landscape evolution could be an overlooked mechanism impacting global ice volume and thereby δ^{18}O values.

4.2 Formation of the Norwegian Channel

It is uncertain how and when the Norwegian Channel was formed, with studies estimating the time of formation to be between 0.35-1.1 Ma – with more recent studies suggesting younger ages (e.g., Sejrup et al., 2003; Hjelstuen et al., 2012; Løseth et al., 2022). In this study, we have assumed that the entirety of the Norwegian Channel formed after 0.5 Ma (MLQ).

A previous modelling study suggests that the NCIS was active in stages with streaming in the inner parts of the channel leading up to, and deactivated during, glacial maximum because of the saddle forming from the merging of the BIIS and the SIS (Boulton and Hagdorn, 2006). In that study, the NCIS was mostly active near the shelf break at LGM and during retreat the ice stream funnelled ice along the entire length of the Norwegian Channel (Boulton and Hagdorn, 2006). We find in this study, that ice streaming was active in the inner parts of the channel before the saddle formed between the BIIS and the SIS, after which ice streaming velocity increased dramatically in the outer parts of the Norwegian Channel near the shelf break and mostly deactivated in the inner parts of the Norwegian Channel, consistent with other literature based on observations of e.g. subglacial landforms combined with dated sediment cores (Sejrup et al., 2016). However, our reference experiment does not mimic at any time an NCIS spanning the entire trunk of the Norwegian Channel, which would significantly contribute to ice mass loss from rapid grounding line retreat as is supported by observations

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(Sejrup et al., 2016). We cannot with this model setup rule out the occurrence of continuous ice streaming in the entire Norwegian Channel after the LGM. Indeed, some processes central to reproducing realistic ice stream behaviour is not included in iSOSIA, including enhanced basal melt owing to basal friction, leading to accelerated thinning in regions with rapid ice sliding as well as effects of internal friction and temperature on ice viscosity which can greatly amplify sliding velocities (Millstein et al., 2022; Bondzio et al., 2016). These mechanisms could contribute to highly elevated sliding velocities, especially in the NCIS, and could facilitate a propagation of the streaming activity in the outer parts of the channel to the inner parts.

Despite the channel being filled with sediment in the reconstructed bathymetry in the MLQ experiment we find an ice streaming pattern similar to that of the reference model, with even higher sliding in the outer parts of the Norwegian Channel from 23-17 ka near the shelf break because of the faster advance in the North Sea in this experiment. In the PREQ experiment streaming is limited to a pattern across channel towards the west because of the sediment wedge sloping along the shoreline leading the ice towards the middle North Sea. We find it likely that the channel could initially have been formed during multiple glacial periods since ~0.5 Ma before the main formation occurred in recent glacial periods (~0.35 Ma; Løseth et al. 2022). This would be in agreement with studies on the North Sea Fan (NCIS depocenter), suggesting that 90% of the sediments in this fan are younger than 0.5 Ma (Hjelstuen et al., 2012).

5. Conclusion

We have used a higher-order ice sheet model to investigate the effect of landscape morphology on the SIS evolution and dynamics. Three different experiments where conducted: (i) a reference experiment resembling the last glacial cycle using modern-day topography and bathymetry, (ii) a mid-late-Quaternary (MLQ) experiment with glacial morphological features in the present-day bathymetry filled with sediment, and (iii) a pre-Quaternary (PREQ) experiment, simulating the SIS on a reconstructed pre-glacial topography and bathymetry. We find in the MLQ experiment that removing glacial morphological features in the bathymetry allows for faster and further southward expansion at similar climatic conditions allowing for a larger ice sheet. On the contrary we find in the PREQ experiment that the early Quaternary bathymetry did not allow for the SIS to advance as far westward, thereby limiting the size of early glaciations and preventing a merge between the BIIS and the SIS. Looking at the prominent glacio-morphological feature, the Norwegian Channel, we find that the PREQ experiment do not allow for significant ice streaming in this area and that the channel was more likely formed since ~0.5 Ma. Furthermore, our results suggest ice streaming occurred in distinct stages along the trunk of the channel with high ice sliding in the inner parts before LGM and sliding in the outer parts of the channel close to the shelf break during LGM. Our results also show that sliding in the inner parts of the channel deactivated because of divergent ice flow when the BIIS and the SIS merged and formed a saddle across there North Sea.

6. Code/Data availability

Code and/or data will be made available upon request.

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7. Author contribution

Gustav Jungdal-Olesen: Conceptualization, Methodology, Software, Formal analysis, Writing, original draft, Visualization. Vivi K. Pedersen: Conceptualization, Methodology, Supervision, Writing, review & editing, Funding acquisition. Jane L. Andersen: Writing, review & editing, Visualization. Andreas Born: Resources, Writing, review & editing

8. Competing interests

The authors declare that they have no conflict of interest.

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