Modeling the Pleistocene History of the Greenland Ice Sheet

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MODELING THE PLEISTOCENE HISTORY OF THE GREENLAND ICE SHEET

A Dissertation Presented

by

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ABSTRACT

MODELING THE PLEISTOCENE HISTORY OF THE GREENLAND ICE SHEET

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One of the most profound and immediate consequences of anthropogenic climate change is sea level rise, which in large part is driven by the melting of polar ice sheets. The Greenland Ice Sheet (GrIS) contains enough ice to raise global sea level by ~7 meters. Fluctuations of the GrIS in response to past climate change provide an opportunity to better understanding the stability of the ice sheet during periods of climatic change. In this thesis, we use numerical ice-sheet models to understand the causes and consequences of past fluctuations of the Greenland ice sheet.

In Chapters 3 and 4, we examined the last deglaciation (21,000 years ago until present day). The last deglaciation is the most recent time when the ice sheet retreated significantly, shrinking from its advanced state during the Last Glacial Maximum to a minimum configuration slightly smaller than the present-day ice sheet. We evaluate simulations of the deglaciation against a database of observations in order to understand the causes and drivers of ice-sheet retreat around the margin of the ice sheet for different climate scenarios. In Chapter 3, we show how abrupt climate change and changes in seasonality affected different regions of the ice sheet by modulating the timing of
deglaciation around the margin. In Chapter 4, we analyze the mass balance processes that drove retreat in each region in order to identify the most salient drivers of retreat for each major drainage of the GrIS. Chapters 3 and 4 are in preparation for publication.

In Chapter 5, we studied the initiation of the ice sheet during the warm Pliocene. We show that early fluctuations of the ice sheet can lead to the development of a large proglacial lake, which has major implications for landscape evolution, abrupt climate change, and ice-sheet stability during the Pliocene and Pleistocene. These proglacial lakes have the potential to affect erosional processes, and we argue that they may be responsible for carving Petermann canyon, a geomorphological feature which has important implications for future ice-sheet stability. Chapter 5 has been submitted to *Geology* for publication.

In Chapter 6, we used different proxy records from the high Arctic to examine GrIS stability over the last 800,000 years. We show that proxy records from Lake El’gygytgyn (Arctic Siberia) and IODP Site 982 (North Atlantic Ocean) lead to divergent ice sheet histories which are nevertheless consistent with recent sea-level targets and the observation of cosmogenic nuclides in the bedrock below central Greenland. This study demonstrates the potential for numerical ice-sheet models to be used to assist with the identification of locales where additional data constraints could have the largest impact on our understanding of the history of the ice sheet. Chapter 6 is in preparation for publication.
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CHAPTER 1

INTRODUCTION

1.1 Introduction

Sea level rise (SLR) is one of the most profound economic, social and environmental issues facing humanity today. The flooding associated with sea level rise is alone projected to cost up to 3% of global GDP annually by the end of this century if emissions continue unabated (Jevrejeva et al. 2018). In the United States, sea level rise will disproportionately impact communities of color and those in low-income areas, further exacerbating the issues of environmental justice which plague the country today (Hardy et al. 2017). Globally, the displacement of hundreds of millions of people will have cascading social, political, and environmental impacts as those in low-lying areas, especially in the global south, are forced inland by rising seas (Geisler & Currens, 2017).

The rate of global SLR has nearly tripled since 1890 and has continued to accelerate over the satellite era (Hay et al. 2015, Nerem et al. 2018). The relatively modest rates of SLR in the 19th and most of the 20th centuries were driven primarily by increased oceanic heat uptake due to anthropogenic warming (IPCC 2013, Hay et al. 2015). In the late 20th century, SLR accelerated as glaciers globally began to retreat. Though many low-latitude glaciers have all but disappeared, and oceanic heat uptake is less efficient as the surface ocean warms and the Atlantic Meridional Overturning Circulation slows, SLR continues to accelerate (Nerem et al. 2018). This discrepancy is driven by the rather delayed response of polar ice sheets in Greenland and Antarctica, which have only recently started to lose mass and contribute to changes in global sea
level (Mouginot et al. 2019, Rignot et al. 2018). These massive ice sheets, which if completely melted would raise global eustatic sea level by approximately 60 meters (Morlighem et al. 2017, Bamber et al. 2013), are now losing mass at an accelerating rate (Mouginot et al. 2019, Rignot et al. 2018), in step with a corresponding acceleration in the rate of SLR. Yet the response of polar ice sheets to future warmth remains deeply uncertain (IPCC 2013) and incorporating their response into consensus projections of sea level change by 2100 remains contentious, despite their emerging role as the most important drivers of sea level change in the next decades, centuries, and millennia.

The Greenland Ice Sheet (GrIS) contains enough ice to raise global sea level by 7.36 meters (Morlighem et al. 2017), and recent work demonstrates that some Greenland outlet glaciers are rapidly retreating, with the potential to mobilize large volumes of interior ice in the coming centuries. The GrIS has experienced periods of mass loss during the last century, primarily in response to rising air temperatures, but was more or less in balance until the later part of the 20th century (Bjørk et al. 2012, Kjeldsen et al. 2015). Mass loss from the GrIS has increased by a factor of six since the 1980s, and the ice sheet is now driving global sea level to rise by 0.85 mm yr\(^{-1}\) (Mouginot et al. 2019), far outpacing the current contribution of the Antarctic ice sheet (0.75 mm yr\(^{-1}\), Rignot et al. 2019). However, projections of how the GrIS will respond to future warmth are divergent and depend critically on pathway that atmospheric carbon dioxide (CO\(_2\)), and accordingly earth’s average temperature, follow over the coming centuries (Aschwanden et al. 2019).

Understanding how GrIS volume has varied in the geologic past in response changes in climate can inform efforts to predict more accurately how the ice sheet will
change in the future, and thus refine projections of the future contribution of the GrIS to global sea level. Such efforts benefit from a range of geologic archives (ice cores, sediment cores, etc.) which help to reconstruct both the past forcing (i.e. magnitude of the change in climate) and the response to said forcing (i.e. magnitude of ice loss). These data can then be used as targets for ice sheet models, which are a critical tool for predicting future sea level as they incorporate a range of physical processes which determine the integrated response of the ice sheet.

1.2 The History of the Greenland Ice Sheet

1.2.1 The Holocene

People have lived along the margins of the Greenland ice sheet for at least 5,000 years (Raghavan et al. 2014). The earliest evidence of population by the Saqqaq people around 5,000 years ago in Western Greenland indicates they were skilled hunters and foragers, able to harvest resources as large as bowhead whales (*Balaena mysticetus*) (Seersholm et al. 2016). They lived in permanent settlements and utilized seasonal camps for subsistence (McGhee 1996). A genetically similar population, Independence I, settled in Northeastern Greenland contemporaneously (McGhee 1996). These early Greenlanders were remarkably adaptable, weathering the harsh climate of the high latitude Arctic and persisting through millennia of climatic and environmental change, including changes in relative sea level and adjustment of the ice sheet margin (McGhee 1996, Jensen 2005). Their presence in Greenland overlapped with a later population of Inuit (Thule) that migrated from North America around the 12th century CE. Genetic mixing occurred between the groups but predates by millennia the earliest archaeological
evidence of coexistence and is thus assumed to have occurred before the Thule arrived in Greenland (Raghavan et al. 2014). The cause of the disappearance of the Saqqaq and Independence people ~700 years ago is still unknown (Raghavan et al. 2014). Today, descendants of the Thule live all around the margin of the ice sheet and continue the subsistence practices that have sustained humans in Greenland for thousands of years.

Beginning in ~985 CE, Greenland underwent five centuries of European occupation during the Late Holocene by Norse explorers. The Norse built two settlements, commonly termed the Western (Sermersooq municipality) and Eastern (Kujalleq municipality). The Norse cultivated livestock and grew imported crops to sustain themselves, though through time, they increasingly relied on marine resources for sustenance. Many explanations have been offered for their disappearance around ~1450 CE, including climate change, disease, and overexploitation of natural resources, but it is likely that no one factor fully explains the decline of the Norse. During the last century, Greenland has been a locus of northern hemisphere geopolitical posturing, with competing claims by the United States, Norway, and Denmark compromising the autonomy of the island. In 2013, Greenland obtained selvstyrelse, or self-governance, but remains closely politically and economically aligned with Denmark.

The rise, decline, and persistence of human populations in Greenland occurred against the background of ongoing climatic and environmental change well documented by studies of lake sediments, moraines, and erratic boulders. The largest changes to the Greenland ice sheet likely occurred as a (lagged) response to the Holocene thermal maximum (HTM), when surface temperatures rose up to 3°C warmer than present (Lecavalier et al. 2018, Buizert et al. 2018, Lasher et al. 2017). This warming was
accompanied by changes in precipitation that may have bolstered the mass balance of the ice sheet (Thomas et al. 2017); however, other studies have found that the ice sheet clearly responded to the HTM temperature perturbation, in places retreating by tens of kilometers from the present-day margin. In Southern Greenland, active regions of the ice margin retreated up to 10s of kms behind their present-day position in response to early Holocene warming, and remained outside the catchment of so-called “threshold lakes” until at least the Neoglacial (~4 ka). For some outlets, the glaciers did not advance to their present-day extent until the 300–500 years ago, indicating that for most of the Holocene the Southern lobe of the Greenland ice sheet was smaller than it is today (Larsen et al. 2016). To the west in Narsarsuaq, the ice sheet was behind its present margin from 6.9–3.0 ka and again from 2.8–0.5 ka (Larsen et al. 2011). Further north, parts of the Western GrIS were up to ~15 km behind their present margin from 5–3 ka, though recent studies near Kangerlussuaq suggest a stable ice front, within 900m of its current position for most of the Holocene (Young and Briner 2015). In contrast, the margin in the Northeast retreated by ~70km between 7.8–1.2 ka (Larsen et al. 2018). These differences in ice sheet evolution during the Holocene may reflect different dynamic processes modulating ice loss or differences in climate which exacerbated or outweighed changes in surface melt in response to Holocene warmth (e.g. Lasher et al. 2017, Thomas et al. 2017). Nevertheless, the radiostratigraphy of the ice sheet suggests that HTM ice velocities were up to 100% higher than present day throughout Greenland, perhaps facilitated by increased delivery of surface meltwater to the subglacial system (Macgregor et al. 2015). The HTM warming and melting of Greenland, which is especially pronounced in the North, may have also been related to other environmental
changes in the Arctic, like the disappearance of sea ice from Greenland’s northern margin (Funder et al. 2011). Understanding the spatial variation in ice sheet response to Holocene warmth is of critical importance to predicting the future response of the GrIS (Briner et al. 2017) and is a central component of this Dissertation.

1.2.2 The last deglaciation

During the LGM, the Greenland ice sheet expanded beyond its present-day margins and far out onto the continental shelf, accommodating an additional 2-6 m sea level equivalent (sle) of ice (Figure 1.1). In the Northwest, the ice sheet coalesced with the Innuitian dome of the Laurentide ice sheet, forming a continuous ice mass that spilled out into an enormous ice shelf in Baffin Bay (Figure 1.1). As the climate naturally warmed from the LGM into the Holocene, Greenland shrunk, reaching a minimum extent inland of its present-day margins sometime during the Holocene Thermal Maximum (Alley et al. 2010). Thus, the last deglaciation provides a test-case for understanding the dynamics of ice-sheet retreat during a period of substantial ice loss from Greenland.

Ice sheet models provide one way to learn more about the deglaciation. Ice sheet models have the advantage of painting a physically consistent and time-continuous portrait of the deglaciation. Simulations can be constrained to fit existing geological data (e.g. Huybrechts et al. 2002) or allowed to freely evolve in response to changing boundary conditions (see chapter 2).

Detailed knowledge of ice-sheet change can be extracted from geological records, but these are subject to a number of uncertainties. Indirect evidence of ice-sheet behavior is available from marine sediment cores (e.g. Larsen et al. 1994, Reyes et al. 2014). Ice-rafted debris (often defined as sediment in the > 2 mm size fraction) can provide a
qualitative indicator of ice conditions on a regional scale, although these particles are also subject to transport by sea-ice, so it is often necessary to establish the geochemical provenance of such debris to explicitly link these records to changes in the ice-sheet (e.g. Bailey et al., 2013). Submerged moraines indicate that the GrIS advanced to the continental shelf break in many regions during the LGM (Funder et al. 2010). Terrestrial moraine complexes can sometimes be assigned a minimum age with $^{14}$C and $^{10}$Be dating, and occasionally provide multiple snapshots of ice-margin position (Levy et al. 2012). Moraines can be difficult to date in the absence of organic material, and their meaning in relation to ice dynamics is often equivocal, as a moraine can be deposited during ice-sheet retreat or advance (Stokes et al. 2006). Crustal rebound rates around the perimeter of the ice-sheet provide some constraint on LGM thickness and retreat history (Peltier et al. 2016). However, these data cannot provide information at a high spatial resolution, and poorly constrained parameters, such as the viscosity and structure of the mantle beneath Greenland, have a large impact on reconstructed ice geometries (Lecavalier et al. 2014). Lake records ("threshold lakes") from the margin of the ice sheet can provide some constraints on the minimum extent of the ice sheet during parts of the deglaciation that were warmer than present (Björck et al. 2002, Young and Briner 2015, Larsen et al. 2015). Other lake records ("isolation basins") are used in conjunction with $^{14}$C dating to reconstruct relative sea level (RSL) around the ice-sheet, which can be linked to ice retreat and advance (Long et al. 2011). It has long been known the age and water-depth inferred from these deposits are subject to uncertainty (Long et al. 2011). However, recent advances in our understanding of ice-sheet/ocean interactions have added additional layers of complexity to the interpretation of RSL indicators in ice-proximal
areas (e.g. Gomez et al. 2010), complicating the use of such records as direct constraints on ice-sheet history (e.g. Lecavalier et al. 2014, Simpson et al. 2009, Huybrechts et al. 2002).

Recent methodological and technical advances have led to a proliferation of studies using cosmogenic isotopes (e.g. $^{26}$Al, $^{10}$Be) to directly date ice-margin retreat. These isotopes are generated in measurable concentrations within the uppermost decimeters of bedrock and glacial erratics through spallation by cosmic rays. As areas become ice-free, surfaces previously subject to erosion beneath the ice-sheet begin to accumulate cosmogenic isotopes at a known rate. Thus, the concentration of these isotopes in surface samples (i.e. boulders and moraines) can be used to precisely infer the date that the surface was exposed (e.g. Balco et al. 2008).

Thanks to a suite of recent studies, surface exposure dates are now available around the entire perimeter of the Greenland ice sheet (Sinclair et al. 2016, and references therein). Importantly, these reconstructions are not subject to the same uncertainties as many of the proxies described above and are emerging as the most reliable source of information about ice-margin retreat. A recent analysis of all published surface exposure dates found that GrIS retreat was spatially asynchronous (Sinclair et al. 2016). They found that the east, south and west regions of Greenland became ice-free at 12–11 ka, 11-10 ka, and 10-7.5 ka, respectively (Sinclair et al. 2016). This observation builds on previous work describing asynchronous deglaciation for adjacent outlets in West and East Greenland (Ó Cofaigh et al. 2012, Dyke et al. 2014). Yet existing models of the deglaciation are unable to capture this behavior. In addition, Young and Briner (2015) noted discrepancies in the reconstructed minimum ice sheet configuration between
available ice sheet models and threshold lake data. As a result, there is a growing demand within the community for reconstructions of past-ice sheet behavior which incorporate (1) a realistic climatic forcing, (2) explicitly consideration of ice-sheet/ocean interactions, (3) increased temporal and spatial resolution, and (4) dynamic interactions between the Greenland and Laurentide ice sheets. Better representation of these aspects in a three-dimensional ice sheet model is necessary to better understand what drove the contrasting ice-sheet responses for different sectors during the deglaciation.

1.2.3 The late Pleistocene and Plio-Pleistocene

Recent results, building on a rich history of observations, have undermined the long-held assumption that the Greenland ice sheet has been relatively stable over the Pleistocene. Schaefer et al. (2016) analyzed the cosmogenic isotopes $^{10}$Be and $^{26}$Al bedrock below the modern-day ice-sheet summit (the GISP2 site; Figure 1.1) and found that the concentrations of these radioactive nuclides are too high for this site to have been continuously ice-covered during the entire Pleistocene. Their analysis suggests that the ice sheet has been absent for at least 300,000 of the last 1.4 million years, but their results do not constrain the exact number of exposures at that site. In addition, the rate of change of the nuclides with depth implies that no more than 2 meters of bedrock erosion has occurred during the same period, implying that the site has seldom seen temperate ice.

Two plausible hypotheses have been put forward to explain the observation of high cosmogenic isotope concentrations beneath the present-day summit. The first requires that the site underwent a long period of continuous exposure (~300,000 years) before ice-sheet expansion at ~1 Ma resulted in an ice sheet whose interior has been
stable since that time. The second implies that, following an initial glaciation around 1 Ma, the incipient ice sheet melted away during many, but not all, subsequent interglacials (i.e. MIS 47, 31) allowing the GISP2 site to remain ice-free for about 20,000 years during some glacial-interglacial cycles. (A third hypothesis, that the Greenland ice sheet has melted completely away during every interglacial period of the Pleistocene, is unlikely considering the widespread presence of Eemian ice below the modern ice sheet and the much older ages obtained for genetic material and gases from deep ice cores (Willerslev et al. 2007, Yu et al. 2016, Dahl-Jensen, personal communication).) Both hypotheses could be consistent with other findings, but they also present new paradoxes that must be examined if we are to improve our understanding of the ice sheet’s stability in a warming climate.

The first hypothesis requires that the GISP2 site was ice-free for a period of 300,000 years from ~1.4 – ~1.1 Ma, although the duration is slightly more robust than the exact timing (Schaefer et al. 2016). This implies that the site remained warm enough to inhibit ice-sheet growth for 7-8 glacial cycles known from marine records (Lisiecki and Raymo 2005). Proximal marine records (e.g. Larsen et al 1994, Bierman et al. 2016, and many others) indicate that ice-rafted debris and eroded material sourced from Greenland was continuously deposited over this time period, suggesting that some ice was present on Greenland despite the interior being ice-free. These observations require a mechanism for inhibiting continental-scale ice-sheet growth on Greenland under glacial climate conditions during the Early Pleistocene. Previous work on glacial inception showed that, as the ice-sheet expanded, a feedback between atmospheric moisture transport and the growing ice-sheet resulted in a Föhn effect that limited ice sheet growth (Langen et al.,
2012). That study looked at ice growth under modern climate conditions and suggested that glacial climate conditions would be necessary for ice-sheet accumulation to overcome such a feedback. However, they did not examine thresholds for this behavior, nor whether the Föhn feedback would be sufficient to inhibit ice growth over multiple glacial cycles during the Early Pleistocene.

The second hypothesis requires that the ice sheet retreated during some interglacial periods in the Early Pleistocene but remained stable during the Late Pleistocene. This implies that the ice sheet’s sensitivity to climate changes has changed over time. Other work has suggested that the Arctic surface temperatures have varied between interglacial periods, which could drive different ice-sheet responses. Melles et al. (2012) analyzed sediments from Lake El’gygytgyn (Northeast Arctic Russia) and identified a number of “super-interglacial” periods, when the Arctic seems to have been exceptionally warm. Based on the results of that study, these super-interglacial periods have characterized Arctic climate for 10% of the last ~1.6 Ma (Melles et al. 2012). Further work on the same sediments has revealed that at least some of these “super-interglacials” were characterized not only by exceptional warmth, but by their long duration. Based on a temperature proxy reconstruction bracketing MIS 31, de Wet et al. (2016) suggested that MIS 33 and 31 were separated by a cold period only a few thousand years in duration in the Arctic, resulting in an interglacial climate for ~50 ka. In addition, a modelling study by Solgaard et al. (2011) found that the ice sheet probably retreated during interglacial periods in the Late Pliocene and into the early Pleistocene, consistent with the observation of interbedded glacial tills and interglacial deposits at Kap København, North Greenland (Funder et al. 2001).
This second hypothesis is somewhat more problematic to reconcile with other evidence (i.e. soils dated at 2.7 Ma suspended in the basal ice from the same core), especially given the very low rates of erosion (<2 m Ma\(^{-1}\)) the Schaefer et al. (2016) analysis requires. (The more that an ice sheet is required to advance and retreat across a site, the more likely it is to be warm-based and highly erosive there – for comparison, the rates of erosion beneath the Laurentide ice sheet during the Pleistocene were on the order of 40 m Ma\(^{-1}\) (Balco et al. 2004).) However, this hypothesis also highlights the potential role of other boundary conditions in shaping the long-term behavior of the ice sheet, which many other studies have argued for. For example, Koenig et al. (2014) found that reduction of sea-ice extent and imposition of Pliocene-like sea-surface temperatures resulted in rapid reduction of ice-sheet extent. Many studies have found that changes in sea-ice extent result in dramatic changes in Arctic surface temperatures (i.e. Ballantyne et al. 2014) and, while some have suggested that modern sea-ice configurations have persisted over the last 2.5 Ma, others have argued that differences in sea-ice cover could explain the diverse climatic responses of the Arctic to glacial-interglacial cycles in the early Pleistocene (e.g. Keisling et al. 2016).

At present, few long-term simulations of the GrIS show retreat patterns consistent with either of the hypotheses put forward by Schaefer et al. (2016). Some work has suggested that during past warm interglacial periods of the Pleistocene the GrIS has experienced very limited mass loss. For example, during the Eemian, which was +8°C warmer than present over central Greenland (NEEM Community Members 2013), one study saw little change in Greenland ice volume (Helsen et al. 2013). Thus, there remains
a gap in our understanding of the processes that affect ice sheet behavior during climatic periods similar to the present day.
Figure 1.1 Extent of the Greenland Ice Sheet at present day (white, with 1 km thickness contours in black) and modeled for the LGM (blue). Gray colors indicate bathymetry contours; note that during the LGM, the GrIS advanced onto the continental shelf in many places (Funder et al. 2001). Locations mentioned in the text are annotated with black dots and text boxes. Major ice core sites, from north to south, are Agassiz, Camp Century (CC), NEEM, EastGRIP, GISP2, GRIP, Renland, and Dye3. JI – Jakobshavn Isbræ. ZI – Zakariæ Isstrøm.
CHAPTER 2

METHODS

2.1 Ice Sheet Modelling

Three-dimensional numerical ice sheet models can be used to examine the plausibility of various paleo-ice-sheet retreat scenarios, which can reveal, for example, the sensitivity of different regions of the ice sheet to various forcings and help refine predictions of future mass loss (e.g. DeConto and Pollard 2016). Due to positive feedbacks between ice sheet growth and snow accumulation, it has been difficult for ice sheet models to reproduce the magnitude of retreat inferred from geological records under known climate forcing scenarios. Recent improvements in ice sheet modelling (Pollard et al. 2015) and their application to Antarctica (DeConto and Pollard, 2016) have bridged this gap. Antarctic simulations now capture the dynamic behavior and severe ice loss during Pleistocene interglacials long inferred from geologic records but seldom captured by model simulations of the ice sheet. Despite similar issues in the Northern Hemisphere outlined above, this modelling approach has yet to be applied to the GrIS.

The Penn State Ice Sheet Model is a three-dimensional thermomechanical ice sheet model which uses a combination of the Shallow Ice Approximation (SIA) and the Shallow Shelf Approximation (SSA) to simplify the physics of ice flow. Ice behaves as a non-Newtonian viscoelastic fluid, with a nonlinear proportionality between strain and stress summarized by a temperature-dependent constitutive equation:

\[ \dot{\epsilon} = A(T)\tau^n \]  

(1)
Standard practice based on field observations and laboratory experiments constrains the value of $n$ to be about 3 (Patterson 1961). This results in a complex and computationally expensive system of equations, and approximations have been proposed to make spatially and temporally extensive simulations of ice flow feasible. One approximation assumes the lateral expanse of the ice is large compared to its thickness, resulting in the driving stress being balanced solely by the basal shear stress (SIA). Another approximation, originally developed for floating ice shelves (where basal shear stresses are necessarily zero), assumes that longitudinal stretching dominates ice flow (Patterson 1961). We use a hybrid approach that solves both the SIA and SSA equations for every grid cell, and subsequently scales the contributions of each flow law based on the magnitude of the basal sliding coefficient ($C$ in Equation 2). This approach achieves computational efficiency while maintaining many important aspects of ice flow, including the effects of three-dimensional stress fields on ice viscosity and the propensity of some parts of the grounded ice sheet to slide at the ice-bedrock interface, thus behaving more like an ice shelf. (i.e. ice streams). The latter process, termed basal sliding, is captured via a basal sliding law:

$$u_b = C|\tau_b|^{m-1}\tau_b$$  \hspace{1cm} (2)

Here, the value of $m$ corresponds to the ability of the bed below the ice sheet (in many places, a column of deformable till) to transmit stresses upstream, $C$ corresponds to a spatially variable basal sliding parameter which we deduce from the present-day ice-
sheet geometry using inverse methods (Pollard and DeConto 2012), and $\tau_b$ is the basal stress.

The assumptions described above do a reasonable job of capturing the large-scale flow features for most of the GrIS. However, they break down near the grounding line, where the ice transitions from resting on land to floating in the ocean, and where all stresses are important for capturing ice dynamics. To get around this, Schoof (2007) developed a parameterization based on an analytical solution to the coupled SIA-SSA system which describes the flux of ice across the grounding line:

$$q_g = \left( \frac{\overline{\alpha}(\rho_0)^{n+1}(1-\rho_i/\rho_0)^n}{4^n C_s} \right)^{\frac{1}{m_s+1}} \left( \frac{t_{xx}}{\tau_f} \right)^{\frac{n}{m_s+1}} \left( \frac{m_s+n+3}{m_s+1} \right) h_g$$  \hspace{1cm} (3)

This parameterization allows the model to capture grounding-line migration, which is known to be an important part of ice advance and retreat (Schoof 2007, Pollard and DeConto 2012). The full details of the parameters above can be found in Pollard and DeConto (2012), but the most notable feature is the farthest right term, which dictates that the discharge across the grounding line scales as the ice thickness to the power of $\sim 5$ (Schoof 2007). Although the hybrid model we employ is in some ways a crude representation of reality, it captures the major features of ice flow in Greenland (e.g. Figure 3.1a), compares favorably to more complex models in intercomparison studies (e.g. Pattyn et al. 2012), and permits simulation of Greenland ice sheet history over the timescales relevant to interpreting geological records of ice-sheet and sea-level change.
2.2 Climate Modelling

The most important factors driving ice sheet behavior on the timescales considered here are climatic. Changes in temperature and precipitation directly impact the ice sheet’s mass balance, which in turn effect the dynamics of the ice sheet. In this work, we use a range of strategies to develop the climatic forcing used to force the ice sheet model: a proxy-driven climatology, regional climate modelling, and an index method which merges the two.

For the last 21,000 years, we developed a climate reconstruction based on a combination of the modern temperature and precipitation fields over Greenland (Box et al. 2013a,b) and δ\textsuperscript{15}N records from three deep ice core sites (Buizert et al. 2014). The modern climatology (temperature and precipitation) was created using shallow ice core records from across Greenland and climate reanalysis data (RACMO 2.2) (Box et al. 2013a,b). The modern climatology was scaled back in time using δ\textsuperscript{15}N measurements on trapped air from the NEEM, NGRIP, and GISP2 deep ice core sites (Buizert et al. 2014). This generates a climatology that varies in time as well as in space, because the magnitude of temperature change varies at the three ice core sites based on the forcing mechanism (Buizert et al. 2014). Finally, to understand the effects of abrupt climate change and seasonality on GrIS evolution over the last 21,000 years, we created two additional reconstructions: one where abrupt climate changes are removed and replaced with smoothed temperature changes, and one where the seasonality of temperature change is adjusted to match the TRACE-21k simulations (Liu et al. 2009).

We compare these reconstructions with climatologies generated by the RCM RegCM3 for every millennium over the last 21,000 years (Pal et al. 2007). The RCM has
been tuned to better capture polar climate by validating model output for modern conditions against RACMO 2.3 (Noël et al. 2015, Figure 3.3a,b). This tuned version of the RCM can then be run using a variety of boundary conditions, including earth’s orbital configuration (Berger et al. 2004), atmospheric greenhouse gas concentrations (Bereiter et al. 2015, Schilt et al. 2010), and ice sheet configuration (Argus et al. 2014, Peltier et al. 2015). To create the RCM boundary conditions, we first run an atmosphere-only (slab ocean) GCM to equilibrium using the appropriate boundary conditions (Genesis; Thompson & Pollard 1994a,b). Once equilibrium is reached (~50 model years), the GCM is run for 10 additional model years, during which RCM boundary conditions are saved every 6 hours. This 6-hourly output is then used to force the RCM for ten years, which is averaged to produce a mean annual climatology.

For the final chapter of this thesis, we used an index method which combines the proxy-based approach and the modeling-based approach (e.g. DeConto and Pollard 2009). We selected high-resolution proxy records which capture climatic variability in the Arctic over the last 800,000 years. The proxy records were normalized to the change between LGM and modern day to create an index. Then we assigned RCM output files to different index values (e.g. LGM climate for the LGM index value, modern climate for the modern index value), and a long-term climatology was created by linearly interpolating between the RCM climatologies based on the proxy records. We use an ensemble approach with this technique in order to force the ice sheet model with a large number of climatologies and select models to analyze based on their ability to meet pre-established data targets.
2.3 Model-data integration

A central strength of running ice sheet models for the past is that their performance against established datasets can be used to establish their accuracy and thus build confidence in their predictions of future ice sheet change. Model-data comparison is complex because it requires close examination of the shortcomings, biases, and multiple interpretations of both model output and proxy data.

In this thesis, the central data that we use for comparison to ice-sheet model output are in situ cosmogenic isotope records from around and beneath the Greenland ice sheet. Cosmic rays, forms of high-energy radiation, constantly bombard the earth but attenuate rapidly within most media including rock, soil, water and ice (Balco et al. 2008). Radioactive cosmogenic nuclides are produced within minerals when the minerals are exposed to cosmic rays. For example, bombardment of the mineral quartz (SiO$_2$) by cosmic rays produces both the cosmogenic nuclides $^{26}$Al (from $^{28}$Si) and $^{10}$Be (from O$_2$). These nuclides accumulate at a known rate and have a known half-life (Balco et al. 2008), so their concentrations reflect the total amount of time a sample has been exposed to the atmosphere, provided the sample has not been covered by much ice or soil or experienced significant erosion.

Around much of the GrIS margin, cosmogenic nuclide concentrations can be used to unambiguously determine the time that ice sheet last occupied the area provided that the samples are devoid of inherited nuclides from a past exposure (e.g. Sinclair et al. 2016). This requires either sampling erratic boulders which are unlikely to have been left in the same orientation during a previous period of exposure, or that surfaces sampled (bedrock or boulders) have been glacially eroded in the uppermost ~meter, so that any
nuclides inherited in the past have been removed. For most of the periphery of the Greenland ice sheet, careful sampling ensures this criterion is met; in some regions particularly in Northern Greenland, the exposure age and the amount of inheritance can be simultaneously solved for by sampling boulders and bedrock in close geographic proximity (Corbett et al. 2013). Cosmogenic nuclides used in this way provide insight into the timing of deglaciation around the ice sheet margin unburdened by uncertainties that have plagued other proxy records of the timing of deglaciation, particularly relative sea level records, which are complicated both by the ambiguity of radiocarbon dates during the deglaciation and heterogeneity in solid earth structure (e.g. Lecavalier et al. 2014).

On longer timescales, cosmogenic nuclides have revolutionized our understanding of the stability of the GrIS. The history of the Greenland ice sheet is a topic of intense debate. The lack of a clear Antarctic signature of ice-sheet collapse during the most recent interglacial period, the Eemian (~125 ka), would require near-total collapse of the Greenland ice sheet to meet the global sea level budget during the same period (Holloway et al. 2016, Sime et al. 2019). Direct records of Eemian ice in Greenland suggest that this was not the case but cannot constrain the history of the GrIS beyond ~130 ka, presumably because older ice from has been lost to basal melting (NEEM Members 2013, MacGregor et al. 2015). In contrast, analysis of the suspended sediment from the lowermost meters of the GISP2 deep ice core site suggested that the ice sheet had been stable and non-erosive there for 2.7 Ma (Bierman et al. 2014). This longstanding debate was reinvigorated with the observation that the concentrations of in situ $^{10}$Be in the bedrock core recovered from beneath GISP2 were much higher than would be expected were the landscape to have
remained ice-covered for the last few million years (Schaefer et al. 2016). Instead, these new data revealed that central Greenland had either undergone a long, relatively recent period of ice-free conditions (300 kyr duration no earlier than 1.1 Ma) or frequent deglaciation during the Pleistocene. This observation was complemented by another insightful application of cosmogenic isotopes, measurement of \(^{10}\)Be and \(^{26}\)Al in marine sediments off the margin of the Greenland ice sheet (Bierman et al. 2016). These records, though temporally coarse, revealed continuous yet dynamic delivery of once exposed glaciogenic sediments to marine outlets since at least 7 Ma.

These cosmogenic nuclide data, in both their straightforward exposure-age and more complex multiple-exposure applications, provide novel and useful targets for paleo ice sheet models which can constrain the processes and mechanisms which led to the destabilization of the GrIS in response to past climatic change. They provide a central set of modelling targets used throughout this thesis.
CHAPTER 3

ROLE OF SEASONALITY AND ABRUPT CLIMATE CHANGE IN DRIVING GREENLAND ICE LOSS DURING THE LAST DEGLACIATION

3.1 Abstract

During the last deglaciation, the Greenland ice sheet (GIS) reduced in volume by 25–100%. The orbital changes were punctuated by a series of abrupt climate events (i.e. the Bølling-Allerød and Younger Dryas), but these climate events did not leave a clear imprint in ice-marginal terrestrial landforms and are more muted in ocean records of climate change. Thus, it is unclear whether abrupt climate change was a salient feature of deglacial climate at Greenland’s margins, and the role those changes played in the retreat of the ice sheet. Here we seek to identify the impact of abrupt climate change on the deglaciation of the GIS by applying a range of forcing scenarios to a numerical ice sheet model and evaluating the timing and progression of ice retreat relative to published $^{10}$Be and $^{14}$C ice-margin data. To isolate the impact of abrupt climate changes relative to other factors, we forced the ice-sheet model with an ice-core-based climate reconstruction (including abrupt climate changes) and an asynchronously coupled regional climate model (forced by changing greenhouse gas concentrations and evolving orbital configurations). We find that, for most regions, including abrupt climate change is critical in trying to fit to the ice-margin data, except in west Greenland, where local surface ocean temperature changes likely exerted a primary control on ice-sheet behavior. Our results provide information about the regionally heterogeneous sensitivity of the
Greenland ice sheet to dynamic and climatic factors which are often ignored in simulations of the last deglaciation. Critically, we show that the abrupt climate changes recorded in the ice-sheet interior were key drivers of ice-margin retreat for most sectors of the ice sheet during the last deglaciation.

3.2 Introduction

Understanding the role of climate change in driving the last deglaciation of the Greenland Ice Sheet (GIS) is important for predicting future mass loss. During the transition between the Last Glacial Maximum (LGM) and the Holocene, referred to here as the last deglaciation, the GIS retreated from an advanced grounded position on the continental shelf and shed 25–100% of its mass (i.e. Funder et al. 2011, Lecavalier et al. 2014). During the deglaciation, changes in CO$_2$, insolation, and the disintegration of the large Northern Hemisphere ice sheets warmed Greenland’s climate, nudging the ice sheet towards a new equilibrium. Superimposed on these warming trends were rapid climatic adjustments, termed “abrupt climate changes,” likely in response to changes in the Atlantic Meridional Overturning Circulation (AMOC) (Alley et al. 1997).

Although abrupt climate changes are documented by ice cores in the ice sheet interior, the impact of these events on the Greenland margin is not clear, and discrepancies between reconstructed marginal ice behavior and ice-core climate records remain unresolved. Reconstructions from ice-marginal lakes indicate that summer temperatures during the Holocene Thermal Maximum (HTM) were 3–5°C above preindustrial (Frechette et al. 2009, Axford et al. 2009, Axford et al. 2011, Axford et al. 2013, Lasher et al. 2017), a result corroborated by temperature reconstructions from the
Agassiz ice cap (Lecavalier et al. 2017) and supported by the predominance of melt layers in HTM ice from that site (Lecavalier et al. 2017). These elevated temperatures are not seen in records of temperature from the interior of the ice sheet. The lack of a strong HTM signal in interior Greenland has been proposed as a reason for poor model-data agreement in some regions (Lecavalier et al. 2014), and questions the general validity of extrapolating climate records from Greenland’s interior to its margins. This is especially problematic because past modelling studies have used $\delta^{18}$O-based temperature reconstructions to force ice-sheet change, though these reconstructions are known to be strongly affected by factors other than temperature, such as moisture source origin and pathways (Langen and Vinther 2009, Charles 1994) and precipitation seasonality (Werner 2000). Despite this uncertainty, the $\delta^{18}$O data indicate that the abrupt events during the deglaciation changed mean annual temperatures by ~8ºC over a period of <50 years (Alley 2000; Steffensen et al. 2008).

Abrupt climate change has not been identified as a major control on the position of the ice margin around most of the ice sheet. Due to the expanded ice sheet geometry during the LGM, few marginal records of climate change extend beyond the HTM. However, the records that do exist suggest that other pronounced and abrupt climatic changes were muted at the margins of the ice sheet, at least seasonally.[references needed]. Although temperature reconstructions from ice cores fingerprint the Younger Dryas signal as being strongest in the south, no evidence for a coherent response of the ice sheet to this climatic perturbation has yet been documented in terrestrial records Southern Greenland. For example, Björck et al. (2002) found a limited expression of the Younger Dryas (YD) in southern Greenland based on a summer temperature
reconstruction. In contrast, other studies found that the ice margin stabilized, or kept retreating, during this prolonged cooling event (Hall et al. 2008, 2010, Jennings et al. 2006, Kelly et al. 2008, Kuijpers et al. 2003). Ice margin advance during the Younger Dryas is observed in Northern Greenland (Larsen et al. 2010) and parts of Western Greenland (Disko Bugt; Ó Cofaigh et al. 2013), indicating that the YD may have had an impact on some sectors of the ice sheet. However, the reasons that some regions were more sensitive to this abrupt climate change than others remain unclear.

The proliferation of studies using cosmogenic isotopes (especially $^{10}$Be and $^{26}$Al) to date the age of landscape exposure provides a fresh source of information about the timing and dynamics of deglaciation around Greenland (Sinclair et al. 2016 and references therein). These exposure ages can be used as targets for ice-sheet model simulations to test the role of abrupt changes in the deglaciation (Sinclair et al. 2016). Such models can provide a physically consistent picture of how the whole ice sheet changed in response to the changing climate. Yet existing simulations of the deglaciation have not been evaluated against these exposure dates and have other shortcomings that limit our understanding of the temporal and spatial effects of abrupt climate change on the ice sheet. Here we provide Greenland deglacial ice-sheet model simulations which are higher resolution than previous studies, consider additional aspects of ice dynamics, incorporate an ocean forcing parameterization, and utilize more realistic climate forcing scenarios that are not subject to the same uncertainties as $\delta^{18}$O paleothermometry (Buizert et al. 2018). We present simulations of GIS retreat through the deglaciation, at higher spatial resolution than previous studies, and using a range of climate forcing scenarios. We show that the inclusion of abrupt climate change in the forcing results in
ice sheet simulations in better agreement with exposure data in most regions, and map the spatial sensitivity of the ice sheet to the inclusion of abrupt climate change to better understand how it may respond to such climate forcing in the future.

3.3 Methods

We use the Penn State/UMass Ice Sheet Model (Pollard and DeConto, 2012) to model the GIS during the last deglaciation. The model is a 3D hybrid shallow ice/shallow shelf thermomechanical ice sheet model, with Weertman-type basal sliding and ice-flux across the grounding line parameterized after Schoof et al. (2007), to permit realistic grounding line migration. For all of the simulations, the LGM ice sheet was initialized by spinning up the model with LGM climate forcing for 40,000 years, yielding a reasonable match to the present-day temperature field of the ice sheet measured at borehole sites. We run the model at 10 km² resolution. The ice sheet model incorporates calving of floating ice as a function of the large-scale stress field, and both submarine and ice-front melt rates are calculated via an ocean melt parameterization similar to that in Martin et al. (2011). As the ice sheet evolves, a lapse rate correction is applied to the temperature and precipitation rates in each grid box to reflect growth or decay of the ice sheet relative to the ice sheet geometry associated with the climate forcing. The elevation of the bed adjusts to the changing ice sheet via an Elastic Lithosphere Relaxing Asthenosphere (ELRA) model with a single time constant (3000 years; Pollard & DeConto 2012).

To study the impact of climate forcing on the deglaciation of the GIS, we used three different climate forcing scenarios. In the first scenario, we apply a deglacial
climate forcing that has been described in detail by Buizert et al. (2018). Briefly, the reconstruction employs a spatial pattern based on modern temperature and precipitation (Box et al. 2013a,b), but scaled in time using a nitrogen isotope reconstruction of temperature at Greenland ice core sites for the last 21 kyr. Seasonality is adjusted following the Trace21k (Liu et al. 2009) transient simulations using CCSM3. The inclusion of variable seasonality has a substantial effect on the mass balance of the ice sheet (Buizert et al. 2017); here we also explicitly examine the role of seasonality in driving the deglacial retreat of the ice margins. We refer to this scenario as “VAR” (Buizert et al. 2017). To consider the role that time-evolving seasonality plays, we use the same climate forcing but with seasonality fixed to modern; we refer to this scenario as “FIX.” These climate forcings capture the known climatic evolution over the summit of the ice sheet, but they do not allow for changes in the spatial pattern of climate related to, i.e. changes in large-scale atmospheric circulation, sea ice, or sea surface temperatures (SSTs). To address this, we performed another simulation with a Regional Climate Model (REGCM3) run at 40km resolution and asynchronously coupled the ice sheet model every 1 ka through the deglaciation. In this simulation, global greenhouse gas concentrations (Monnin et al. 2001), orbital parameters (Laskar et al. 2004), Northern Hemisphere ice cover (Peltier et al. 2015), and sea surface conditions (sea ice/SSTs) (Liu et al. 2009) were updated every 1ka and recoupled to the evolving Greenland ice sheet (Figure 3.3). We refer to this scenario as “RCM.” Finally, to isolate the impact of ice dynamics on the deglaciation of the GIS, we ran an additional simulation with the FIX climate forcing for an expanded domain covering both Greenland and the easternmost area of the Canadian Arctic. The only difference between this simulation and the FIX
simulation is the expansion of the westernmost boundary from 75ºW Longitude (standard domain) to 96ºW Longitude (expanded domain). Although the resolution of the model is still too coarse to resolve some ice-dynamical processes, i.e. fast flow or ice-ocean interaction within narrow fjords, a major feature of the LGM GIS that is captured here is its buttressing against the Innuitian Ice Sheet across the Nares straight and a Baffin Bay ice shelf. The physical effects of this buttressing on ice flow are captured by the model at this resolution. We refer to this scenario that includes expanded Innuitian ice as “INN.”

In all simulations, ocean temperatures were reduced by 2ºC during the LGM, and evolved in time following a Labrador Sea subsurface temperature reconstruction (Winsor et al. 2012). This reconstruction was chosen as it represents one of the few available subsurface (100-200m water depth) temperature reconstructions in the region. Temperatures at this depth reflect the influence of warm Irminger Current waters, which exert a primary influence on submarine melt rates in modern-day fjords around Greenland (Straneo et al. 2010). In addition, all simulations were conducted using a mass-conserving bedrock map (Morlighem et al. 2014).

To assess the model-data fit for each of the simulations, we extracted ice thickness histories for all grid cells where 10Be data are available. For grid cells where more than one 10Be date exist, the ages were averaged to provide a mean target for the model. The results were then given a binary score dependent on whether or not the model deglaciated at the time indicated by the 10Be exposure age data (defined as within one or two standard deviations of the mean). Many 10Be sampling studies provide information about deglacial processes at a higher spatial resolution than we can use in this study; future efforts to downscale model output to finer, fjord-resolving grids will be necessary.
to fully take advantage of the available data. Nonetheless, our approach allows us to identify regional differences and to highlight the relationship between the chosen boundary conditions, climate forcing, and the resulting deglacial histories.

3.4 Results

In this section, we first describe the differences between the LGM configuration of the ice sheet using the different climate forcing scenarios and the two domains described above. We then describe the transient response of the ice sheet and evaluate, on a regional scale, the performance of the model against $^{10}$Be ages for each simulation.

During the LGM, reduced oceanic and atmospheric temperatures allow the ice sheet to advance far beyond its present grounding line (Figure 3.2b). In Northwest Greenland, the ice sheet grounds in Nares Strait and extends to the edge of the standard domain. With the extended domain, the ice sheet grows thicker in the Nares Strait and the enhanced buttressing leads to a thicker ice sheet throughout Northern Greenland. All across the northern margin, the ice sheet extends past the present-day coastline and grounds on the continental shelf at depths between 100-200m. The only exception is in the Lincoln Sea, where the ice advance is relatively modest, but a large ice tongue develops (Figure 3.2). In Northeast Greenland, the ice sheet grounds ~150 km off the coast, consistent with reconstructions from that region (Funder et al. 2011). The grounded ice sheet here reaches a maximum thickness of 2000 (2300) m in the standard (INN) domain. In East Greenland, grounded ice advances and reaches the edge of the continental shelf in the Scoresby Sund region. South of Scoresby Sund, large ice shelves develop and the ice sheet grounds on the continental shelf off the coast of the Kangerdlugssuaq, Sermilik, and Bernstorffs fjord systems (Figure 3.2c). South of
Bernstorffs, ice sheet advance is limited by warm ocean temperatures, leaving large parts of the continental shelf ice-free (Figure 3.2c). West Greenland shows the greatest sensitivity to the choice of domain. For the INN model domain, convergent ice flow results in a large ice shelf that covers (but does not ground in) Baffin Bay, and allows the grounding line along the West Coast of Greenland to advance onto the continental shelf. This ice shelf does not develop for the Greenland-only domain (Figure 3.2). However, in both the standard and extended domains, increased discharge via Jakobshavn Isbræ results in an advance in the grounding line onto the continental shelf in Ilulissat Fjord/Disko Bugt and the development of a large ice shelf there (Figure 3.2c).

For the LGM configuration, the surface mass balance (50 Gt yr$^{-1}$) is balanced by calving and ocean melt in approximately equal proportions, resulting in a steady state configuration (Figure 3.4). For the VAR forcing, ice sheet volume and extent varies little from 20 ka – 16 ka. Surface mass balance increases slightly from 20 ka to 17 ka due to increasing accumulation rates, but by 17 ka this effect is balanced by increased losses from ocean melt, and by 16 ka the net mass balance becomes negative and the ice sheet begins to shrink in volume. The ice sheet loses mass at a modest rate (-50 Gt yr$^{-1}$) from 16 ka to 14.5 ka, at which point rapid atmospheric warming drives widespread surface melting. Mass loss peaks at -250 Gt yr$^{-1}$ at ~14ka and then falls to -50 Gt yr$^{-1}$ from 12.5–11 ka, during the Younger Dryas, removing much of the shelf and marine-based ice and bringing the ice sheet, in most sectors, out of contact with the ocean. Three periods of extensive surface melt occur during the early Holocene, driving mean wastage rates of -100 Gt yr$^{-1}$ until 6 ka when a Holocene minimum configuration is reached at 6.72 m sle. Following the Holocene minimum, a reduction in surface melt and progressive cooling
towards the present-day result in positive ice-sheet mass balance of ~50 Gt yr\(^{-1}\). At present day, the modelled ice sheet has a volume of 7.36 m s.e. (Figure 3.4). The FIX forcing differs from this nominal simulation in that the LGM ice volume is ~15% greater, due to enhanced accumulation rates when modern seasonality is applied during the glacial period. The mass balance of the FIX tracks the VAR simulation, but does remains positive until ~14.5 ka, and does not begin aggressively losing mass until ~11.5 ka, i.e. following the Younger Dryas. The RCM scenario differs in that mass balance stays mostly positive prior to the Holocene, with surface melt rates that are generally similar to the VAR scenario, indicating the predicted seasonality in the RCM is similar to that of Trace21k (Liu et al. 2009). For the RCM scenario, the climate stays generally cool until ~10 ka, when increased greenhouse gas concentrations and a warm orbital configuration lead to regional temperature increases and drive peak mass loss rates of ~400 Gt yr\(^{-1}\) at ~9 ka. In contrast to the FIX scenario, the RCM scenario also shows a Holocene minimum prior to present day, with a minimum ice sheet volume of ~7.34 m s.e. at 4 ka (Figure 3.4, 3.6). Finally, the INN scenario reaches a much larger LGM maximum volume of ~15 m s.e. at 18 ka due to increased buttressing with the Innuitian ice sheet and the associated advance father onto the continental shelf. This scenario shows strong mass loss peaks both at 14.5 ka and 11.5 ka, as the abrupt warmings associated with these transitions put extensive areas of this larger ice sheet into the ablation zone, leading to significant (up to 1500 Gt yr\(^{-1}\)) surface melt rates (Figure 3.4).
3.4.1 North Greenland

In general, ice retreat in North Greenland occurs early in the deglaciation and responds primarily to surface melting following the Bølling-Allerød warming (Figure 3.4). However, retreat in Northwest and Northeast Greenland differ in their timing due to the influence of the Nares Strait ice shelf in the Northwest (Figure 3.5). To the north, Petermann glacier is the only major fast-flow feature that remains after the disintegration of the Nares Strait ice shelf (Figure 3.5). By 6.5 ka, it has retreated following Holocene maximum warmth to ~100 km behind its present-day grounding line, after which the glacier begins to readvance down its trough (Figure 3.8).

$^{10}$Be data indicate that deglaciation in North Greenland (Figure 3.10e) occurred at about 12.5 ka (Sinclair et al. 2016). The average deglaciation date in the FIX, VAR, no ACC, RCM and INN scenarios are 10.0, 13.9, 14.7, 7.5 and 10.5ka, respectively. The VAR scenario captures the highest proportion of the $^{10}$Be boulder ages within their published uncertainty for Thule, but for Johannes V Jensen Land and Store Koldewey, the INN scenario is equally good or better (Table 3.1, Table 3.2, Figure 3.6–3.10). Relative to other regions in North Greenland, the deglaciation in Thule is delayed due to buttressing by a large ice shelf covering the Nares Strait; the INN scenario may overestimate the impact of this buttressing at the shelf-adjacent Thule site, delaying deglaciation and resulting in mismatch with the $^{10}$Be data here. The ice margin in northeast Greenland reaches ~ 50-100 km behind its modern-day extent at ~7 ka, with the most pronounced retreat near the outlets of Nioghalvfjerdsfjord and Zachariae glaciers (Figure 3.9). In Johannes V. Jensen Land, the minimum extent of the ice sheet is within 30 km of the present-day margin (Figure 3.9).
3.4.2 East Greenland

In general, deglaciation in East Greenland occurs early in the deglaciation at about 14 ka, nearly synchronously with North Greenland. In Kangerdlugssuaq and Sermilik, the model shows significant thinning during the deglaciation but most locations where data is available remain ice-covered because of well-developed ice streams there and narrow fjord geometries that are not resolved by our 10 km model grid (Figure 3.2).

$^{10}$Be data indicate that deglaciation in Scoresby Sund (Figure 3.10g) occurred on average at 13.6 ka. The average deglaciation date in the FIX, VAR, no ACC, RCM and INN scenarios are 9.7, 14.5, 16.0, 7.0 and 11ka, respectively. The VAR scenario captures the highest proportion of boulder ages here (Table 3.3). After deglaciation, the ice margin stays relatively close to its Holocene extent, although the ice sheet in the southern part of the domain shows a late Holocene retreat at ~4 ka up to 100 km behind the present-day margin due to retreat along the trunk of Vestfjord Glacier (Figure 3.9).

Deglaciation recorded in Kangerdlugssuaq and Sermilik (Figure 3.10h) is difficult for the model to capture, because the data locations lie along narrow fjord systems that are just below the resolution of the simulations. None of the simulations capture individual boulder ages accurately and in all simulations ice re-advances across presently ice-free areas after the early Holocene. These regions also host major outlet glaciers, so even aggressive forcing scenarios do not result in full deglaciation of the fjords under modern forcing at 10 km resolution. We note that a finer grid scale is likely necessary to accurately simulate the dynamics of deglaciation in these settings.
3.4.3 South Greenland

$^{10}$Be data indicate that deglaciation in Bernstorffs and Narsarsuaq (Figure 3.10i) occurred on average at 10.8 ka. The average deglaciation date in the FIX, VAR, no ACC, RCM and INN scenarios are 6.3, 8.9, 6.2, 1.3 and 8.6ka, respectively. The VAR and FIX scenario capture the highest proportion of the boulder ages in these regions, suggesting the ice-core-derived climate forcing best represents deglacial climate change in southeast Greenland, but that the seasonality of the forcing does not have a significant impact on the deglaciation here. Nevertheless, the VAR simulation deglaciates ~1–2 ka earlier than the FIX simulation, giving a better match to the mean $^{10}$Be-inferred age here (Figure 3.10i). Following deglaciation, the ice margin retreats along the trunks of major outlet glaciers, up to 40 km behind the modern ice margin. The timing of minimum glacier extent varies from 10 ka to 4 ka, and both the VAR and RCM scenarios show the ice retreating slightly behind its modern-day margin (Figure 3.8–3.10).

Deglaciation in Paamiut and Nuuk (Figure 3.10f) occurred on average at 11 ka. The average deglaciation date in the FIX, VAR, no ACC, RCM and INN scenarios are 8.6, 6.8, 7.0, 12.9 and 5.5ka, respectively. The VAR, FIX, and RCM scenarios do an equally good job of simulating deglaciation in these regions (Table 1). After deglaciation, the ice margin remains within 20 km of the modern margin in both Paamiut and Nuuk (Figure 3.5–3.10).

3.4.4 West Greenland

Deglaciation histories in West Greenland are the most variable, and the most data exist in these areas to constrain the spatiotemporal dynamics of retreat. In general, $^{10}$Be
data show a north-south gradient in the average age of deglaciation, with southernmost areas in West Greenland (Sisimiut-Kangerlussuaq) deglaciating later than the northernmost areas (Upernavik) (Figure 3.10a-d). The INN and RCM simulations both result, for different reasons, in delayed deglaciation in this region, resulting in offsets relative to the ice-core-climate forced simulations (VAR and FIX) of up to 5ka, and leading, in general, to a better match with $^{10}$Be for south- and central-west Greenland (Table 3.11, Figure 3.5–3.10).

$^{10}$Be data indicate that deglaciation in Sisimiut-Kangerlussuaq (Figure 3d) occurred on average at 9.9 ka. The average deglaciation date in the FIX, VAR, no ACC, RCM and INN scenarios are 10.1, 13.4, 12.7, 8.4 and 10.0ka, respectively. The FIX and INN scenarios do the best job of capturing individual boulder histories in this region. During the early Holocene, the ice margin retreats up to 100 km behind the modern margin, with progressively greater retreat towards the north (Figure 3.8–3.10).

$^{10}$Be data indicate that deglaciation in Disko Bugt (Figure 3.10c) occurred on average at 9.1 ka. The average deglaciation date in the FIX, VAR, no ACC, RCM and INN scenarios are 8.5, 12.4, 12.0, 9.2 and 10.7ka, respectively. The RCM scenario does the best job of capturing individual boulder histories in this region. During the early Holocene, the ice margin here also retreats up to 100 km behind the modern margin, with less retreat towards the north and more retreat along deep troughs; for example, the trunk of Jakobshavn retreats approximately 100 km behind its present-day margin (Figure 3.8–3.10).

$^{10}$Be data indicate that deglaciation in Uummannaq (Figure 3.10b) occurred on average at 9.6 ka. The average deglaciation date in the FIX, VAR, no ACC, RCM and
INN scenarios are 9.1, 13.7, 12.6, 8.6 and 12.2 ka, respectively. The FIX scenario does the best job of capturing individual boulder histories in this region (Table 3.3). During the early Holocene, the ice margin retreats ~50 km behind the modern margin (Figure 3.8–3.10).

\(^{10}\text{Be}\) data indicate that deglaciation in Upernavik (Figure 3.10a) occurred on average at 13.2 ka. The average deglaciation date in the FIX, VAR, no ACC, RCM and INN scenarios are 8.9, 13.0, 12.5, 14.7 and 10.1 ka, respectively. The VAR scenario does the best job of capturing individual boulder histories in this region. Deglaciation in Upernavik occurs in two pulses; the ice sheet retreats onto land, exposing some of the locations where \(^{10}\text{Be}\) data is available, by 14 ka (Figure 3.6). A second phase of the deglaciation occurs at 10 ka, when the grounding line begins to retreat again, culminating in a retreat behind the present-day margin of ~30–40 km (Figure 3.8–3.10).

3.5 Discussion

Previous studies have made different, and somewhat contradictory, observations about the response of the Greenland ice sheet to abrupt climate changes during the deglaciation. In a synthesis of existing data, Sinclair et al. (2016) showed that deglaciation progressed clockwise around the Greenland margin, with the earliest deglaciation in East Greenland (12–11 ka), followed by South Greenland (ca. 11–10 ka) and finally West Greenland (ca. 10–7.5 ka). Our simulations indicate that the asynchronous nature of the deglaciation may be explained by a combination of factors, but especially the regional sensitivity of the ice sheet to abrupt climate changes during the deglaciation. In North Greenland, the VAR and INN scenarios capture the greatest proportion of \(^{10}\text{Be}\) data, with the INN scenario providing a slightly better match in
Johannes V Jensen Land and the VAR scenario providing a slightly better match in Thule (Table 3.1–3.2, Figure 3.10e); both simulations do an equally good job of capturing the ages in Scoresby Sund (Table 3.1–3.2, Figure 3.3g). This suggests that variable seasonality and buttressing with the Inuitian ice sheet are the salient controls on ice sheet retreat here. In addition, the RCM and FIX simulations do a poor job of capturing the $^{10}$Be data, suggesting that the abrupt climate changes not captured by the RCM are a key factor for driving retreat. The sensitivity of this region to abrupt climate changes, confirmed by our simulations, agrees with the observation that glaciers in Peary Land advanced during the Younger Dryas, a phenomenon not seen anywhere else in Greenland (Larsen et al. 2016).

In East Greenland, the VAR simulation does the best job at capturing the $^{10}$Be data. Although some studies have been hampered by nuclide inheritance in this region (i.e. Kelly et al. 2008), a comparison of boulder and bedrock $^{10}$Be ages from a local ice cap and the ice sheet margin in Milne Land revealed synchronous retreat during the Younger Dryas (Levy et al. 2016). The authors suggested rising summer temperatures during the Younger Dryas were responsible for this behavior. Although rising summer temperatures during this period are also captured in the RCM scenario due to increasing summer insolation, this simulation deglaciates much later than the $^{10}$Be data would indicate, suggesting that both seasonality and abrupt climate fluctuations were important for driving retreat here. Farther south, Dyke et al. (2014) suggested that oceanic warming drove deglaciation of Kangerdlugssuaq Fjord, while early Holocene warming was responsible for the slightly later deglaciation in the neighboring Bernstorffis Fjord.

Although all the simulations presented here include the same parameterization of ocean-
driven melting, the only the INN simulation captures the Kangerdlugssuaq data. Interestingly, this suggests that the buttressing provided by the Innuitian ice sheet can have far-reaching effects via its impact on ice thickness and accompanying changes in ice dynamics (Figure 3.2c). Indeed, for the INN simulation, the ice-dynamic changes are substantial even at the latitude of Kangerdlugssuaq as the large ice shelf in Baffin Bay allows the ice sheet to advance further onto the continental shelf in West Greenland and leads to upstream slowing and thickening (Figure 3.2c). The Bernstorffs data are indeed best matched in the VAR and FIX simulations, suggesting that while the seasonality of forcing may not have a strong impact here, the abrupt climate changes during the deglaciation do. The RCM simulation does not capture any of the boulder ages in Southeast Greenland, suggesting that the deglaciation in this region is strongly driven by Greenland-wide climate forcing and less sensitive to local factors. However, future work to more directly elucidate the impact of changing ocean temperatures on the fjord systems here is necessary to understand their sensitivity to different factors during the deglaciation (Dyke et al. 2014).

In Southeast and South Greenland, the FIX and VAR scenarios both capture the 10Be data better than the RCM scenario, although the FIX scenario does slightly better in Nuuk (Table 3.1, Figure 3h,i). This is contrary to the expectation, based on local terrestrial temperature reconstructions, that seasonality during the Younger Dryas was greater than today, i.e. summer temperatures were not much colder than today (Björck et al. 2002). Nevertheless, the inclusion of the abrupt temperature changes is necessary for capturing the spatial pattern of ice retreat. The INN simulation also captures the 10Be
data equally as well as the FIX simulation, due to the fact that they employ the same climate forcing.

Moving up the west coast, the FIX, RCM, and INN simulations capture more of the $^{10}$Be data than the VAR scenario in Sisimiut-Kangerlussuaq, Disko Bugt, and Uummannaq/upernavik, respectively (Table 3.1–3.2, Figure 3.10a-d). This suggests that, although the abrupt climate changes recorded in the ice core record influenced deglaciation as far up the west coast as Sisimiut-Kangerlussuaq, local climatic features and ice dynamics become more important from Disko Bugt and northward. In Disko Bugt, offshore mapping revealed multiple phases of retreat and readvance, with a large (~200 km) advance of Jakobshavn Isbæ at 12.2 ka, correlative with the Younger Dryas (Ó Cofaigh et al. 2013). However, the short duration of this event suggests it may be related to a transient surge rather than a true response to Younger Dryas climate. In Disko Bugt, the RCM simulation does better at capturing the $^{10}$Be-based histories than ice-core-based climate forcing (Table 3.1, Fig3.11c,d,f). This suggests that the Holocene thermal maximum (HTM), a feature not seen in the ice core reconstructions, but prominent in the regional climate modelling due to a peak in local summer insolation and increased SSTs in Baffin Bay, was a more important driver of ice retreat than the abrupt climate changes during the deglaciation in these regions. Holocene climate reconstructions (based on chironomid assemblages and δ$^{18}$O) from South and West Greenland have shown summer temperatures peaked 3–5°C above preindustrial during the HTM in these regions, similar to the magnitude of the response in the RCM (Figure 3.3, i.e. Frechette et al. 2009, Axford et al. 2009, Axford et al. 2011, Axford et al. 2013, Lasher et al. 2017). This
suggests that ice-core-based climatologies may not be the most accurate temperature forcings for understanding the glacial history of these regions.

Importantly, the inclusion of the abrupt climate change captured in the ice-core records (especially the Bølling-Allerød warming) substantially improves the model-data fit in most regions (Table 3.1). The main differences between the ice-core-derived climate forcing (VAR and FIX) and the modelled climate forcing (RCM) is that the former includes abrupt climate forcing, but with a modest HTM (3.3°C), whereas the latter shows a more prominent HTM but lacks the necessary mechanisms to produce a high amplitude Bølling-Allerød warming or Younger Dryas cooling. We also used a climate reconstruction (no ACC) which does not feature abrupt climate changes during the deglaciation. The HTM in the RCM scenario occurs from 7-6 ka, with temperatures peaking at ~5°C greater than preindustrial over parts of the ice sheet, especially the margins (Figure 3.3). Although the VAR and FIX simulations generally do a better job at capturing the boulder histories than the RCM simulation (Table 3.1), the RCM climatology does slightly better than all other reconstructions in the Disko Bugt region (Table 3.1). A HTM warmer than preindustrial in Disko Bugt is well-documented by climate reconstructions from lake sediments around Jakobshavn Isbræ, which show that temperatures peaked at ~7ka at ~5°C warmer than preindustrial (Axford et al. 2013). In addition, a change in the seasonality of precipitation between 6–4ka in West Greenland has been linked to reduced sea ice in Baffin Bay and the Labrador Sea (Thomas et al. 2016) driven by a warmer and more vigorous West Greenland Current (Perner et al. 2013). The RCM simulation captures these features more faithfully than the ice-core based reconstruction and gives a better match to the $^{10}$Be data, suggesting that local
Holocene oceanic conditions and their atmospheric consequences were a primary control on deglaciation in Western Greenland. However, the fact that the RCM simulations give a poor fit to the $^{10}$Be data in most other regions suggests a pronounced HTM was not as salient a factor in driving ice retreat for all sectors of the ice sheet (Table 3.1).

Although ocean forcing was included in our simulations, further work needs to be done to understand the role of oceanic warming on the regional pattern of ice-sheet deglaciation (see Chapter 4). Although abrupt climate changes have a strong expression in Greenland surface air temperature, they reflect changes in North Atlantic oceanic circulation (Buizert et al. 2014, Buizert et al. 2015). Changes in the circulation affect ocean temperatures at the surface and at depth, potentially resulting in warming of a few degrees over centennial timescales (Liu et al. 2009). Ultimately, changes in ocean circulation are transferred to the atmosphere and thus the two processes are tightly linked. Our results show that much of the deglaciation can be explained by the atmospheric warming, and that regional differences in the timing of deglaciation can be attributed to the regionally variable sensitivity to abrupt climate changes, local atmospheric effects, and ice dynamics during the deglaciation. However, such oceanic warming can also have a direct effect on melting parts of the margin that are in direct contact with warm waters. Some of the remaining data-model mismatch may be linked to regionally heterogeneous oceanic temperature histories and physical processes, including the potential for sudden changes resulting from incursions of warm water masses on continental shelves during past abrupt warming events (i.e. Liu et al. 2009). Indeed, the records of oceanic temperature change from different regions around Greenland show that the histories for different sectors of the ice sheet are different (i.e. Jennings et al. 2006, Winsor et al.).
In addition, improved bathymetry around the margin suggests that some sectors may be more prone to incursions of warm water than others (Morlighem et al. 2017). Inclusion of these factors in future studies will be necessary to resolve the remaining discrepancies between the modelled ice history and that inferred from the landscape.

The RCM forcing generally leads to a delay in deglaciation ages relative to $^{10}$Be data (Figure 3.6–3.10). However, it has a similar LGM maximum and modern-day ice volume to the VAR simulation, and these are the only simulations which capture an early Holocene minimum behind the present-day margin of the ice sheet. This minimum represents a reduction of the ice sheet by 45-60 cm s.l., roughly 6–8% of the modern ice sheet volume. The retreat of Greenland behind its present-day margin is well documented in South and Southwest Greenland (i.e. Carlson et al. 2014) and West Greenland (i.e. Young and Briner 2015), although the timing of this minimum is spatially heterogeneous (Larsen et al. 2017). We find that one explanation for the difference in timing of the Holocene minimum may be the different sensitivities of these regions to abrupt and gradual climatic changes during the deglaciation. The RCM simulation compares favorably to $^{10}$Be data in West Greenland and reaches a Holocene minimum at 4ka (Figure 3.9). In contrast, the VAR simulation compares more favorably to $^{10}$Be in Southwest Greenland and reaches a Holocene minimum at 8ka (Figure 3.8). Thus, the differences in the timing of minimum ice volume in these regions may be explained by different climatic histories elucidated by comparison with $^{10}$Be data. In other regions where a Holocene minimum is predicted by model simulations, no data presently exists for comparison (i.e., Northeast Greenland).
Other factors may explain why all simulations fail to capture some of the $^{10}$Be data within their published uncertainties. Firstly, regions where prominent ice-streams controlled ice dynamics during the LGM (i.e. Disko Bugt, Uummannaq) show, in general, poor model-data fit, with the model capturing fewer of the $^{10}$Be ages within their published uncertainty. Although we can accurately simulate ice dynamics for the present day, uncertainties in the bathymetry and basal properties of the continental shelf make simulation of dynamics in these areas problematic. For our analysis, we assume that all $^{10}$Be ages <19ka from the published database are equally valid, but cosmogenic isotope inheritance may still affect some of the measured ages, especially in north and east Greenland (Sinclair et al. 2016). This is confirmed by the predicted evolution of the basal thermal state in the ice sheet model for North Greenland; in Thule and Johannes V Jensen land, all of the simulations indicate the ice sheet is cold-based at the LGM, which would facilitate the preservation of older cosmogenic nuclides. In contrast, the ice sheet model predicts warm-based ice in all other regions for which $^{10}$Be data exist.

Additional data, especially relative sea-level (RSL) records, exist to constrain the evolution of the ice margin over time. These data have been used before as targets for ice sheet model simulations (i.e. Lecavalier et al. 2014). Although the utility of these data for assessing ice sheet models is clear, their interpretation hinges on accurate and detailed information about earth structure beneath the ice sheet. It is becoming increasingly clear that the distance to, and structure of, the lithosphere underlying Greenland is spatially heterogeneous (i.e. Rohogzhina et al. 2016, Stevens et al. 2016); additional work is needed within our modelling framework to accommodate these emerging observations including a coupled Earth/gravitation model (e.g., Gomez et al. 2013) and incorporate
RSL records into our analysis framework. However, the mechanisms we identify here as being important controls on the spatiotemporal pattern of ice retreat remain important.

3.6 Conclusion

We present simulations of the evolution of the Greenland ice sheet over the last deglaciation at higher spatial resolution than previous reconstructions, and importantly, considering processes that have previously been not been included. Forcing the ice sheet model with variable, time-evolving seasonality in the temperature record leads to a better simulation of deglaciation relative to exposure age data in nearly all regions, relative to a simulation where seasonality is fixed at present day values, especially in North and East Greenland. In North and much of West Greenland, inclusion of the Innuitian ice sheet is important for accurately resolving the retreat history, as the buttressing provided by the adjacent ice mass significantly delays deglaciation in those regions. In Southwest Greenland and Disko Bugt, asynchronously coupling a regional climate model to the evolving ice sheet produces the best match to deglaciation histories, suggesting a strong control of local SSTs and the Holocene Thermal Maximum on the deglaciation history there. However, this simulation does a poor job of capturing the deglaciation history other sectors of the ice sheet, at least in part due to the absence of abrupt climate change in the climate forcing. This suggests that abrupt climate changes were important in driving deglaciation of the Greenland ice sheet following the LGM. Ice-sheet retreat occurs earlier in the simulations using ice-core derived climate forcing that includes abrupt changes, indicating that a highly variable climate forcing gives better model-data agreement than a steadily warming one, despite the initial and final states being broadly similar.
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Table 3.1. Measured and simulated deglaciation timing. Bold type and darker boxes refer to the best-fit scenarios in each region.
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Table 3.2 Effect of using an expanded model domain (Inuitian) on deglaciation timing. Bold type and darker boxes refer to the best-fit scenarios in each region.
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Table 3.3 Fraction of $^{10}$Be deglaciation ages fit by the simulations. Bold type and darker boxes refer to the best-fit scenarios in each region.
Figure 3.1 Reconstructed temperature forcing used to drive ice sheet model simulations, shown as an anomaly relative to modern. Red: Fixed seasonality. Blue: Variable seasonality, with winter temperatures (DJF) and summer temperatures (JJA) shown separately. Yellow bars highlight two periods of abrupt climate change during the deglaciation: Heinrich Stadial 1 (HS1) and the Younger Dryas (YD).
Figure 3.2 Comparison of modern and last glacial maximum ice sheets. a) Modern ice sheet. b) LGM (21ka) ice sheet with standard domain. c) LGM (21ka) ice sheet with expanded Innuitian domain.
Figure 3.3 Simulated (RegCM3) temperature and precipitation for selected time slices during the last deglaciation. a) RCM under modern boundary conditions. b) RACMO 2.3 (Noël et al. 2015). c–k show RCM climatologies using boundary conditions from: c) 1ka, d) 3ka, e) 5ka, f) 7ka, g) 8ka, h) 9ka, i) 10ka, j) 12 ka, k) 14.5ka, l) 16ka.
Figure 3.4 Evolution of Greenland ice volume over the last 20 ka under different forcing scenarios. A) Total ice volume for four experiments: VAR, FIX, RCM, and INN. B) Total mass balance for each of the experiments. The minimum mass balance for the INN scenario at 11.5ka is -1700 Gt yr$^{-1}$. C) Surface melt rates for each of the experiments.
Figure 3.5 Timeslices of ice-sheet extent for the scenarios described. a) fixed seasonality, b) variable seasonality, c) no abrupt climate change (Buizert et al. 2018). Timeslices shown are, from left to right: 16ka, 12ka, 8ka, 4ka. For 16ka, the reconstructed extent of the LGM GrIS is shown (Funder et al. 2011). For 4ka, purple dots indicate $^{10}$Be observations which are fit (within two sigma) by the modelled deglaciation timing; pink dots are not fit (Sinclair et al. 2016).
Figure 3.6 Ice margin positions throughout the deglaciation for the scenarios described in the text: a) 16ka, b) 15ka, c) 14ka, d) 13ka. Orange: seasonality fixed at modern (Buizert et al. 2018). Green: variable seasonality (Buizert et al. 2018). Blue: No abrupt climate change (Buizert et al. 2018). Yellow: RegCM3 climatology. Circles show $^{10}$Be data which indicate the ice margin had retreated inland of the sample by the date in each panel.
Figure 3.7 Ice margin positions throughout the deglaciation for the scenarios described in the text: a) 12ka, b) 11ka, c) 10ka, d) 9ka. See Figure 3.6 for full description.
Figure 3.8 Ice margin positions throughout the deglaciation for the scenarios described in the text: a) 8ka, b) 7ka, c) 6ka, d) 5ka. See Figure 3.6 for full description.
Figure 3.9 Ice margin positions throughout the deglaciation for the scenarios described in the text: a) 4ka, b) 3ka, c) 2ka, d) 1ka. See Figure 3.6 for full description.
Figure 3.10 Summary of deglaciation timing in $^{10}$Be data (filled probability distribution), $^{14}$C data (black probability distribution), and selected ice sheet model simulations: Innuitian domain (blue circles), RCM climatology (red circles), fixed seasonality (gray circles), and variable seasonality (black circles). Colored regions in the central map refer to the following panels: a) Upernavik, b) Uummannaq, c) Disko Bugt, d) Sisimiut-Kangerlussuaq, e) Thule, Johannes V Jensen Land, Store Koldewey, f) Nuuk, Paamiut, g) Scoresby Sund, h) Kangerdlugssuaq, Sermilik, i) Bernstorffs, Narsarsuaq.
CHAPTER 4

DISENTANGLING CONTRIBUTIONS OF ATMOSPHERIC AND OCEANIC FORCING IN DRIVING PAST DEGLACIATION AT GREENLANDIC OUTLET GLACIERS

4.1 Abstract

The disintegration of polar ice sheets is causing global sea level rise to accelerate. The Greenland ice sheet has been losing mass for at least the past three decades due to increased ocean melting, discharge (calving), and ocean melt. Longer records are necessary to place these losses into context and understand the long-term stability of the ice sheet in response to these losses. The deglaciation of the ice sheet out of the Last Glacial Maximum (LGM) is the most recent time that Greenland lost significant mass to these three processes, and therefore provides a framework for understanding the spatiotemporal importance of different mass balance processes. Here we use data-constrained ice-sheet model simulations to understand the processes that drove the deglaciation in each of Greenland’s major drainages. We show that, while oceanic processes played the strongest role in driving deglaciation around much of the ice sheet, surface melting was especially important in controlling the timing of deglaciation throughout northwest and east Greenland. Our findings implicate surface melting as a critical process for the future of the northern Greenland ice sheet.
4.2 Introduction

Continued oceanic and atmospheric warming has led to an increase in mass loss from the Greenland ice sheet (GrIS; Mouginot et al. 2019). However, the importance of these processes for the modern ice sheet is regionally and temporally variable. Incursions of warm water and increases in surface melt can also lead to dynamic changes which enhance mass delivery to the ablation zone and marine margins (Straneo et al. 2013, Zwally et al. 2002). Enhanced oceanic melting triggered the break-up of the ice tongue at Jakobshavn Isbræ, resulting in acceleration and enhanced mass loss in the early 2000s (Holland et al. 2008, Velicogna et al. 2014); recently, the same glacier has slowed down substantially due to local oceanic cooling, limiting mass loss (Khazendar et al. 2019). For example, west Greenland has a well-defined ablation zone that extends far inland (Noël et al. 2015), and increased surface melt, to a point, is likely accommodated by a well-established network of subglacial channels (e.g. de Fleurian et al. 2016). In contrast, northeast Greenland has historically had very little surface melt, but the recent proliferation of seasonal melt ponds near the outlets of the Northeast Greenland ice stream may have significant dynamic implications if the water eventually makes its way to the bed (e.g. Rathmann et al. 2017). In order to better understand the implications of ongoing changes for the long-term stability of the ice sheet, a longer-term perspective on the response of the ice sheet’s margins to different mass balance processes is crucial. However, obtaining such a perspective is limited by the short instrumental records of mass loss we have access to (e.g. Mouginot et al. 2019). Slightly longer records have been compiled through the use of historical photographs, and these reveal that the ice sheet’s mass has fluctuated on multi-decadal timescales in response to changes in climate.
A more long-term perspective is necessary to understand how the ice sheet is likely to respond to the coming centuries of warming atmospheric and oceanic temperatures (IPCC, 2013).

Here, we use simulations of the last deglaciation to extend our understanding of the GrIS sensitivity to oceanic and atmospheric mass loss properties. We use ice sheet model simulations which provide the best fit to a database of over 800 $^{10}$Be ages to partition the mass lost to calving, oceanic melting, and surface melting prior to the last deglaciation, which occurred diachronously around Greenland but nevertheless represents the most recent time when Greenland lost a significant amount of mass (~3–7 meters of sea level equivalent, or ~25–100% of its modern volume).

### 4.3 Methods

We used the data-constrained simulations presented in Chapter 3 to calculate the influence of ocean melting, calving, and surface melting on the evolution of each drainage within the GrIS prior to deglaciation. Drainages were first defined by the modern flow field and extended by following contours of ice velocity to the edge of the LGM ice sheet. For each time step, we calculated the total mass lost to ocean melting, calving, and surface melt within each grid box and then summed all of the grid boxes within each drainage to provide an integrated perspective of the factors driving ice sheet change.

To closer examine the role of oceanic processes in driving the deglaciation of the ice sheet, we ran four sensitivity tests using different ocean melt parameters and calculated their impact on the timing of deglaciation around the margin. The evolution of the ocean melting factor (O), a tunable parameter in our modeling framework, and ocean
temperature is shown in Figure 4.1. Ocean melt rate (OMR) is calculated in the ice sheet model following (Martin et al. 2011):

\[
OMR = \frac{O \cdot K_f \cdot \rho_w \cdot c_w}{\rho_i \cdot L_f} (T_o - T_f)^2
\]  

(1)

Where OMR is ocean melt rate (m a\(^{-1}\)), \(\rho_w\) is the density of water (1000 kg m\(^{-3}\)), \(c_w\) is the specific heat of seawater (4218 J kg\(^{-1}\) K\(^{-1}\)), \(\rho_i\) is the density of ice (917 kg m\(^{-3}\)), \(L_f\) is the latent heat of fusion (335000 J kg\(^{-1}\)), \(K_f\) is a transfer factor (15.77 m a\(^{-1}\) K\(^{-1}\)), and \(T_o\) is the ambient ocean temperature and \(T_f\) is the pressure melting temperature of the ice at the depth where it is in contact with the ocean (Martin et al. 2011, Pollard and DeConto 2016). We created a timeseries of \(T_f\) based on a subsurface temperature reconstruction (Winsor et al. 2012) and four timeseries of \(O\) to examine the role that the timing and magnitude of oceanic melting have on regional deglaciation (Table 4.1).

4.4 Results

The ocean forcing parameterization modulates the ice sheet’s response to deglacial climate change (Table 4.1). However, the best-fit scenario identified in Chapter 3 generally does the best job of fitting the available \(^{10}\)Be data regardless of the ocean forcing applied (Table 4.1). An earlier peak in ocean forcing (\(O_4\)) leads to an improved fit in Scoresby Sund and Paamiut, suggesting that warm Irminger current waters may have influenced these sites slightly sooner than data from the North Atlantic indicate (Figure 4.1). In many regions, especially throughout west Greenland (e.g. Nuuk and Sisimiut-Kangerlussuaq) the model fit in insensitive to the range of ocean forcings applied even though ocean melting is the largest component of mass lost prior to deglaciation. In northwest Greenland (Upernavik and Thule), the deglaciation is similarly insensitive to
the ocean forcing applied, consistent with surface melting being the most salient driver of deglaciation there.

Prior to regional deglaciation, the largest amounts of mass lost are as follows: surface melt ($1779 \times 10^2$ Gt) in northwest Greenland (Figure 4.2); ocean melt ($4097 \times 10^2$ Gt) in central west Greenland (Figure 4.3); ocean melt ($5202 \times 10^2$ Gt) in southwest Greenland (Figure 4.4); ocean melt ($6978 \times 10^2$ Gt) in southeast Greenland (Figure 4.5); surface melt ($2365 \times 10^2$ Gt) in east Greenland; and ocean melt ($1611 \times 10^2$ Gt) in north Greenland (Figure 4.6). In summary, for northwest and east Greenland, surface melt is the largest component of mass loss prior to deglaciation (46% and 64%, respectively). In central west, southwest, and southeast Greenland, ocean melt is the largest component of mass loss prior to deglaciation (61%, 58%, and 55%, respectively. In north Greenland, surface melt and ocean melt drive a similar fraction of the total mass loss (36% and 45%, respectively).

4.5 Discussion

During the LGM, the ice sheet is especially vulnerable to ocean melting, because in its advanced state its margins are marine-terminating nearly everywhere around Greenland. Thus, oceanic processes are expected to play a prominent role in driving mass loss everywhere. However, the configuration of the ice sheet in different regions modifies its sensitivity to mass balance processes. Glaciers which occupy deep troughs (e.g. Uummannaq) experience higher rates of ocean melt due to the lower pressure melting point at depth, whereas low-slung regions of the ice sheet are especially sensitive to abrupt changes in temperature which can rapidly change the size of the ablation zone. We
analyze our results in this context, keeping in mind that as the ice sheet evolves its sensitivity to mass balance processes will change. The advance of the ice sheet onto the continental shelf is limited in east Greenland due to the low accumulation rates during the last glacial maximum, which limits the potential influence of oceanic processes compared with west Greenland, where the ice flux is higher and the ice sheet can thus maintain a more advanced position on the continental shelf (see Figure 3.2).

Our ice sheet model incorporates a calving scheme based the divergence of the ice flow field (Pollard and DeConto 2016), which does poorly against observations of current calving regimes in Greenland (Choi et al. 2018). However, this scheme has been extensively validated in Antarctica (Pollard and DeConto 2016). The expanded LGM GrIS, with its marine terminating margin and extensive ice shelves, was probably more similar to modern Antarctica than to modern Greenland (see Figure 3.2). Thus, our choice of calving parameterization is appropriate for the deglaciation, but future studies to look at processes such as the early Holocene disintegration of remnant ice tongues would benefit from considering alternative calving parameterizations.

Recent work has found that oceanic forcing is critical for simulating glacial-interglacial fluctuations of the GrIS (Tabone et al. 2018). Our results confirm the finding that these processes are critical for capturing the dynamics of the deglaciation around most of the ice sheet and should be considered in future work (e.g. Cuzzone et al. 2019). However, our results add nuance to this conclusion by demonstrating that ocean melting played a stronger role in some drainages compared to others during the last deglaciation, and highlight the critical role played by surface melt in certain drainages.
Our results show that the deglaciation of East Greenland, including the catchments of the Northeast Greenland Ice Stream, was driven in large part (64%) by surface melting. This suggests that surface melting may have also played an important role in the retreat of the northeast margin of the Greenland ice sheet to ~70 km behind its present-day margin during the middle Holocene (Larsen et al. 2017). In this context, the increasing prevalence of seasonal surface meltwater in northeast Greenland may be a harbinger of more dramatic changes.

Our results also indicate that surface melt played a dominant role in driving deglaciation in Northwestern Greenland (46%). Today there is little surface melting in northwest Greenland, and recent increases in mass loss mostly attributed to dynamic processes (Khan et al. 2010). During the LGM, convergent flow between Ellesmere Island and northwestern Greenland (Nares Strait) led to the development of an extensive ice shelf there that extended south into Baffin Bay (Jennings et al. 2018). This ice shelf did not completely disintegrate until after the Bølling-Allerød, when summer temperatures warmed considerably. One of the reasons surface melt played such a large role in the deglaciation of northwestern Greenland is due to the break-up of the ice shelf, which no longer exists. However, as in northeast Greenland, northwest Greenland is one of the few regions where floating ice tongues still occupy glacial fjords (Reilly et al. 2017). Ice tongues restrict ice flow by providing buttressing against fjord walls, and their collapse can lead to a cascade of dynamic changes (Van der Veen et al. 2011). The importance of surface melt for driving the collapse of ice shelves in northwestern Greenland during the deglaciation suggests that the same processes could become
responsible for future disintegration of remaining ice tongues, some of which already show extensive evidence of seasonal surface melt features (Hill et al. 2017).

We find that ocean melt was the dominant driver of mass loss from central west, southwest, southeast, and northern Greenland over the last 20ka. These regions are notably also areas where large tidewater glaciers, some of which sit in deep subglacial troughs are still in contact with the ocean (Joughin et al. 2012). Ocean melt has driven a large proportion of the total mass lost in these regions over the last 20ka and should thus be closely monitored as these regions continue to receive incursions of warm water. As long as there are marine terminating margins these oceanic processes can excite dynamical changes which can propagate far inland and mobilize large volumes of ice, which may continue to be shed to the ocean until the oceanic forcing abates (e.g. Khazendar et al. 2019).

4.6 Conclusion

The Greenland ice sheet is currently losing mass at an accelerating rate due to both oceanic and atmospheric warming. This study provides much needed perspective on these ongoing changes by identifying the processes that have driven mass loss for each major drainage of the Greenland ice sheet over the last 20,000 years. We find that surface melt dominates mass loss in northwest and east Greenland, whereas oceanic processes drive mass loss along the southwest and southeast coast as well as in north Greenland. As boundary conditions (i.e. sea ice, overturning circulation) evolve, the dominant processes may change. However, this study marks an important step towards predicting the processes that are likely to drive mass loss into the future.
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Table 4.1 Effect of different ocean forcing parameters on model fit to available $^{10}$Be data. FIX: seasonality fixed at modern, VAR: seasonality variable over the last 21ka, no ACC: no abrupt climate change. Suffixes 1,2,3,4 refer to different ocean melt parameterizations (Figure 4.1). Bold type and darker boxes refer to the best-fit scenarios in each region.
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Table 4.2 Total mass lost to calving, ocean melt, and surface melt ($10^2$ Gt) prior to deglaciation within each drainage of the Greenland ice sheet. Bold type and darker boxes refer to the best-fit scenarios in each region.
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**Table 4.3** Fraction of total mass lost to calving, ocean melt, and surface melt prior to deglaciation within each drainage of the Greenland ice sheet.
Figure 4.1 Ocean forcing used to drive ice sheet model simulations. a) Temperature perturbation over the last 21,000 years based on a North Atlantic subsurface temperature reconstruction as an anomaly from modern (Winsor et al. 2012). b) Four different realizations of ocean melting factor $O$ from equation 1.
Figure 4.2 Mass balance for northwest Greenland for the last deglaciation. Green: total mass balance. Red: surface melt. Blue: calving. Purple: ocean melt. Gray histograms and black probability distribution show the timing of modelled deglaciation for grid cells where 10Be data are available. Black line shows the mean deglaciation timing. The total mass lost in this region prior to the mean deglaciation due to calving (blue box), ocean melt (purple box), and surface melting (red box) are shown in that order in the upper left of each box ($10^2$ Gt). The map of Greenland shows the region of interest filled with the color that corresponds to the mass balance process to which the greatest amount of mass was lost prior to deglaciation.
Figure 4.3 Mass balance for central west Greenland for the last deglaciation. See Figure 4.2 for full description.
Figure 4.4 Mass balance for southwest Greenland for the last deglaciation. See Figure 4.2 for full description.
Figure 4.5 Mass balance for southeast Greenland for the last deglaciation. See Figure 4.2 for full description.
Figure 4.6 Mass balance for east Greenland for the last deglaciation. See Figure 4.2 for full description.
Figure 4.7 Mass balance for north Greenland for the last deglaciation. See Figure 4.2 for full description.
CHAPTER 5

PLIO-PLEISTOCENE MEGAFLOODS AS A MECHANISM FOR GREELANDIC MEGACANYON FORMATION

5.1 Abstract

The present-day Greenland ice sheet sits on a complex network of canyons thought to be pre-glacial and fluvial in origin. Such a history for these features implies they have exerted influence on the ice-sheet since its inception. The largest of these subglacial canyons, one of the longest in the world, terminates in northwest Greenland at the outlet of the Petermann Glacier. Yet the geologic history of the canyon, and similar features in Northern Greenland, remain unknown. Here we present numerical model simulations of the early history of the Greenland ice sheet and show that interactions between climate, the growing ice-sheet, and pre-existing topography may be largely responsible for the excavation of the canyon via repeated catastrophic outburst floods. Our results have implications for interpreting sedimentary features beneath the Greenland Ice Sheet and around its marine margins and suggest a previously undocumented mechanism for causing abrupt climate changes prior to the establishment of widespread, episodic ice-cover in North America and Eurasia.

5.2 Introduction

Subglacial topography exerts a primary control on ice-sheet dynamics and is thus an important boundary condition for simulations of ice-sheet behavior. Mapping of the subglacial landscape in Greenland reveals complex networks of canyons, many of which
are thought to have a fluvial, rather than glacial, origin (Bamber et al., 2013; Morlighem et al., 2014; Cooper et al., 2016). An especially prominent trough extends ∼750 km inland from the terminus of Petermann Glacier in northwest Greenland to nearly the summit of the present-day ice sheet (Figure 5.1) (Bamber et al., 2013). The canyon may affect subglacial water routing throughout northwest Greenland, and thus has the potential to influence a large portion of the ice sheet. Yet despite the canyon’s influence on the overlying ice sheet and its rank as the longest canyon in the world, little is known about the processes that led to its formation.

The morphology of the V-shaped canyon suggests it was formed by running water prior to extensive glaciation of Greenland (Bamber et al., 2013; Cooper et al., 2016). While it is clear from off-shore records that the extent and duration of ice-cover in the northern hemisphere increased about 2.7 million years ago (Ma), it remains possible that Greenland hosted some land-based ice as early as the Eocene (Eldrett et al., 2009; DeConto et al., 2008). In fact, a recent study argued for persistent ice cover in the high-topography, high-precipitation environment of Southeast Greenland over the last 7 million years (Bierman et al., 2016). This does not exclude ice-free conditions at times in the rest of Greenland during this period – indeed, another recent study argued for episodic ice-free conditions throughout the rest of Greenland as recently as the last interglacial period (125 thousand years ago, ka) based on basal material below the present-day location of the ice-sheet summit (Schaefer et al., 2016). The temporal and spatial extent of ice-free conditions remains mostly unknown, even during the Late Pleistocene, clouding the timing and mechanism(s) of mega-canyon formation. Here we present a new mechanism for the formation of the mega-canyon based on ice-sheet model
simulations of the Greenland Ice Sheet’s (GIS) early history. We show that climate and bedrock topography exert strong controls on the dynamics of ice-sheet inception, and as a consequence, the early formation of the GIS may have been accompanied by rapid retreat of the northern GIS margin and the repeated, catastrophic drainage of large proglacial meltwater lakes with volumes comparable to those left in North America by the waning Laurentide ice sheet during the last deglaciation.

5.3 Methods

5.3.1 Climate Forcing

To generate climate forcing, we used the current version of the GENESIS GCM (Alder et al., 2011), run in slab ocean mode using the ice-free Northern Hemisphere topography and a modern orbit. The GCM was run for 60 model years (~30 years beyond equilibrium), saving 6-hourly output to drive a Regional Climate Model (Pal et al. 2007) run at 40-km resolution. The results of the RCM were averaged over 10 years and the temperature and precipitation fields used to drive the ice-sheet model.

To force the model through simulated glacial-interglacial cycles for 1.2 Ma, we generate a synthetic temperature anomaly that varies from warm interglacial-like temperature to cool glacial-like temperatures and apply this to the ice-free climate state to approximate glacial-interglacial climate changes over Greenland (Figure 5.2). We chose this simplified approach to the climate forcing rather than attempting to simulate the time-evolving climate which is computationally infeasible on these long timescales. Simulations of the evolution of the GIS through actual glacial-interglacial cycles must explicitly consider external drivers and a range of potential climate scenarios and
uncertainties in the coupling between the ice sheet and climate system, which we leave to future work. Nevertheless, our approach takes into account three fundamental characteristics of global climate evolution over the period that the GIS was becoming persistent on a continental scale (Lisiecki and Raymo, 2005): glacial periods become increasing colder and are more variable than interglacial periods, and randomly located "super-interglacial" periods cause warming above normal levels (Figure 5.2). This approach mimics the effects of orbital cyclicity on GIS surface mass balance over the 1.2 Ma of the simulation (Figure 5.2).

The presence and geometry of the ice sheet on Greenland strongly influence its climate. Two key factors are the high albedo and topography of the ice sheet, which directly control surface temperatures and precipitation. To understand pre-glacial Greenland climate, we ran a regional climate model (RCM) with deglaciated boundary conditions. The boundary conditions were set by removing ice cover everywhere except Greenland’s eastern mountains, resulting in an ice sheet with a total sea level equivalent (sle) of ~0.16 m remaining on Greenland (and consequently, raising global sea level by 7.20m). The bedrock was then allowed to rebound to isostatic equilibrium, resulting in elevation changes ranging from 10s-100s of meters near the coast to 100s-1000 meters inland. We make no estimation of the total volume of sediment lost to glacial erosion since the inception of the ice sheet, assuming instead that most of the bed topography predates glaciation as has been argued by other studies (e.g. Cooper et al., 2016), so the effects of glacial erosion are ignored. This assumption is further supported by measured and inferred rates of glacial erosion which are orders of magnitude smaller than the changes due to isostatic adjustment (1—10 m Ma⁻¹, Egholm et al. 2017).
5.3.2 Ice Sheet Model

We use the Parallel Ice Sheet Model (PISM) (Bueler and Brown, 2009; PISM Authors, 2016) run at 10 km resolution with a positive degree-day (PDD) scheme for mass balance. The PDD scheme used coefficients of 3 and 8 mm °C⁻¹ day⁻¹ for melting of snow and ice, respectively, and surface temperature was adjusted to the ice-sheet geometry using a lapse rate of 6.5 °C km⁻¹. Though the resolution is comparably coarse and the mass balance scheme simple, these choices enable us to run multiple realizations of ice-sheet inception on timescales of Ma.

We note that on the timescales considered here, changes in boundary conditions are poorly constrained, and a simple modelling approach is ideal to study processes and mechanisms that may be modified by, but are ultimately independent of, other changes to the climate system. As has been documented by many previous studies, a strong hysteresis effect controls the growth and decay of ice sheets due in large part to positive climatic feedbacks related to albedo and elevation (modulated by glacial isostatic adjustment) as an ice sheet grows (Weertman, 1961; Abe-Ouchi and Blatter, 1993). In order to initiate retreat, at least one of two main mechanisms must be active: negative surface mass balance or dynamic thinning and retreat. A warming climate can lead to a change in the accumulation area ratio which is unfavorable for the ice sheet, leading to surface melting and mass wastage. In addition, grounded ice retreating into overdeepened basins, if destabilized, can lead to rapid ice sheet disintegration via the marine ice sheet instability (Weertman, 1974; Schoof, 2007). PISM captures both of these processes, which is an advantage over simpler models. For the purposes of this study, we emphasize
that while the modelled timing and extent of Greenland inception and retreat phases are sensitive to the climate forcing applied as well as parameter choices for the ice-sheet model and its coupling to the atmosphere, the mechanism for generating outburst floods is independent of these choices. At the ice sheet’s peak volume, much of the interior of northern Greenland is grounded 1km below sea level. Rapid warming into an interglacial period, as is characteristic for glacial-interglacial transitions during the Plio-Pleistocene, leads to the destabilization and collapse of the northern margin before the bedrock has time to adjust. This leads to the potential for substantial terrestrial water storage, which would then be vulnerable to drainage via known processes (i.e. ice dam disintegration). Relatively little is known about Greenland’s climate prior to the onset of major glaciations there and what information is available paints a picture strikingly different from the Greenlandic climate we know from modern observations and ice-core records, with flora and fauna indicating reduced sea ice extent and much warmer temperatures that today (Funder et al., 2001; Rybczynski et al., 2013). However, it is clear from marine and terrestrial archives that the ice sheet has also waxed and waned through the Pleistocene, and it is the consequences of that fundamental observation that we wish to highlight here (Flesche Kleiven et al., 2002; Schaefer et al., 2016), with the potential to cause repeated flood events.

5.3.3 Water Routing and Erosion

To calculate paleotopographies at the highest resolution possible, we calculated the lithospheric deflection by subtracting the bedrock elevation for each timestep of PISM from its modern value. This field, which was at the resolution of the model (10 km), was then downscaled to the resolution of available bed products (150m) to calculate flowpaths
based on the local topographic gradient (Schwanghart and Scherler, 2014). We calculated all of the marine-terminating stream networks for deglaciated topographies to define proglacial lake catchments. Then, starting from the marine terminus of the stream network, we defined a sill as the first landward topographic maximum above sea level. The elevation of this sill was used as the elevation contour for defining the maximum instantaneous volume of a proglacial lake.

To estimate the amount of erosion caused by megafloods, we use two approaches. First, we use modern studies of bedrock erosion caused by catastrophic floods to contrain the order of magnitude of landscape change we would expect from our modeled flood volumes. Second, we use a parameterization that predicts spillway erosion rate:

\[
\frac{dz_s}{dt} = -(\rho g S)^a \cdot k_e \cdot (z_l - z_s)^a
\]

Where the vertical erosion rate is a function of the density of water \((\rho)\), acceleration due to gravity \((g)\), channel slope \((S)\), rock erodibility \((k_e)\), lake surface height \((z_l)\), sill height \((z_s)\), and the erosion law exponent \((a = 1.5)\) (Garcia-Castellanos & O’Connor 2018). We consider the range of possible values for uncertain parameters to provide a rough estimate of erosion rates for our modeled flood volumes.

5.4 Results

In our simulations, the orbitally oscillating climate leads to fluctuations in ice-sheet volume, but the general cooling trend contributes to the progressive establishment of the GIS (Figure 5.2). After ~600 kyr in our simulations, the GIS begins to oscillate between modern (6—8 m sle.) during interglacial periods and larger (12—14 m sle.) during glacial periods. For the largest glacial ice sheets, a warmer-than-average
interglacial period is necessary to trigger a deglaciation, and the GIS can stabilize or continue to grow through multiple glacial-interglacial cycles (Figure 2). During these skipped interglacial periods, the bedrock continues its isostatic adjustment to the increasing ice load, leading to the development of an overdeepened basin in central Greenland with subglacial outlets in both the northwest (via Petermann) and the northeast (via the Northeast Greenland Ice Stream (NEGIS)) (Figure 5.1).

When a warm interglacial occurs after a period of prolonged stability, the GIS rapidly disintegrates due to the positive surface mass balance-elevation feedback (Weertman, 1961) and the ice retreating into regions with submarine bed elevations. Large parts of western Greenland, i.e. within the modern-day ablation zone, begin to melt due to surface imbalance. As mass loss continues, two styles of deglaciation are observed. In both cases, retreat of the ice sheet along troughs in the northwest and northeast creates a saddle which rapidly ablates, revealing a large subaerial basin with direct access to the ocean blocked only by small remnant ice caps (Figure 5.1). Variations in the thickness of ice in northeast and northwest Greenland prior to deglaciation determines whether the first ice-free route is observed through the Petermann canyon (Figure 5.3) or the outlets of NEGIS (Figure 5.1).

In many individual deglaciations, the Petermann canyon provides a transient outlet to the ocean (Figure 5.3). In some scenarios, a pathway is also available via the northeast, near the modern outlets of Nioghalvfjerdsfjord and Zakariæ. These deglaciation scenarios depict a previously undescribed mechanism with implications for landscape evolution. Following glacial periods when the ice sheet bed has isostatically adjusted to overlying ice load, rapid retreat reveals an overdeepened subaerial basin. Such
a basin could have collected runoff and meltwater on land, and an outburst flood could have been triggered by the disintegration of remnant icecaps near the margin or bedrock rebound; subglacial transport of meltwater is likely, but due to the steep surface slopes of the retreated ice sheet relative to the bedrock topography, subglacial hydropotential paths would closely follow surface potential (Figure 5.3). The mechanism for these outburst floods would then have been similar to those described for the Laurentide ice sheet during the last deglaciation (Teller et al., 2002; Clarke et al., 2004). To understand the potential of a large subaerial lake in central Greenland to cause outburst floods, we mapped the volume and hydropotential pathways during the two deglaciation scenarios described above (Figure 5.3). The volume of the lake below sea level peaks between 2,000 to 4,000 km³ during deglaciations. If the sill elevation is considered, the volume of the lake peaks between 15,000 and 18,000 km³ (Figure 5.2). The duration of time these basins exist as subaerial lakes is ~1–3 kyr, a timescale that is dictated by the viscoelastic response of the mantle to the reduced ice load. Though these ice-free depressions form an ice-dammed basin, the earth model does not account for the load caused by the accumulation of water. The lakes we simulate can be hundreds of meters deep in places where the glacial ice sheet reached its maximum thickness (~3 km), meaning this process would slow (not stabilize) the disappearance of these transient lake basins.

The subsequent regrowth phase is slightly different than the retreat phase, as the ice sheet gradually expands from inception points over Camp Century, the northeast coast, and southern Greenland (Figure 5.1). Aridity in the northwest, influenced by the Greenland’s topography and simulated storm tracks, restricts glacial inception there, and ice must instead advance from other regions to cover the area.
Based on the outcropping lithified sedimentary rocks mapped in ice-free regions of North Greenland (Dawes 2004), we assume that the erodability of the bedrock in North Greenland is on the low end of $k_e$ estimates (Garcia-Castellanos & O’Connor 2018). Taking the isostatically compensated channel slope ($0.3 \times 10^{-3}$; Bamber et al. 2013) and estimating 100 meters water elevation above sill height prior to an outburst flood (Figure 5.3B,E), we predict spillway erosion rates ranging from 0.5 to 5 meters per year in response to outburst flooding. Based on this model, we would expect erosion rates to grow over time as the canyon develops and the channel steepens.

5.5 Discussion

Though PISM is a complex model capable of realistically simulating the dynamics of modern ice sheets (i.e. Aschwanden et al. 2016) we use it here not to capture reality but as a tool to explore feedbacks in the climate-ice sheet-bedrock system following initial glaciation of the GIS. Our simulations point to an erosive mechanism that has been recognized for the formation of other large proglacial canyon networks (Larsen and Lamb, 2016) but not yet considered for the landscape beneath the GIS. Thus, although it is not possible to say when the canyon may have formed, our results provide a framework for understanding the formation of the canyon via linked geomorphological, climatic, and glaciological processes and provide a new hypothesis that can be tested by additional modelling and targeted field studies both beneath, and along the margin margins of, the GIS.
5.5.1 Mega-canyon formation via repeated proglacial lake outburst floods

Following the Last Glacial Maximum (LGM), the meltwater lakes left by retreating ice sheets in the Northern Hemisphere led to catastrophic floods that fundamentally altered surrounding landscapes (Larsen and Lamb, 2016; Teller et al., 2002; Clarke et al., 2004). These discharge estimates for these events range from 0.3 to 5.2 Sv. During the Holocene, erosive outburst floods in glacial systems have estimated minimum discharge rates of $10^{-3}–10^{-2}$ Sv (Baynes et al., 2015; Walder and Costa, 1996; Grinsted et al., 2017).

Our simulations predict that during periods of rapid ice-sheet retreat following glacial maxima, subaerial meltwater lakes are periodically formed in northern Greenland with a volume up to 18,000 km$^3$, twice the volume of Lake Agassiz prior to the Younger Dryas (Teller et al., 2002). Water routing analysis indicates that these lakes would have drained either to the northwest, following Petermann canyon, or to the northeast via the outlets of NEGIS (Figure 5.3). If drained completely in one year, this corresponds to a discharge of ~0.6 Sv, similar to discharge estimates for LGM scenarios but higher than those seen during the Holocene.

Our back-of-the-envelope calculations suggest that for the rock types found in North Greenland (lithified sandstones) erosion rates in response to these outburst floods would be 0.5–5 m/yr. The maximum depth of the Petermann canyon is ~800 meters, which places 160–1600 years of flood-style erosion by outburst flooding. This is consistent with the jökulhaup model for catastrophic ice-dammed floods, which suggests that Lake Missoula went through many fill-drain cycles during the last deglaciation (Waitt 1985). The flood volumes we reconstruct correspond to ~0.05 m of sea level. 
equivalent (s.l.e.), during periods when the ice sheet lost at least ~8 m s.l.e. in volume (Figure 5.2). These outburst floods would have happened late in the deglaciation when North Greenland was mostly ice free, but this mass balance analysis suggests that a single deglaciation would have provided enough melt during this phase to support up to 10s of outburst flood events. Thus, our back of the envelope 160–1600 years of flood-style erosion could be accomplished over more than one and up to tens of interglacials.

Observations of modern catastrophic floods provide another constraint on the total erosion we would expect from these floods. The estimated discharge for the floods we propose (0.6 Sv) is two orders of magnitude higher than that catastrophic, erosive floods observed in the historical record. Wilson et al. (2019) looked at a 2015 glacial lake outburst flood (GLOF) in Chile with peak discharge of $5 \times 10^{-3}$ Sv and resulted in a maximum of 38 meters of erosion in a sediment-laden alluvial plain. Cook et al. examined GLOFs in the Himalaya with a peak discharge of $2.3 \times 10^{-3}$ Sv which caused 1–10 meters of erosion into sediment. Finally, Lamb & Fonstad (2010) studied an outburst flood in Texas with a peak discharge of $1.4 \times 10^{-3}$ Sv which caused 10s of meters of incision in a bedrock canyon. While the discharge estimates for these cases are low compared to what we estimate, all of the studies looked at erosion over smaller spatial scales (1–10 km) than Petermann canyon (~750 km). If we assume that the higher discharge of the floods we reconstruct had a similar erosive potential over a larger spatial scale, with 1–20 meters of erosion per flood event, we require 40–800 flood events, consistent with the physically motivated estimate of canyon incision occurring over multiple interglacials.
5.5.2 Distribution of sediments around and beneath Greenland

Sediments both off- and onshore are a potential source of information about past outburst floods. Seismic studies have mapped extensive Plio-Pleistocene sediment packages north of Greenland which may reflect periods of enhanced terrestrial erosion. In northern Baffin Bay, high-resolution seismic surveys reveal thick Pliocene- and Pleistocene-age sediment deposits (Gregersen et al., 2013; Knutz et al., 2015). These sediments can be reasonably well dated thanks to regional correlation of seismic reflectors and ODP ocean drilling site 645 (Knutz et al., 2015, 2019). Off Northeast Greenland, multiple sediment cores and good coverage by seismic studies permit more complete mapping of sedimentary units (Berger and Jokat, 2009). In the Boreas (77°—79°N) and Molloy (79°—80°N) basins, sedimentation rates tripled from 3.4 to 10.3 cm kyr\(^{-1}\) during the Pliocene, falling to 7.3 cm kyr\(^{-1}\) during the late Pliocene and Pleistocene (Berger and Jokat, 2009). One explanation for this change in sedimentation rate is that early glaciations in Greenland were more erosive, consistent with the mechanism we propose.

Sediments beneath the ice sheet itself also support our proposed mechanism. Receiver function analysis to identify the structure of the mantle beneath Greenland required \(\sim\)100m of sediment beneath the NGRIP ice-core site to fit the observed seismic returns there, in contrast to other ice-core sites which did not require a sediment package to fit the returns (Dahl-Jensen et al., 2003). In northeast Greenland, a ground-based active seismic study mapped a package of dilatant till \(\sim\)40 km wide and \(>8\) m thick beneath the main trunk of NEGIS (Christianson et al., 2014). While none of these data prove the
existence of repeated outburst flood events, they are collectively consistent with our proposed mechanism of landform evolution.

5.5.3 Comparison with marine records of ice-rafted debris

IRD has been used to identify periods of enhanced iceberg discharge from northern hemisphere ice sheets during the LGM (Flesche Kleiven et al., 2002) (Figure 5.1). Iceberg discharge during the LGM was closely associated with the release of large volumes of meltwater; meltwater slows the melting of icebergs and facilitates the transport of entrapped clasts to the open ocean. IRD could also be an indicator of past meltwater floods originating from Greenland. IRD are sourced from all over the northern hemisphere, but their origin can sometimes be clarified by their geochemistry (Bailey et al., 2012). During the late Pliocene and early Pleistocene, such studies indicate that Greenland was an important source of icebergs to the North Atlantic (Naafs et al., 2013).

Marine sediment cores from around Greenland have been analyzed for IRD. In Baffin Bay, studies of ODP Site 645 reveal persistent deposition of IRD and larger presumably iceberg-derived material throughout the Plio-Pleistocene. In the early Pliocene, 10% of the sediments consist of beds up to 1-meter-thick attributed to rapid emplacement of poorly sorted sandy sediment (Hiscott et al., 1989b). These were interpreted as either debris flow or sudden accumulations of ice-rafted sediments; the possibility they were flood deposits was not considered (Hiscott et al., 1989b). Differentiating ice-rafting and meltwater pulse deposits is difficult, especially in ice-proximal settings (i.e. Baffin Bay); indeed, sedimentary evidence of Heinrich events during the LGM indicates that both meltwater and iceberg discharge occurred during these events (Hemming, 2004). The source of IRD indicates that most are derived from Baffin
Island and West Greenland (Korstgaard and Nielsen, 1989). However, the presence of quartzarenite clasts in the Pleistocene section of the core points directly to a source in northern Greenland, the Thule Basin (Hiscott et al., 1989a), suggesting at least some of the material was sourced from the region we propose contributed to outburst flooding (Figure 5.3).

5.5.4 Other examples of canyon formation via repeated lake outburst floods

In the Channeled Scablands of western Washington state, repeated outbursts of glacial lake Missoula carved canyons kilometers across and several hundred meters deep. The estimated discharge rate during the formation of these features is $0.1$ to $3 \times 10^6$ m$^3$ s$^{-1}$ or $0.1$ to $3$ Sverdrups (Sv) (Larsen and Lamb, 2016). Although the routing of other meltwater floods associated with the retreating Laurentide ice sheet is less certain (e.g. Condron and Winsor, 2012), the volume of proglacial Lake Agassiz as estimated from shorelines is 9500 km$^3$ preceding the Younger Dryas, and 163,000 km$^3$ preceding the 8.2ka event; if released over a period of one year (as it often assumed in modelling studies), these volumes correspond, respectively, to rates of 0.3 and 5.2 Sv (Teller et al., 2002; Clarke et al., 2004).

5.5.5 Implications for Plio-Pleistocene abrupt climate change

These outburst floods are a potential driver of abrupt climate change through influence on ocean circulation before glaciation of the northern hemisphere reached magnitudes similar to those observed during the LGM. The potential for relatively low-volume yet potent outburst floods may be easier to integrate with evidence for abrupt climate changes during periods when sea-level variations are minimal.
Many studies have linked iceberg discharge and meltwater floods to rapid and severe changes in global climate during the LGM (Alley et al., 1993). Though older climate archives generally lack the resolution to study similarly abrupt climate changes, some authors have argued that abrupt climate changes also occurred during past glacial periods in the Plio-Pleistocene (Naafs et al., 2013). These are often attributed to Dansgaard-Oeschger-style oscillations in the subpolar North Atlantic. Although the exact mechanism for these cycles is unknown, it is likely that variations in freshwater input to the North Atlantic played a role. For instance, Bolton et al. (2010) showed that stable isotope measurements on planktonic foraminifera from the subpolar North Atlantic reveal high-frequency variability in surface conditions there during the early Pleistocene (2.6 — 2.4 Ma). This is in keeping with results from the Mediterranean Sea during MIS 100 (~2.5 Ma), where sea surface conditions responded in sync with variations in ice-rafted debris in the North Atlantic (Becker et al., 2006). Thus, it is clear that abrupt climate changes influenced global climate prior to the LGM, yet potential mechanisms for this variability are lacking.

The impact of meltwater outburst floods on global climate is highly dependent on their routing. Condron and Winsor (2012) used a high-resolution ocean circulation model to study the impact of a 5 Sv meltwater pulse sustained over one year, meant to represent a Lake Agassiz outburst flood around the time of the Younger Dryas. They found that routing the water through the St. Lawrence Valley (into the North Atlantic) had little impact on the strength of the AMOC circulation, while water routed through the Mackenzie Valley (into the Arctic Ocean) resulted in a weakened AMOC. The size of the meltwater pulse indicated by our modelling is an order of magnitude smaller than the one
used in that study, which is unsurprising given the size of the GIS relative to the LGM Laurentide; however, the northerly outlets where the water is routed are ideally located to be directly entrapped in the East Greenland Current and advected to the North Atlantic with minimal dispersion (Figure 5.1). Studies of the potential impact of these floods have not been performed despite their obvious potential importance for AMOC and global climate (Ganopolski and Rahmstorf, 2001).

5.6 Conclusions

Our results suggest a mechanism for the formation of an extensive geomorphological feature that exerts a first-order control on ice dynamics in northwest Greenland. Based on a combination of numerical modeling and Plio-Pleistocene sediment accumulation patterns, we argue that repeated retreat and advance of the GIS during its Pliocene/Pleistocene inception may have led to large meltwater outburst floods. The potential for these floods is set by interactions between the climate, expanding ice-sheet and pre-glacial bedrock topography. Thus, despite uncertainty in the timing of such events, the possibility exists and is consistent with several lines of independent evidence.

Large meltwater lakes in North Greenland are consistent with accumulation of sediments in these regions, which still underlie the NGRIP (Dahl-Jensen et al., 2003) and NEGIS (Christianson et al., 2014) ice-core sites. These sediments are dilatant and unconsolidated (Christianson et al., 2014), consistent with accumulation in a subaerial system. Because the presence of sediment beneath ice sheets is fundamental to their dynamics on both long and short timescales (e.g. Clark and Pollard, 1998; Alley et al., 2007), the origin and extent of such sediments should be mapped and, where possible, sampled.
Thick Pleistocene sediment packages around Greenland, especially individual deposits containing material sourced from north Greenland in the Late Pliocene (Hiscott et al., 1989b), point to sustained production and export of sediments from northwest and northeast Greenland throughout the Pleistocene (Berger and Jokat, 2009). Our simulations suggest that coarse-grain layers in these deposits may be characteristic of floods originating from proglacial lakes left during retreats of GIS. Future work on marine sediment cores from around Greenland may be enriched by considering the potential for such events. If their occurrence can be confirmed with physical evidence, it will alter our understanding of linkages between landscape denudation and ice-sheet evolution in Greenland. Such floods present a plausible scenario for the formation of the largest mapped subglacial feature in Greenland and one of earth’s most spectacular fluvial structures.
Figure 5.1 Map of Greenland. Ocean Drilling Program (ODP) sites mentioned in the text (645, 646, and 909) are annotated. Gray shading shows ocean bathymetry. Colors indicate bed elevation beneath the modern Greenland Ice Sheet. Red and purple outlines indicate the maximum extent of proglacial lakes in Figure 5.3. Deep ice core sites are labelled with filled dots. ODP sites 645, 646 and 909 are labelled with colored squares. Brown line reflects a conservative estimate of Last Glacial Maximum (LGM) ice sheet extent (Funder et al., 2011). Red box indicates the extent of the maps in Figure 5.3. Blue and orange arrows indicate the path of the (cold) East Greenland and the (warm) Irminger Currents, respectively, following (Straneo et al., 2012).
Figure 5.2 Climate evolution and ice volume change for Pleistocene simulations. The simulation shown in Figure 5.3 is plotted with the darkest, thickest line in each panel. Two other realizations with different climate forcings are shown with thinner medium- and light-hued lines. A) Applied temperature anomaly, meant to mimic three known aspects of Pleistocene climate change: 40 kyr cyclicity, gradually intensifying glacial climates, and sporadic "super-interglacials." B) Greenland Ice volume. C) Volume of the proglacial lake below sea level (black line). The time slices shown in Figure 5.3 are highlighted with red and purple boxes.
Figure 5.3 Development and evolution of a large proglacial lake in Greenland initiation experiments. Colors, same as in Figure 5.1, indicate isostatically adjusted topography, white transparent shading shows the extent of the ice sheet, and turquoise lines and polygons show the calculated water routing paths within the ablation zone and the proglacial lake assuming it was filled up to the sill. The hydropotential flowline in the top row was used to extract the profiles plotted in the bottom two rows. The first row shows the extent of the ice sheet just prior to deglaciation. The second and third rows shows the evolution of the extent of the proglacial lake during two deglaciation phases (A–C and D–F). Note that subtle differences in topography dictate that lake outflow is sometimes routed to the northeast (panel F). Panels A–F represent 680, 687, 690, 1000, 1004, and 1008 kyr simulated years.
CHAPTER 6

CONTROLS ON THE PLEISTOCENE STABILITY OF THE GREENLAND ICE SHEET

6.1 Abstract

The future stability of the Greenland ice sheet will have direct and grave consequences for coastal communities globally. Understanding thresholds in future ice sheet stability hinges on our ability to interpret the geologic record of past collapses, which is equivocal about the timing and magnitude of past collapses. Here we use an ice sheet model forced by climatologies developed using two continuous proxy records of Arctic climate evolution during the last 800,000 years, one based on North Atlantic sea surface temperatures and the other based on a terrestrial climate reconstruction from Lake El’gygytgyn. When tuned to sea-level targets for the last 125,000 years, our results show divergent histories of ice cover which can be used to generate testable hypotheses about the exact timing and duration of ice cover across Greenland. Reconstructions from Lake El’gygytgyn are the most consistent with sea level reconstructions from the last 125,000 years and suggest prolonged periods of nearly ice-free conditions in Greenland for much of the last 800,000 years in response to Arctic interglacial warmth.

6.2 Introduction

The Greenland ice sheet (GrIS), if melted completely, would raise global sea level by 7.36 meters (Morlighem et al. 2017). The future behavior of the GrIS critically depends on the magnitude of anthropogenic warming in the coming centuries (Pattyn et
Understanding the thresholds that determine whether the ice sheet will remain relatively stable or undergo substantial retreat is difficult, because ice sheet models differ in the predictions of how the GrIS will respond (Fürst et al. 2015, Vizcaino et al. 2015).

The geologic record provides analogues that can be used to understand the response of the GrIS to past warm periods and thereby constrain its response to future warming. However, inconsistencies between different geological archives lead to divergent and mutually inconsistent interpretations of GrIS stability in the recent past. Some lines of evidence indicate that the ice sheet has been generally stable over the last three million years (e.g. DeConto et al. 2008, Alley et al. 2010). Even during the pronounced warmth of previous interglacials, many authors have argued for minor and/or geographically restricted changes to GrIS volume. Interglacial ice preserved at the base of the NEEM ice core in Northwest Greenland indicates that during the Eemian (Marine Isotope Stage (MIS) 5e, 125 thousand years ago (ka)) the ice sheet remained extensive and continuous despite an >8°C warming at that site (NEEM Community Members 2013). Reyes (2014) argued that during MIS 11, while a portion of Southern GrIS diminished, the ice sheet lost no more than 30-40% of its volume. Argon isotopes in the basal ice of Southern GrIS reinforce this interpretation and demonstrate that some basal ice there is as old as 400 ± 170 ka (Yau et al. 2016). The same approach yielded a basal age of at least 970 ± 140 ka for ice beneath the summit of the ice sheet (Yau et al. 2016). These findings are in accord with an analysis of the basal material from the GISP2 ice core, which found that site had been continuously covered by ice for the last 2.7 Ma (Figure 6.1, Bierman et al. 2014). These results, which argue for the stability of the GrIS
in more-or-less its present configuration for the duration of the Pleistocene, are unsurprising in the context of previous modelling results. DeConto et al. (2008) demonstrated the sensitivity of GrIS glaciation to atmospheric CO$_2$ levels, which have been near or below 280 ppm deglaciation threshold for at least the past 3 million years (Bereiter et al 2015, Martinez-Boti et al. 2015). However, newly emergent results are questioning the long-term stability of the ice sheet. Schaefer et al. (2016) analyzed cosmogenic nuclide concentrations in the bedrock below the GrIS summit and argue that this location must have been ice-free for some 25% of the last million years. These data are difficult to reconcile with the consensus that the GrIS has been stable over the same time frame.

Three-dimensional numerical ice sheet models can be used to examine the plausibility of various paleo-ice-sheet retreat scenarios, which can reveal, for example, the sensitivity of different regions of the ice sheet to various forcings and help refine predictions of future mass loss (e.g. DeConto and Pollard 2016). Due to positive feedbacks between ice sheet growth and snow accumulation, it has been difficult for ice sheet models to reproduce the magnitude of retreat inferred from geological records under known climate forcing scenarios. The Pleistocene provides a useful analogue for understanding GrIS response to future warmth, as global boundary conditions have remained the same for past “interglacial” periods. The rhythmic oscillations of atmospheric CO$_2$ and Antarctic temperature between glacial and interglacial periods over the past 800,000 years are also apparent in the Arctic, but with a much greater amplitude of warming during some so-called “super-interglacials” (Brigham-Grette et al. 2013). That the Arctic is exceptionally sensitive to interglacial warmth has long been apparent.
(Alley et al. 2010), but the community has lacked quantitative insight into the magnitude of Arctic climate change during much of the Pleistocene is lacking, due to the repeated scouring of the landscape by ice-age glaciers and ice sheets.

Here we use for the first time a temperature reconstruction from Lake El’gygytgyn, Arctic Northeast Siberia (Habicht et al., in prep) to force an ice-sheet model to understand the implications of past Arctic warmth for the stability of the GrIS over the last 800 kyr (Figure 6.2). This approach allows us to test competing hypotheses about the ice sheet’s response to past warm periods and refine our understanding of how the GrIS might respond to future atmospheric warming.

6.3 Methods

We use a three-dimensional thermomechanical hybrid ice-sheet/ice-shelf model to test the hypothesis that the Greenland ice sheet remained stable in response to Arctic climate change over the last 800 kyr. We initialize the model by adjusting basal sliding coefficients to fit the modern geometry of the ice sheet (Pollard and DeConto 2012). Our model realistically captures both the modern ice volume and dynamics of the GrIS under modern climate forcing (Figure 6.3).

We use a regional climate model (RegCM3), adapted to Greenland, to generate the climatologies used to force the ice-sheet model. The model performs favorably relative to reanalysis products when run with modern boundary conditions (Figure 6.3). We perform simulations using the following ice-sheet configurations: LGM, modern, Holocene Thermal Maximum, and partially deglaciated. The boundary conditions for each experiment can be found in Table 6.1.
Time-continuous climatologies are produced by weighting between the four RCM climatologies, based on three different continuous paleoclimatic timeseries: the Lisiecki-Raymo stack, an alkenone-based SST reconstruction from IODP site 982 (57.5°N, 15.86°W), and a branched glycerol dialkyl glycerol tetraether (brGDGT) record from Lake El’gygytgyn (67.5°N, 172°E) (Lisiecki & Raymo 2005, Lawrence et al. 2009, Habicht et al. in prep, respectively). The Lisiecki-Raymo stack (LR04) is based on a compilation of globally distributed benthic oxygen isotope records and reflects a combined signal of changes in ice volume and global temperature over the last 5 Ma (Lisiecki & Raymo 2005). The alkenone record from IODP site 982 reflects changes in sea surface temperature in the North Atlantic region (Rockall Plateau) over the last 4 Ma. The site is well situated to record changes in the size of the Greenland and its influence on regional climate (Lawrence et al. 2009). The brGDGT record from Lake El’gygytgyn is the first continuous terrestrial temperature reconstruction from the Arctic over the time interval 800ka – present (Habicht et al., in prep). The record reveals that the terrestrial Arctic exhibited a particularly strong response to so-called “super-interglacials” of the last 800 ka, including MIS 11 and MIS 7. Though Lake El’gygytgyn is located on the other side of the Arctic (Figure 6.1), a comparison between the brGDGT record and the NGRIP ice core oxygen isotope record, commonly interpreted as primarily reflecting temperature, reveals that Lake El’gygytgyn captured the same major patterns of climatic variability during this interval (Figure 6.2).

We use the three proxy reconstructions described above (Figure 6.2) to linearly interpolate between glacial, modern, and warmer-than-modern climate scenarios. The proxy records were normalized so that their value at LGM is 0 and their value at modern
in 1; thus, any periods that are colder than LGM have a value lower than 0 and any
periods warmer than modern have a value higher than 1 (Figure 6.4). As the primary
uncertainty in these climate reconstructions is related to the climatology during past
warm periods, two adjustable parameters were included in the interpolation scheme
(Figure 6.4). We generated ensembles by varying these parameter values within a
prescribed range to create unique climatologies, representing a range of possible climate
histories consistent with each proxy reconstruction (Figure 6.4). Each climatology was
first used to drive to a continuous 125 kyr integration of the ice sheet model. Although
the climatology is thus created off-line from the ice sheet model, we use a lapse rate
correction to determine temperature (5 °C km⁻¹, Abe-Ouchi et al. 2007) and precipitation
decreases exponentially as a function of elevation (Ritz et al. 2001) in each grid box
when calculating surface mass balance. Simulations were scored based on their ability to
meet a suite of sea level targets for this period (Table 6.1); the simulations which met
these criteria were then re-run for 800ka to understand the implications of these long-
term climatologies for the Pleistocene stability of the ice sheet.

To compare the timing and extent of past exposure in simulations against
measurements, we directly calculate the cosmogenic nuclide inventory of bedrock. We
use published elevation-dependent production rates and decay constants to calculate the
concentration of \(^{10}\)Be and \(^{26}\)Al using the GIA-adjusted elevation of exposed bedrock in
each grid cell (Balco et al. 2008, Stone et al. 2000). Erosion is assumed to be minimal
over the last 800,000 years, which is consistent with observations from central Greenland
(Schaefer et al. 2016, Bierman et al. 2014), although this assumption may be an
oversimplification for the margins (Sinclair et al. 2016, Egholm et al. 2017).
We use four sets of sea-level targets for the last 125 ka based on published estimates of ice-volume and/or far-field sea-level change to guide our selection of model parameters for long-term integrations. These targets and the associated references are shown in Table 6.1. Although we do not use MIS 11 (420–395 ka) to guide parameter selection, estimates of GrIS ice volume for this interval are included for reference, as this is one of the most well-studied intervals of sea-level change during the last 800 ka.

### 6.4 Results

Simulations forced by LR04 show little variability in ice sheet volume regardless of parameter choice (Figure 6.5). The early Holocene sea level target is not capture by any of the simulations because the Holocene Thermal Maximum, a well-described aspect of early Holocene Arctic climate, is not captured by LR04. Because none of the simulations described here met our sea level targets, we focus on the results of simulations forced by IODP Site 982 and Lake El’gygytgyn (Figure 6.5).

A subset of simulations forced by the record from IODP Site 982 meet all the prescribed sea level targets. Because of the high index values during the latter part of the Eemian, many of the simulations indicate a near total disintegration of the Greenland ice sheet around 110 and 100 ka. Following the Eemian retreat, ice volume recovers into the last glacial period before dropping during the Holocene thermal maximum and rising again towards modern values (Figure 6.5).

A subset of simulations forced by the record from Lake El’gygytgyn meet all the prescribed sea level targets. Ice loss during the Eemian is constrained to the period 129–119ka before rising into the last glacial period. Ice volume falls in response to the Holocene thermal maximum before rising towards modern values.
We re-ran simulations that met our sea level targets for the last 125 ka over the last 800 ka. The Site 982- and Lake El’gygytgyn-driven climatologies result in markedly different ice sheet histories over this interval (Table 6.3, Figure 6.6). Site 982 simulations indicate a relatively stable ice sheet from 800 to 430 ka, with the first major disintegration occurring during MIS 11 (Table 6.3, Figure 6.6). The ice sheet responds less strongly to MIS 9 and MIS 7, and again disintegrates completely in the Eemian, slightly after the documented MIS 5e highstand (Kopp et al. 2009). Lake El’gygytgyn simulations experience a greater duration of low ice volume (Figure 6.6). The first major disintegration of the ice sheet occurs during MIS 19, and partial to near-total collapses occur during MIS 15, 13, 11, 9, and 7 (Table 6.3, Figure 6.6). In contrast, the ice sheet response to MIS 5e is modest, with less than 2 m slet of ice lost in response peak Eemian warmth.

6.5 Discussion

Cosmogenic nuclides (i.e. $^{10}$Be and $^{26}$Al) in bedrock samples and ice-marginal environments can provide insights into the ice cover history of currently glaciated regions (Schaefer et al. 2016, Bierman et al. 2016, Shakun et al. 2018). These radioactive nuclides accumulate in minerals through spallation by cosmogenic rays which attenuate rapidly with depth in most media (Balco et al. 2008). Their concentrations in surface rocks reflect an integrated signal of the total amount of time that the bedrock has been ice free and exposed. For a simple exposure scenario where previously shielded rocks have been exposed until the present day, determining the date when sample was exposed is straightforward (i.e. Sinclair et al. 2016). For samples that have undergone multiple periods of exposure and coverage, interpreting the exposure history is more difficult.
Multiple exposure cycles can bias exposure ages, but pairing bedrock and erratic boulder samples permits simultaneous calculation of both the simple exposure age and the inheritance of nuclides in the bedrock from previous exposures (i.e. Corbett et al. 2013). On their own, bedrock samples may have a complicated history that cannot be directly constrained by isotope measurements alone. For example, though ice-marginal areas are likely to be eroded and their cosmogenic signature “reset,” areas below central Greenland may have experienced less erosion and therefore retain more of the cosmogenic isotope signature. The first samples measured from beneath the GrIS indicate there has been little erosion over the last 3 Ma, and the concentration of $^{10}$Be predicted by the data from the bedrock just beneath the GISP2 site was approximately $6 \times 10^4$ atoms per gram (Schaefer et al. 2016).

The upper limit for the last time GISP2 was exposed is $\sim$2.5 Ma, longer than the simulations presented here. Our simulations can be used to identify targets for future drilling and generate testable hypotheses about Greenland ice cover over the last 1 Ma. For example, the Site 982 and El’gygytgyn simulations differ not just in the concentration of nuclides accumulated across Greenland, but also in the ratio of $^{26}$Al to $^{10}$Be. These nuclides are measurable in the same samples (i.e. Schaefer et al. 2016), and because they are produced at a characteristic ratio but decay with different half-lives, the ratio in measured samples can reveal the last time that the sample location was ice-free for an extended duration. Our results identify multiple locations where variations in this ratio could be directly linked to provide support for either of the ice sheet histories presented here or, in the case of ratios < 5, rule them out completely (Figure 6.7).
Another way to utilize cosmogenic isotopes is to measure their concentrations in ice-proximal archives, which integrate exposure over an entire catchment but may have more precise ages as they can be dated using other archives in the same core and are more likely to be stratigraphically continuous. $^{10}$Be records from the east coast of Greenland suggest that the southeast may have been more dynamic than the northeast over the last million years, but these reconstructions are limited by coarse temporal resolution (Bierman et al. 2016). Such records may provide more useful constraints at higher resolution or when simulations covering the last ca. 5 Ma are available.

The timing of past ice loss in these simulations depends on the age models of the proxy records used to generate the ice-driving climatologies. The age model for the LR04 stack is based on a simple ice sheet model which follows northern hemisphere summer insolation (Lisiecki and Raymo 2005, Imbrie and Imbrie 1980). The age model for Site 982 was generated by tuning to the LR04 stack (Lawrence et al. 2009), and the age model for Lake El’gygytgyn was additionally tuned to insolation parameters (Nowaczyk et al. 2013). Similar tuning procedures were done for all of the published records and discrepancies in the assigned age of individual interglacials are minor.

The two regions which tend to deglaciate first are the sectors of the ice sheet in southwest and northeast Greenland. Examining the climatologies reveals that these regions are sensitive for different reasons. In southwest Greenland, temperature increases above modern drive the already extensive ablation zone up the relatively modest surface slopes, resulting in rapid increases in surface melting as the climate warms (Figure 6.3). In the northeast, the extremely low precipitation rates cannot keep pace with even small
changes in surface melting, causing this sector to be sensitive to the lapse-rate-driven positive feedback on surface melting.

In this study we have considered only the role of atmospheric forcing in driving GrIS volume over the last 800ka. Today atmospheric warming is the primary driver of changes in Greenland mass balance (Fürst et al. 2015), though changes in oceanic heat transport have driven increased discharge at some outlets (Straneo et al. 2010) and may lead to additional dynamics changes in the future (Fürst et al. 2015). As the purpose of this study was to understand GrIS response to past warm periods, we chose to leave ocean temperatures at modern values, which restricts the maximum size of the ice sheet during glacial periods.

The GrIS interglacial response can critically depend on the size of the ice sheet during the preceding glacial maximum, as the isostatic depression associated with increased ice loading can increase the efficacy of ocean warming and drive mass loss. However, as the GrIS melts, its margin becomes increasingly terrestrial, with marine terminating glaciers confined to deep trough outlets in few locations. Thus, the role of oceanic forcing in driving ice-sheet collapse is confined temporally (early in the deglaciation) and spatially (specific fjords). In addition, there are few records appropriate to reconstruct the behavior of relevant water masses (e.g. warm Irminger Current waters) over the timescale considered here. We therefore leave simulations which consider the combined influence of atmospheric and oceanic temperature change to future work, and note that regardless of the oceanic forcing used, we expect the differences in ice sheet response shown here to remain stark even under different ocean forcing scenarios.
Forcing the ice sheet model with the deep-ocean benthic stack and two terrestrial temperature records leads to divergent ice histories over the last 800 ka (Figure 6.6). This result is important in the context of efforts to understand the connection between globally integrated records from the deep ocean and the changes in surface climatology that led to the growth and decay of large ice sheets during the Pleistocene. Previous authors have speculated that glacial periods with stronger cooling of the deep ocean are associated with less buildup of terrestrial ice (Rohling et al. 2015), which is required from an isotopic mass balance perspective (Winnick and Caves, 2015) but stands in contrast to the commonly used technique of forcing an ice sheet model with the LR04 stack. This suggests there is some decoupling between the deep ocean and terrestrial temperatures on glacial-interglacial timescales which is not simply related to the fact that the atmosphere is variable on a much shorter timescale than the deep ocean. Terrestrial records which independently constrain past surface climate are necessary to improve our understanding of past ice sheet change, and as demonstrated here, imply markedly different histories for the ice sheets. Even changes in surface ocean temperature can differ from terrestrial temperature; the record from Site 982 has a strong obliquity component which may be responding directly to changes in earth’s orbit rather than through all feedbacks (i.e. changes in albedo associated with ice and vegetation) which directly impact the terrestrial climate response and lead to a stronger response to precession forcing (Habicht et al. in prep).

Using the Lake El’gygytgyn record to force the ice sheet model leads to modest ice loss during the Eemian, in contrast the Site 982 record, which shows near-complete retreat the ice sheet ~10ka after the highstand (Kopp et al. 2009). Because ice dating to at
least 90 ka is still present across much of Greenland (Macgregor et al. 2016), and basal ages of ice cores are older than the Eemian (Yau et al. 2016, Bierman et al. 2014, Willerslev et al. 2007), it is unlikely that the ice sheet lost much of its mass during the Eemian. Thus, given the current state of our knowledge of the history of the ice sheet, forcing the ice sheet model with the Lake El’gygytgyn temperature record is most consistent with the available data, and we favor this reconstruction for understanding the longer-term history of the ice sheet prior to the ice sheet. This implies that Greenland may have contributed significantly to both known highstands of the last 800 ka (i.e. MIS 11) and other interglacial periods for which highstands are not well-documented (i.e. MIS 7). Records of this response may have been erased or altered by subsequent highstands (Dutton & Lambeck 2012) or erroneously attributed to more “prominent” interglacials due to inaccurate dating (Burgess et al. 2019). Alternatively, Greenland may have responded to super-interglacial Arctic warmth that was not felt by the Antarctic ice sheet, which may have compensated for Arctic ice loss through increasing surface mass balance (e.g. Huybrechts and Oerlemans 1990), leading to a negligible global sea level response (Raymo et al. 2006). It is worth noting that continuous reconstructions of sea level based on ice sheet models (DeConto and Pollard 2008) and marginal sea hydraulic models for the Red and Mediterranean seas (Grant et al. 2014, Rohling et al. 2013) suggest a sea level highstand during MIS 7 that was of similar magnitude to the MIS 5e highstand, which is compatible with our results when forcing the ice sheet model with the Lake El’gygytgyn-based reconstruction.

Both the Lake El’gygytgyn and Site 982 based reconstructions suggest ~20 ka of exposure at the GISP2 site during MIS 11. MIS 11 is characterized by a long interval of
moderate interglacial warmth in most records, which resulted in a sea-level highstand mapped at many far-field locations (Dutton et al. 2015 and references therein). Although there remains debate about the age of the ice at the base of the GISP2 ice core, and whether or not that ice is \textit{in situ} (Yau et al. 2016), our results show that multiple interpretations of the climatic history of Greenland which match sea-level targets for the last 125 ka suggest a prolonged period of ice loss during MIS 11 and a contribution of the Greenland ice sheet to that highstand of \(~6\) m (Table 6.3).

\section*{6.6 Conclusions}

The history of the Greenland ice sheet remains uncertain. In this study we used proxy records of deep ocean temperature, North Atlantic sea surface temperature, and terrestrial Arctic climate to force a model of the Greenland ice sheet for the last 800,000 years. Our results show that these climate records imply different stability histories for the ice sheets and allow us to generate testable hypotheses about the spatiotemporal stability of the Greenland ice sheet over the Pleistocene which can be directly tested by future drilling. Our results demonstrate the sensitivity of the ice sheet to climate change during the Pleistocene and underscore the need for better records of high-latitude temperature change and associated ice-sheet response.
<table>
<thead>
<tr>
<th>Time period</th>
<th>Age</th>
<th>Estimate of Greenland ice volume (m sle)</th>
<th>Sources</th>
</tr>
</thead>
<tbody>
<tr>
<td>Holocene thermal maximum</td>
<td>7-4 ka</td>
<td>7.01 [6.66–7.36]</td>
<td>Young and Briner 2015, Larsen et al. 2017</td>
</tr>
<tr>
<td>Eemian</td>
<td>129-123 ka</td>
<td>5.21 [4.06–6.36]</td>
<td>Dutton et al. 2016 and references therein</td>
</tr>
</tbody>
</table>

*Table 6.1* Sea level targets for simulations.
<table>
<thead>
<tr>
<th>Experiment</th>
<th>Orbit</th>
<th>Greenhouse gases</th>
<th>GrIS configuration</th>
</tr>
</thead>
<tbody>
<tr>
<td>Last Glacial Maximum</td>
<td>21 ka</td>
<td>190 ppmv CO2, 367 ppm CH4</td>
<td>expanded</td>
</tr>
<tr>
<td>Modern</td>
<td>0 ka</td>
<td>400 ppmv CO2, 1800 ppm CH4</td>
<td>modern</td>
</tr>
<tr>
<td>Holocene Thermal Maximum</td>
<td>7 ka</td>
<td>257 ppmv CO2, 591 ppm CH4</td>
<td>modern</td>
</tr>
<tr>
<td>Partially deglaciated</td>
<td>same as modern</td>
<td>same as modern</td>
<td>partially deglaciated</td>
</tr>
</tbody>
</table>

**Table 6.2** Boundary conditions for climate model experiments. RCM boundary conditions include orbital parameters (Berger et al. 2014), CO2 (Bereiter et al. 2015), CH4 (Schilt et al. 2010), NO2 and past GrIS configurations (expanded: Argus et al. 2014, Peltier et al. 2015; modern: Morlighem et al. 2017; partially deglaciated, this study).
<table>
<thead>
<tr>
<th>MIS</th>
<th>982</th>
<th>El’g</th>
</tr>
</thead>
<tbody>
<tr>
<td>19</td>
<td>9.2</td>
<td>6.85</td>
</tr>
<tr>
<td>17</td>
<td>7.51</td>
<td>0.88</td>
</tr>
<tr>
<td>15</td>
<td>6.7</td>
<td>0.61</td>
</tr>
<tr>
<td>13</td>
<td>9.5</td>
<td>1.81</td>
</tr>
<tr>
<td>11</td>
<td>0.89</td>
<td>0.52</td>
</tr>
<tr>
<td>9</td>
<td>2.3</td>
<td>0.98</td>
</tr>
<tr>
<td>7</td>
<td>5.34</td>
<td>0.75</td>
</tr>
<tr>
<td>5</td>
<td>0.45</td>
<td>5.43</td>
</tr>
<tr>
<td>1</td>
<td>7.2</td>
<td>7.37</td>
</tr>
</tbody>
</table>

**Table 6.3** Greenland ice volume for interglacial Marine Isotope Stages (MIS) of the last 800ka for simulations forced by proxy reconstructions from IODP Site 982 (982; Lawrence et al. 2009) and Lake El’gygytgyn (El’g; Habicht et al., in prep).
Figure 6.1 Map of the Arctic. The proxy reconstructions used in this study come from IODP Site 982 (green dot) and Lake El’gygytgyn (orange dot). The purple star marks the position of the Greenland Ice Sheet Project 2 (GISP2) ice core.
Figure 6.2 Comparison of climate records for the last 800 ka. a) Benthic Oxygen Isotope Stack (Lisiecki and Raymo 2005). b) Site 982 $\text{U}_\text{K}^{37}$-based sea surface temperature reconstruction. (Lawrence et al. 2009). c) Lake El’gygytgyn brGDGT-based terrestrial temperature reconstruction (Habicht et al., in prep). d) Oxygen isotope composition of ice from the NGRIP ice core site for the last ~130ka (NGRIP members 2004). e) Carbon dioxide concentrations from Antarctica over the last 800ka (Bereiter et al. 2015).
Figure 6.3 Comparison of modern climatology with reanalysis and past climates created with the regional climate model RegCM3 for temperature (left) and precipitation (right). Boundary conditions for all the simulations presented here are in Table 2. a) RACMO 2.3 (Noël et al. 2015). RCM Climatologies: b) Modern. c) Last Glacial Maximum (21 ka). d) Holocene Thermal Maximum (7 ka). e) Partially deglaciated.
Figure 6.4 Schematic of interpolation scheme used to scale between climate states over the last 800,000 years. Two changeable parameters determine the linear scaling between modern, Holocene Thermal Maximum, and partially deglaciated scenarios. Black lines are for visualization purposes and illustrate potential ensemble members analyzed in this study.
Figure 6.5 Ensemble performance over the last 125,000 years. Simulations forced by a) Lisiecki-Raymo benthic oxygen isotope stack (Lisiecki and Raymo, 2005). b) Site 982 Sea Surface Temperature reconstruction (Lawrence et al. 2009). c) Lake El’gygytgyn (Habicht et al., in prep). Colored boxes show the sea level targets in Table 6.1.
Figure 6.6 Results of 800,000-year ice sheet model integrations, forced by LR04 (black line), Site 982 SSTs (green line), and Lake El’gygytgyn terrestrial temperature (orange line).
Figure 6.7 Accumulation of cosmogenic isotopes in Greenland bedrock. $^{10}$Be concentration (top row) and predicted $^{26}$Al/$^{10}$Be (bottom row) for simulations forced by Lake El’gygytgyn temperature reconstruction (left column) and Site 982 SSTs (right column).
CHAPTER 7

OUTLOOK AND FUTURE DIRECTIONS

The chapters in this thesis answer some questions about the history of the Greenland ice sheet (GrIS) and its relation to past climate change and identify new avenues for fruitful research.

In Chapters 3 and 4, we identified the factors that drove deglaciation of the Greenland ice sheet margin following the Last Glacial maximum. In Chapter 3, we showed that seasonality and abrupt climate change affect the timing of deglaciation around the GrIS margins. In Chapter 4, we showed that atmospheric warming and oceanic processes drove the deglaciation in different sectors of the ice sheet. Dynamic processes also likely affected the evolution of the GrIS throughout the deglaciation, and that is a topic that should be explored in future research. The major outlets of the modern GrIS, especially the marine-terminating Jakobshavn, Petermann, and Northeast Greenland (Zakariæ, Storstrømmen, and Nioghalvfjerdsfjord) ice streams show significant dynamic changes in our simulations throughout the deglaciation. Multiple moraine complexes both on and off-shore of Jakobshavn indicate the outlet had a complex deglaciation history (Ó Cofaigh et al. 2013). Sediment cores from beneath Petermann suggest multiple collapses of the buttressing ice tongue there, potentially driven by increased surface melting and/or incursions of warm water masses during the Holocene (Reilly et al. 2017). These records should be compared with modelled fluctuations of Petermann in response to atmospheric and oceanic forcing in order to elucidate the mechanisms of past ice loss in northwest Greenland and understand the
controls on the stability of this system, which overlies one of the longest and deepest subglacial canyons in the world, into the future (Morlighem et al. 2014). In central Northeast Greenland, changes in the basal state of the ice sheet and reduced ice accumulation during the LGM result in the ice sheet freezing to the bed, reducing the capacity of the ice sheet to deliver mass to the ocean via its northeastern outlets. This may explain the retreat of northeast outlets far behind their present-day margin in the early Holocene (Larsen et al. 2017. As all of these ice streams are undergoing remarkable modern changes and have a large potential for contributing to future sea level rise, changes in dynamics during the last major period the GrIS lost mass merit closer inspection. In addition, fluctuations of land-terminating portions of the GrIS during the last deglaciation are captured by our simulations and may explain the complex exposure histories recorded in both cosmogenic and threshold-lake records (e.g. Larsen et al. 2011).

Past studies have used relative sea level indicators to constrain the evolution of the ice sheet (e.g. Simpson et al. 2009, Lecavalier et al. 2014). These markers are problematic because their ages mostly rely on $^{14}$C analysis, which is particularly complicated in North Atlantic marine sediments because of the highly uncertain reservoir age correction, which probably changed through the deglaciation, and uncertainty in solid earth structure beneath Greenland. The improvement of solid earth models has made comparison of relative sea level records and ice-sheet model output more robust, resulting in an improved model fit to data in Antarctica (e.g. Gomez et al. 2018). Ongoing work to couple more realistic solid earth models with ice-sheet models coupled both directly during computation and in post-processing, will make relative sea level
indicators more robust targets for paleo ice sheet models and should be incorporated into future work to constraining the ice cover history of Greenland.

The early Holocene is emerging as a powerful analogue for future warming, because it is the most recent period when the ice sheet was smaller than it is today, and a relatively wide coverage of data are available to evaluate model performance (Briner et al. 2017). Ice core records from Northern Greenland show an increased number of melt layers during the early Holocene, in good agreement with reconstructed summer temperatures there up to +4°C above modern (Lecavalier et al. 2017, Lasher et al. 2017). The impact of this increased surface melting on the GrIS is less clear and should be examined in future work. The GrIS may have responded to increased melt through changes in ice dynamics, and some data indicates widespread glacier acceleration in response to early Holocene warmth (MacGregor et al. 2015). If this acceleration is related to increased sliding, it might be explained by increased production of surface meltwater and/or increased access of surface meltwater to the subglacial environment, a phenomenon which may also explain the widespread and enigmatic basal ice structures seen throughout Northern Greenland and commonly attributed to basal freeze-on (Panton & Karlsson 2015, Bell et al. 2014). This hypothesis could be directly tested by analyzing existing model output described in this dissertation. In southeast Greenland, the restricted geometry of many fjords makes their dynamics difficult to capture at the coarse resolution employed in our studies. Further work using nested high-resolution grids to capture these features will be necessary to fully understand their response to the deglaciation.
In Chapter 5, we showed the potential for large proglacial lake outburst floods to influence landscape evolution in north-central Greenland. We also showed that our results are consistent with limited datasets from offshore west Greenland, which indicate northern-sourced sediment delivery to Baffin Bay during the Pliocene. This work has been used to motivate a marine drilling proposal which was reviewed favorably and awaits scheduling at the JOIDES Resolution Review Board. This proposal, led by Dr. Paul Knutz of the Geological Society of Denmark and Greenland (GEUS), seeks to provide spatiotemporal constraints on the evolution of the western GrIS, including its inception and fluctuations throughout the Pleistocene. If the proposed records are developed, they can also be used to provide a circum-Greenland perspective on the extent and timing of past exposure and erosion along the Greenland margin which will complement existing studies (e.g. Bierman et al. 2016). Such records will be useful for evaluating the veracity of long-term ice cover histories like those presented in Chapter 6.

In Chapter 6, we showed that proxy records of Arctic temperature change over the last 800,000 years have dramatically different implications for the past stability of the ice sheet. This work demonstrates the potential for sub-ice cosmogenic nuclide records to constrain GrIS stability over longer timescales. Ongoing work to produce similar records from other regions will provide spatiotemporal “fingerprints” of past ice-sheet collapse which will provide direct and powerful new modelling targets. Future work to take advantage of these records have the potential to shape our understanding of the coupled evolution of Arctic climate and the GrIS over the last few million years.
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