NUMERICAL CLIMATE MODEL SIMULATIONS INVESTIGATING THE ROLE OF ARCTIC SEA ICE EXPORT EVENTS IN MODULATING DEGLACIAL CLIMATE

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NUMERICAL CLIMATE MODEL SIMULATIONS INVESTIGATING THE ROLE OF ARCTIC SEA ICE EXPORT EVENTS IN MODULATING DEGLACIAL CLIMATE

A Dissertation Presented

by

ANTHONY J. JOYCE

Submitted to the Graduate School of the University of Massachusetts Amherst in partial fulfillment of the requirements for the degree of

DOCTOR OF PHILOSOPHY

September 2019

Geoscience
NUMERICAL CLIMATE MODEL SIMULATIONS INVESTIGATING THE ROLE OF ARCTIC SEA ICE EXPORT EVENTS IN MODULATING DEGLACIAL CLIMATE

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Department of Geosciences
DEDICATION

To my mother, Eileen.
ACKNOWLEDGMENTS

I’d like to acknowledge and thank all the individuals who have been supporting me in the process of my doctoral work and for being part of an exciting journey.

First and foremost, I would like to thank my advisor, Alan Condron for providing me with his invaluable guidance and patience throughout my research. He truly inspired me on many different levels and excited my curiosity about climate modeling. Professionally, he helped me acquire the necessary skills and tools to analyze data in many fashions and gave me the freedom to explore and apply my thoughts to this project. Thank You!

Raymond Bradley is thanked for sharing his knowledge about pre- and post-LGM climate conditions. He has proven to be an indispensable resource in conducting this research and creating thought experiments to enhance my research. I also greatly appreciate his support in testing this hypothesis. I’d also like to thank Brian Rose for explaining the fundamentals of sea ice algorithms in climate models, as well as providing excitement for everything modelling. A great thanks is also given to Rob DeConto for teaching me the fundamentals of modeling during my master’s degree as well as guiding me through my early stages of computer programming for climate data analysis. Thank You. Last but not least, I’d like to thank Hans Johnston for being on my committee and having so much curiosity regarding my project.

I’d like to thank my fiancé Briana. She has been my emotional support blanket for the last 4 years always being supportive during all my hardships. She has always provided faith, patience and love. She always understood no matter the issue. Finally, I’d like to
thank my Mom. She has always had as much interest in my endeavors as I had. Thank you Mom.
ABSTRACT

NUMERICAL CLIMATE MODEL SIMULATIONS INVESTIGATING THE ROLE OF ARCTIC SEA ICE EXPORT EVENTS IN MODULATING DEGLACIAL CLIMATE

SEPTEMBER 2019

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Directed by: Professor Alan Condron

Periods of abrupt climate cooling during the last deglaciation (20,000-8,000 yrs ago) are often attributed to glacial outburst floods slowing the Atlantic meridional overturning circulation (AMOC). Yet, despite over 40 years of research, conclusive evidence that such events significantly impact climate remains elusive. This study uses a climate model to investigate an alternative freshwater forcing mechanism in which the episodic break-up and mobilization of thick perennial Arctic sea ice might have disrupted large-scale climate. The results presented here show the first evidence that (1) the Arctic Ocean stored enormous volumes of freshwater during colder periods as thick masses of sea ice, and (2) that massive sea ice export events to the North Atlantic are generated whenever the transport of sea ice is enhanced either by changes in atmospheric circulation, rising sea level submerging the Bering land bridge, or glacial outburst floods draining into the Arctic Ocean from the Mackenzie River. Of relevance for understanding the key drivers of past abrupt climate change, I found that the volumes of freshwater released to the Nordic Seas are similar to, or larger than, those estimated to have come from terrestrial outburst floods, including a discharge around 12,900 years ago that is often considered the cause of the Younger Dryas cooling. The results from
my thesis thus provide the first evidence that the storage and release of Arctic sea ice helped modulate deglacial climate change.
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<tr>
<td>Acceleration</td>
<td>$Du/Dt$ Variable (m s$^{-1}$)</td>
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<td>Acceleration of gravity</td>
<td>$g$ 9.81 m s$^{-2}$</td>
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<td>Angular momentum</td>
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<td>Air density</td>
<td>$\rho_a$ 1.225 kg m$^{-3}$</td>
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1.1. Introduction

During the last deglaciation (~20,000-8,000 yrs. BP), periods of increased freshwater discharge to the North Atlantic and Arctic Ocean often coincided with the onset of centennial-to-millennial length periods of climate cooling (Clark et al., 2001) and impacted climate far beyond the Arctic Ocean (Knies et al., 2007; Spielhagen and Bauch, 2015). Major hemispheric and global warm-to-cold climatic transitions, such as Heinrich Events, the 8.2-kyr-event, the Pre-Boreal Oscillation (PBO), and the Younger Dryas have all been linked to an increase in terrestrial freshwater runoff to the North Atlantic (Bond and Lotti, 1995; Broecker, 1994; Broecker et al., 1989; Clark et al., 2001; Manabe and Stouffer, 1997). The last (i.e. most recent) of these cold events is known as the Younger Dryas (12,900-11,700 cal. yr. BP; Figure 1.1). This episode of cooling was originally identified by the reappearance of the Arctic-Alpine flowering plant, Dryas octopetala, in lake sediments in Scandinavia (Jensen, 1938) and is most clearly seen in climate records across the North Atlantic and western Europe (Emiliani, 1978; F. Ruddiman and McIntyre, 1981). In ice cores from Greenland, it is marked by a ~7°C cooling over a century and an even faster (~20-40 year) warming ~1,200 years later at its termination (Alley, 2000; Rasmussen et al., 2006). A reorganization of the major ocean currents, including the Gulf Stream and Atlantic Meridional Overturning Circulation (AMOC), in response to increased freshwater forcing from the sub-Arctic and Arctic regions, is commonly evoked as the main mechanism driving these events. Indeed, major changes in global ocean circulation have
been closely linked to periods of increased freshwater forcing (Clark et al., 2002). For example, dramatic reductions in North Atlantic Deep Water (NADW) formation has been shown through observations of lowered benthic δ¹³C near the Portuguese and Moroccan margins (Keigwin and Lehman, 1994; Vidal et al., 1997; Willamowski and Zahn, 2000; Zahn et al., 1997) and further supported by excess Protactinium/Thorium ratios from Bermuda Rise sediments (McManus et al., 2002).

Figure 1.1: Reconstructed Greenland temperatures derived from isotopic ratios (blue line) showing the main deglacial cold episodes: 8.2-kyr-event, Pre-Boreal Oscillation (PBO), Younger Dryas, Inter-Allerød cold period, and the Older Dryas (yellow bars). The Younger Dryas spans the interval ~12,900-11,700 cal. yr. BP. (Source: Condron & Winsor 2012).

The AMOC (Figure 1.2) is considered a vital component of the global climate system, as it regulates the global thermal heat budget by transporting ~1 Petawatt (PW; 1x10¹⁵ Watts) of heat northward from the tropics to the high northern latitudes. In the North Atlantic, warm, salty waters flowing northwards in the upper ocean from the tropics to the
higher latitudes cool in the Labrador and Norwegian Seas and sink to depth via a process known as deep convection, to form NADW (e.g. Cunningham et al., 2007; Srokosz et al., 2012). Increased freshwater runoff to the ocean has been shown in climate models to restrict deep convection by creating a buoyant, freshwater “cap” in the Labrador and Norwegian Seas that inhibits NADW formation and slows the transport of heat to the high latitudes. In response, the sub-Arctic, Western Europe and North America cool (Manabe and Stouffer, 1988).

Figure 1.2: The northern branch of the Atlantic Meridional Overturning Circulation (AMOC). Warm, salty waters in the upper layer of the Atlantic Ocean (red-orange-yellow arrows) flow northward, cool, and sink to depth to create southerly flowing deep water (blue-green arrows) (Source: NASA GISS)

Although freshwater input has been repeatedly cited as the major cause of abrupt climate change in the literature for the last three-to-four decades (Broecker et al., 1989;
the precise sensitivity of the AMOC to changes in high-latitude freshwater forcing is still poorly understood, and highly model dependent (Figure 1.2). Climate models generally show that ~0.1 Sv (1 Sv = 10^6 m^3 s^{-1}) of freshwater applied consistently to the North Atlantic between 50°N - 70°N latitude weakens the AMOC, while 1.0 Sv immediately ‘shuts off’ the overturning cell (Ganopolski and Rahmstorf, 2001; Manabe and Stouffer, 1997; Rahmstorf, 2002; Stouffer et al., 2006). However, the source of the freshwater thought to have disrupted the AMOC in the past remains poorly understood and widely debated (Broecker, 2006; Keigwin and Jones, 1995; Lowell et al., 2005; de Vernal et al., 1996). The mechanism commonly put forward centers on the idea that the overturning cell is weakened by a sudden, and massive freshwater discharge from glacial lakes situated along the southern edge of the Laurentide ice sheet, typically Lake Agassiz (Figure 1.3), or the rerouting of freshwater from one drainage outlet to another (see Section 1.2.1; Clarke et al., 2004; Lowell et al., 2005; Meissner and Clark, 2006).
Figure 1.3: Changes in the strength of the Atlantic Meridional Overturning circulation (AMOC), or Thermohaline Circulation, to a 0.1 Sv freshwater perturbation to the subpolar North Atlantic. The figure shows the significant difference in AMOC sensitivity to the same freshwater perturbation in 14 climate projection models. (Source: Stouffer et al. 2006).
Figure 1.4: The possible drainage pathways (blue arrows) that stored freshwater from post-glacial lakes may have taken initiating cold climate events during the last deglaciation (Source: Tarasov & Peltier 2005).

Studies have shown the AMOC might operate in two different modes; one that comprises of a colder ‘glacial’ state in which the overturning is shallower and weaker, and the second a more vigorous overturning similar to today. It has also been suggested that the two modes vary in sensitivity to freshwater input (e.g. Ganopolski and Rahmstorf, 2001) with a near modern “warm” phase of the AMOC being more sensitive than a glacial “cool” phase. The ability of NADW to form in different areas during warm and cold climate regimes may also lead to differences in the sensitivity of the climate system to changes in high latitude freshwater forcing, such as a southward migration of deepwater formation sites in the North Atlantic from the Greenland Sea to south of Iceland (Ganopolski and Rahmstorf, 2001).
1.2 High Latitude Freshwater Forcing

There are three main high-latitude freshwater forcing mechanisms commonly put forward as drivers for past abrupt climate change: (i) land-based meltwater from glacial lakes and ice sheet collapse (Broecker, 1994; Broecker et al., 1989), (ii) massive iceberg discharge events, commonly associated with Heinrich Events (Alley and MacAyeal, 1994; Heinrich, 1988), and (iii) changes in sea-ice cover in the North Atlantic and sea ice export events out of the Arctic Basin (Li et al., 2010; de Vernal et al., 1996). Here, I give an overview of each mechanism, and also discuss the role of freshwater forcing in Dansgaard-Oeschger (D-O) events as the mechanisms driving these millennial-length sawtooth shaped climate events remain the subject of considerable debate and uncertainty. It is also important to realize that the role of freshwater forcing in the global climate and ocean systems is by no means clear and remains a topic of much scrutiny and fierce debate. Yet, in order to make accurate future climate predictions, it is critical to understand how changes in freshwater forcing drove past climate change.

1.2.1 Meltwater Floods

Freshwater forcing produced by the melting of the major Northern Hemisphere ice sheets (e.g. Younger Dryas, 8.2-kyr-event) during deglaciation had frequently been cited as a mechanism capable of triggering past changes in climate by inhibiting deepwater convection and slowing the AMOC (Clark et al., 2001; Keigwin et al., 1991; Lowell et al., 2005; McManus et al., 2004; Meissner and Clark, 2006). Broecker et al. (1989) first suggested that the Younger Dryas cold episode (12,896 – 11,703 cal. yr. BP; Rasmussen et al., 2006), was triggered by a sudden re-routing of glacial Lake Agassiz overflow from the Mississippi drainage system to the Gulf of St. Lawrence weakening the AMOC (Figure
1.5). Around the same time, Manabe and Stouffer (1988) used a climate model to show that freshening the North Atlantic (between latitudes 50° N-70° N) led to a dramatic reduction in the strength of the AMOC, supporting Broecker’s idea that increased high-latitude freshwater runoff might have triggered past abrupt climate change. More recently, Condron and Winsor (2012) showed that meltwater released from the St. Lawrence Valley would not instantaneously cover the sub-polar North Atlantic with freshwater as Broecker suggested but would have turned to the right due to the Coriolis force, and flowed south along the east coast of North America as a buoyant coastal current (Condron and Winsor, 2011, 2012). As such, meltwater from the Gulf of the St. Lawrence seaway did not significantly influence NADW formation or the overturning circulation in these experiments. However, the authors did find that meltwater routed from the Mississippi River to the Arctic Ocean, via the Mackenzie River, significantly reduced the AMOC, leading them to suggest that a more northern meltwater outlet likely triggered the Younger Dryas.

Using a glacial systems model, Tarasov and Peltier (2005) also showed that during the Younger Dryas episode, glacial meltwater from the North American ice sheets may have drained into the Arctic Ocean via the Mackenzie River Valley, with a flux of 0.01-0.16 Sv over a 100-yr. period. Their model results are supported by terrestrial evidence on the Mackenzie delta of a massive flood event at the onset of the Younger Dryas (Murton et al., 2010) and more recent results from marine sediments recovered from near to the mouth of the Mackenzie River showing a freshwater signal at this time. In addition, it has been proposed that the freshwater released from the Mackenzie River might have flushed sea ice and freshwater stored in the Arctic Ocean into the North Atlantic through the Fram
Strait (e.g. Not and Hillaire-Marcel, 2012), increasing the impact of any terrestrial freshwater forcing on the AMOC.

![Proposed drainage pathways originating from Lake Agassiz](image)

Figure 1.5: Proposed drainage pathways originating from Lake Agassiz. Prior to the Younger Dryas, meltwater was routed primarily into the Gulf of Mexico via the Mississippi (D). The onset of the Younger Dryas cooling was originally thought to coincide with the opening of the eastern outlet (C) or Hudson Strait (B), although there is increasing evidence that the Arctic route (A) may have opened at this time (Source: Teller et al., 2002)

A second example of climate change attributed to massive freshwater discharges is Meltwater Pulse 1A (MWP – 1A), which occurred ~14,000 yr. BP and was associated with 14-18 m of global eustatic sea level rise over a ~350-year period (Deschamps et al., 2012). Although the global sea level records suggest there was an input of freshwater into the ocean, the source location and timing of the event are still debated (e.g. Carlson, 2009; Clark et al., 2002; Gregoire et al., 2012; Stanford et al., 2006; Weaver et al., 2003). For
example, isotope dating methods have also linked MWP-1A to the onset of the Older Dryas cold episode (Stanford et al., 2006) while other studies have linked this meltwater event to the initiation of the warm Bølling episode (Clark et al., 1996). As such, associating climatic changes in response to MWP-1A remain difficult to reconcile as a climatic warming in response to a large meltwater discharge is hard to explain with existing theory on how freshwater forcing impacts climate. In addition, studies using glacial-isostatic adjustment models differ on the source location of MWP-1A and suggest that the LIS was not the sole source and suggest that the Antarctic ice sheet may have played a considerable role (Clark et al., 2002; Tarasov et al., 2012). Ice sheet modeling studies show that if MWP-1A originated from North American ice sheets, a “saddle collapse” between the Laurentide and Cordilleran Ice Sheets could have discharged $3.8 \times 10^6$ km$^3$ yr$^{-1}$ (0.24 Sv) of freshwater causing ~9 m of sea level rise over 500 yrs. (Gregoire et al., 2012). The magnitude of this freshwater pulse is ~50-60% of the recorded sea level rise associated with this period, and the remaining ~40% could possibly be explained from melting European and/or Antarctic ice sheets (Gregoire et al., 2012). Other modeling studies hypothesize a drainage of the rapidly collapsing LIS into the Arctic, Gulf of Mexico or the Mid-Atlantic (Labrador Sea and Gulf of St. Lawrence) (Tarasov and Peltier, 2005) which could then weaken the AMOC. In any case, assuming MWP-1A was in response to the Bølling-Allerød warm interval (and triggered the Older Dryas cold period) a Northern Hemisphere source would certainly be feasible. On the contrary, if MWP-1A coincided with the initiation of the Bølling warm period it has been suggested that Antarctica could be a source of the freshwater as some models (Weaver et al., 2003) have shown that a freshwater flux from Antarctica would disrupt Antarctic Bottom Water (AABW) formation.
that triggers a compensating increase in NADW formation and increase in AMOC. However, substantiating these results has proven difficult (e.g. Ivanovic et al., 2017) as does terrestrial evidence for an Antarctic contribution (e.g. Liu et al., 2015).

Additional uncertainties surrounding the role of terrestrial freshwater forcing in causing abrupt climate change are also evident for the Younger Dryas. For example, at the onset of the Younger Dryas (~12,900 years BP), it has been estimated that ~9,500 km³ of freshwater were released from Lake Agassiz into the North Atlantic (Teller et al., 2002). This volume of freshwater is, however, ~15-20 times less than the volume of freshwater estimated to have been discharged into the North Atlantic at the onset of the 8.2-kyr-event (~8,200 years BP), yet the millennial length Younger Dryas cold episode lasted 10-times longer than the 8.2-kyr-event. Furthermore, a lack of any significant sea level rise (5±2 m) over the Younger Dryas period, or perhaps even a sea level lowering during this time, supports the notion that the flux of freshwater from land to the subpolar North Atlantic was probably quite small (e.g. Abdul et al., 2016; Lambeck et al., 2014).

1.2.2 Heinrich Events

Heinrich Events were first recognized in deep-sea sediment cores along the North Atlantic belt between 35-50°N (Ruddiman, 1977). Each event is identified as an abrupt transition to an interval of high ice rafted detritus (IRD) concentrations, high magnetic susceptibility (Grousset et al., 1993), and a reduction in foraminifera abundance (Bond et al., 1992; Grousset et al., 1993; Heinrich, 1988). It is believed that each occurrence is attributed to a massive release of icebergs into the North Atlantic, and there are six such events (H1 to H6) dated within the past 70 kyr (Bradley, 1999). IRD layers contain mainly quartz grains, but some layers exhibit a high abundance of limestone and dolomite
suggesting a similar origin (Dowdeswell et al., 1995). Layer thicknesses (avg. 10-15 cm) decrease from the west Atlantic to east Atlantic, suggesting material originated primarily from the LIS (probably Hudson Bay) (Broecker, 1994; Dowdeswell et al., 1995) and propagated eastwards with the Gulf Stream.

Figure 1.6: Time series of post-glacial sea level rise (meters) for the last deglacial. Colored points with error bars indicate sea level readings from various locations in both the Atlantic and Pacific Ocean. Sea level rise associated with the Younger Dryas (12,896-11,703 cal. yr. BP) is < 5.0 m globally, indicating very little to no sea level rise. (Source: Robert Rohde, Berkeley Earth)

Icebergs entering the sub-polar gyre from Canada would have freshened the North Atlantic and reduced NADW formation. Multiple studies show that during a typical freshwater-event, 0.02-0.08 Sv (3.0-12.0 x 10^{14} m^3) of freshwater may have been released over a 500-year interval into the sub-polar ocean (Roberts et al., 2014). Evidence of a reduction in sea surface salinities (Vernal et al., 2000) have been found in North Atlantic
marine sediment cores (Madureira et al., 1997) and ice cores from the Greenland ice sheet (Steffensen et al., 2008), whereas warming in the Southern Hemisphere can be seen in Antarctic ice cores, indicating a possible bi-polar seesaw effect (Blunier et al., 1998). Furthermore, changes in precipitation related to freshwater-events are recorded in marine sediment cores from Cariaco Basin (Peterson et al., 2000), lake sediments from Bolivia (Baker et al., 2001; Fritz et al., 2010; Placzek et al., 2013), and speleothems from Brazil (Wang et al., 2004), consistent with wetter conditions further south and drier conditions further north, which are thought to be related to a southward shift of the Intertropical Convergence Zone (ITCZ) during Northern Hemisphere cold events (Broccoli et al., 2006; Chiang, 2009; Chiang and Friedman, 2012). Changes in marine sediment cores from the Arabian Sea (Deplazes et al., 2013; Schulz et al., 1998) and in loess from China (Porter and Zhisheng, 1995) indicate a weakening of the Asian Monsoon, most likely coincident with changes in North Atlantic circulation.

The role of freshwater from melting icebergs in triggering climatic cooling is not however that obvious. Originally it was proposed that the melting icebergs slowed AMOC (e.g Broecker 1996), but more recent research has identified a significant (~1500 yr) lag between the onset of climatic cooling and the actual deposition of IRD (normally at the end of D-O events), suggesting that they may not be the direct cause of a slowdown in the strength of the AMOC and that perhaps they are the response to regional or hemispheric changes in climate and/or internal ice sheet dynamics.

1.2.3 Dansgaard-Oeschger Events

Dansgaard-Oeschger (D-O) Events (e.g. Dansgaard et al., 1993) are millennial length sawtooth shaped, rapid climate changes observed in the Greenland ice core oxygen
isotopic ($\delta^{18}O$) record. They are typically associated with a $10 \pm 5^\circ$C temperature rise in a few decades followed by a gradual cooling ($100’s$-$1000’s$ of yrs.; Bond et al., 1999). Heinrich events may be associated with D-O events as they typically precede D-O warmings, but not every D-O event has a Heinrich Event (Bond and Lotti, 1995). During a D-O event, warming in Greenland is coincident with warming in the Nordic Seas, wetter conditions in Europe (Genty et al., 2003) and South America (Peterson et al., 2000), and enhanced summer monsoons in the Indian Ocean, suggesting an intensification of the Northern Hemisphere hydrological cycle. Aridity has also been recorded in the southwestern United States (Wagner et al., 2010) indicating changes in moisture patterns were not identical everywhere, but rather regional in the sub-tropics and tropics. Isotopic studies of Greenland ice during D-O events suggest a southern moisture source from a region where sea surface temperatures (SST) were higher, indicating extensive warming (somewhere) in the Northern Hemisphere (Johnsen et al., 1989). Cooling in Antarctica (EPICA community members, 2006) has also been demonstrated in Southern Hemisphere ice cores coinciding with increasing temperatures in Greenland (Bradley, 1999; Lemieux-Dudon et al., 2010) demonstrating again, a bi-polar temperature regime.

The causes of D-O events remain uncertain, but theories about changes in ocean overturning, sea ice, and freshwater forcing have been put forward (Dansgaard et al., 1993; Dokken et al., 2013; Li et al., 2010). Unlike Heinrich Events, D-O events are thought to have been attributed to changes in sea ice cover over the Nordic Seas and northeast Atlantic and changes in the North Atlantic storm track (Dokken et al., 2013). During an interstadial (warm period), the northward displacement of the storm track mechanically mixes surface waters and brings latent heat into the Nordic Seas, which in turn inhibits sea ice formation.
Conversely, during stadials (cool periods), a more quiescent storm track is inferred with a southward displacement of the jet stream decreasing the amount of latent and sensible heat extracted from the North Atlantic, thereby making the environment more conducive to sea ice formation (Li et al., 2010; Li and Battisti, 2008). Generally, D-O events have been linked to changes in synoptic scale, cyclone propagation. These changes in the storm track help create D-O warm and cool events, leading to the saw tooth pattern evident in the Greenland ice core records.

In addition to the atmosphere, it is also believed that oscillatory modes of the AMOC (Manabe and Stouffer, 1988; Stommel, 1961) [described previously in section 1.1] produce warm stadials and cold interstadials in intermediate water masses of the Nordic Seas affecting sea ice formation and triggering D-O type events. If a decrease in freshwater input occurs between 50° N and 70° N (Nordic Seas) then NADW formation is enhanced, with an incursion of warm, salty waters accompanied by sea ice free conditions and warming over Greenland, producing a D-O-like event (warm phase of AMOC). Further evidence for this is examined in deuterium excess of the Greenland ice cores suggesting the Nordic Seas as a moisture source during Greenland interstadials due to an observed 3°C rapid decrease in source temperature of precipitation (Dokken et al., 2013; Steffensen et al., 2008) and sea ice free conditions. A decrease in sea ice would have increased precipitation over the Fennoscandian Ice Sheet (FSIS) and freshwater runoff into the Nordic Seas. This is thought to have increased sea ice formation, severing sensible heat flux from the ocean to the atmosphere and triggering the gradual cooling observed with D-O events. This cycle demonstrates that changes in both atmospheric and ocean circulation
exhibit control over freshwater forcing but determining which branch of the climate system changes first remains unresolved.

In summary, three major triggers of climate have been described that invoke changes in high latitude freshwater forcing from sea ice, land-based ice sheets, icebergs, and glacial lakes as a mechanism for weakening the AMOC. In this dissertation, I will test an alternative hypothesis to the classic glacial meltwater discharge idea by investigating whether the growth and export of thick, multiyear, sea ice out of the Arctic to the North Atlantic could have weaken the strength of the AMOC in the past. Unlike freshwater forcing from terrestrial sources, the melting of sea ice will not significantly increase global sea-level as the ice is already floating in the ocean. Furthermore, a sea ice thickness of just 10 m over the entire Arctic would store ~44,350 km$^3$ of freshwater and far exceed the volume of freshwater estimated to have been stored in glacial Lake Agassiz at the onset of the Younger Dryas.

This hypothesis will be tested by running a series of numerical climate model (MITgcm) experiments to answer three main questions:

1) What is the range in equilibrium Arctic sea ice thickness during the Last Glacial Maximum (LGM)? Could enough freshwater have been stored in the Arctic as ice, such that the eventual break-up of ice would have weakened AMOC by inhibiting NADW formation in the Nordic and Labrador Seas?

2) What mechanisms are capable of mobilizing thick, stagnant, multiyear Arctic sea ice and cause it to be exported to the North Atlantic?
3) How sensitive is the global ocean circulation, and more specifically the AMOC, to increased sea ice export from the Arctic to the subpolar North Atlantic?

1.3. Glacial and Modern Arctic Characteristics

During the Last Glacial Maximum (LGM; MIS 2; ~21,000 years BP), conditions might have been favorable for the thermodynamic growth of much thicker sea ice. The Barents and Scandinavian Ice sheets covered much of northwest Europe severing intrusion of North Atlantic warm water through the Barents Sea, and sea level ~120 m lower than today left the Bering Strait exposed, cutting off the connection to the north Pacific Ocean and leaving the Fram Strait as the only link between the Arctic and North Atlantic Oceans (Bradley and England, 2008; Landvik et al., 1998). As a result, this reduced the area of the Arctic by ~50%. The only oceanic heat exchange between the Arctic and the North Atlantic took place at the Fram Strait, creating an isolated Arctic basin. Additionally, a shift in atmospheric circulation permitting the jet stream to dip further south, essentially steering cyclones away from the Arctic Basin, could have subsequently reduced the mobilization of the confined ice and reduced atmospheric heat flux into the region (Bradley and England, 2008; Li and Battisti, 2008; Seager and Battisti, 2007). Combined with a low obliquity angle and a cold, relatively deep, halocline, conditions in the Arctic might have been ideal for sea ice to grow unabated to considerable thickness (10’s – 100’s m).

Proxy reconstructions of sea ice cover over the North Atlantic, based on surface ocean temperatures, suggest that during glacial conditions sea ice extended as far south as Iceland and the Faeroe Islands during winter with seasonally ice-free conditions in the eastern Nordic Seas (up to 80° N) in the summer as sea surface temperatures reached >3°C (Pflaumann et al., 2003). In the central Arctic, reconstructions of climate based on data
from the Greenland Ice Sheet indicate that temperatures were 25 °C lower than modern (Cuffey and Clow, 1997; Dahl-Jensen, 1998). Reconstructions from biomarkers and sediment deposition rates indicate that the entire central Arctic Ocean was covered by thick perennial sea ice at this time, while conditions in the western basin may have been cold enough for an ice shelf to form along parts of the north coast of Arctic Canada (Jakobsson, 2002; Nørgaard-Pedersen, 2003; Xiao et al., 2015). Sediment deposition rates in both the Eurasian and Ameriasian basin are <1 cm kyr⁻¹, i.e. low, suggesting very thick sea ice with few open leads for sediment to deposit to the ocean floor persisted during the summers (Figure 1.7). As the climate warmed, the mobilization and export of this thick sea ice into the North Atlantic might have supplied enough freshwater to the ocean to trigger a slowdown in the AMOC, although this has never been tested before.

In the modern Arctic, a cold halocline occupies the upper ~50 m of the water column and is partly maintained by meltwater runoff from the continents surrounding the Arctic Ocean. Below the halocline, warmer (~1-4 °C) and saltier (~34-35 psu) waters are encountered between ~200-700 m depth (Rudels, 2001) and are identified as the Atlantic Layer; a water mass originating from the North Atlantic that enters the Arctic in the West Spitsbergen Current (WSC) at Fram Strait (Rudels, 2001). The cold halocline acts to isolate the surface waters from the heat supplied by the Atlantic Layer and permits a thermodynamic equilibrium sea ice thickness of ~2–3 m (Rothrock et al., 1999; Vinje et al., 1998; Wadhams and Davis, 2000). Ice is initially formed by a freezing (congelation) process at the surface caused by heat loss to the atmosphere and is known as frazil ice (Walker and Wadhams, 1979). Less than 50% of this ice remains in the Arctic Ocean for more than 1 year as it is transported from the Laptev Sea and the East Siberian Sea to Fram
Strait by the Transpolar Drift (Dickson et al., 2007). Sea ice that stays in the Arctic for more than one year is considered multiyear sea ice and can grow to 5-8 m thick. Ice this thick has been observed off of Ellesmere Island and the northern coast of Greenland and is most likely attributed to land-fast ice or physical deformation processes such as ridging. Some observations have shown sea ice flows ~10-12 m thick with a crystalline structure, suggesting the thickest multiyear sea ice could have formed by slow thermodynamic congelation growth over several decades (Walker and Wadhams, 1979).

Figure 1.7: Sedimentation rates in the Amerasian and Eurasian Basins. Low sedimentation rates corresponding with low eustatic sea level illustrate oceanic conditions during the LGM. Sedimentation rates $< 1$ cm kyr$^{-1}$ suggest very thick sea ice with few leads for sediment to melt out. Additionally, this indicates a very rigid “lid” of ice cover over the Arctic that persisted during the summer. (Source: Jakobsson et al. 2003)
1.4. Evidence for exceptionally thick Arctic sea ice

In the early 16th century, Robert Thorne, an English merchant, advocated (to Henry VIII) exploring passages in the Arctic with a belief he could find an easier route to the Far East (Mills, 2003). It was thought, by some, that the central Arctic might be a large expanse of open water surrounded by sea ice in the northern parts of the subpolar North Atlantic making it easy to reach the Far East (Struzik, 2009). This “Open Polar Sea” originated from the Ancient Greeks who hypothesized that above a certain latitude, there was continuous sunlight for months keeping temperatures above freezing (Figure 1.8). Of course, such beliefs were soon proven to be very wrong.

Figure 1.8: Map picturing the “Open Polar Sea” theory of the early 18th and 19th century in which the Arctic Ocean was believed to be entirely free of sea ice. Drawn by Silas Bent, a U.S. Navy cartographer circa 1875. (Source: Silas Bent; public domain)
Many of the diaries and journals kept by 19th and early 20th century Arctic explorers refer to regions of exceptionally thick and extensive ice in the western Arctic ocean that was much thicker than in the modern-day (1970s to present). These features had a characteristically rounded, hummocky surface and stood out as being quite distinct from the disturbed and fractured multi-year sea-ice which the explorers generally had to deal with. They are frequently referred to as ‘Ice Islands’ as they are often found far from land and cover many hundreds of square kilometers. Such ice was documented by Nares (Nares, 1878) who introduced the term “paleocrystic ice”, to describe the exceptionally old and thick ice (“floes...of gigantic thickness with a most uneven surface and covered with deep snow...”) that he encountered off the northern coast of Ellesmere Island (Canada) (Figure 1.9). In fact, this “sea of ancient ice” (Alcock, 1876) – extending up to 480 km along the northern coast of Ellesmere Island into the Arctic Ocean– was traversed by A. Markham during his attempt to reach the North Pole in 1876 (Markham, 1880).

Other encounters with thick, very-old, sea ice was later documented north of Alaska by Mikkelsen (1907) and the Norwegian explorer Storker Storkerson who spent a summer ‘adrift’ on a large piece of sea ice ~15 m thick in the Beaufort Sea (Crary et al., 1955) (Figure 1.10). Furthermore, Cook (Cook, 1913) noted, "With our dogs bounding and tearing onward, from the 87th parallel to the 88th parallel we passed for 2 days over old ice without pressure lines or hummocks. There was no discernible line of demarcation to indicate separate fields, and it was quite impossible to determine whether we were on land or sea ice. The barometer indicated no perceptible elevation, but the ice had the hard, wavering surface of glacial ice with only superficial crevasses." He subsequently speculated as to the genesis of these features: "From my observations I had come to the
conclusion that ice does not freeze to a depth of more than 12 or 15ft [3.5 to 4.5 m] thick during a single year. Occasionally we crossed fields 50ft [15m] thick. These invariably showed signs of many years of surface upbuilding.... probably the result mostly of addition to the superstructure. Frequent falls of snow, combined with alternate melting and freezing in summer...are mainly responsible for the growth in the thickness of the ice on the Polar Sea..."

Figure 1.9: Thick, multiyear “paleocrystic” sea ice. This illustration appeared in the newspaper *The Graphic* in 1876 and depicts the Polar Sea laden with huge clasts to boulder size pieces of ice, as viewed by Commander Markham and Lieutenant Parr during their spring sledging expedition (*British Arctic Expedition, 1875-1876*). The ice in the foreground is estimated to be 15-18 m (50-60 ft.) high. A clear example of what Captain Nares named “Paleocrystic Ice” to describe very thick, multiyear ice they encountered in the western Arctic. (*Source: NASA; The Graphic*)

When American polar explorer Robert Peary (Peary, 1907) traversed the northern coast of Ellesmere Island (*Figure 1.12*) as far west as Cape Aldrich in 1909 he also described extensive ice shelves with “long, prairie-like swells” that shared many of the characteristics with modern-day ice shelves (*Figure 1.11*). Unlike the Antarctic, however,
the thick Arctic sea ice observed by these explorers was not glacially-fed. Instead these regions are formed by sea-ice that is repeatedly thickened by the accumulation of snow and superimposed ice on the surface, as well as the freezing of seawater onto the underside of the ice (Dowdeswell and Jeffries, 2017). Hence, the amount of ice growth in the vertical is limited by a combination of the amount of surface accumulation vs. ablation, and by the amount of heat supplied to the base of the ice by the ocean. Direct measurements (by drilling) and remote sensing showed that the ice shelves along the Canadian coast were able to become at least 35-50 m thick by this process. Indeed, radiocarbon dates suggest that parts of these ice shelves are several hundred-to-thousands of years old (Crary, 1958).

Figure 1.10: The photograph above was taken in the 1940’s by the United States Airforce (U.S.A.F) and was one of the last remaining extensive regions of thick (multiyear) sea ice in the Arctic (foreground). In the background, much thinner (single year) sea ice can be seen. Thick sea ice in the foreground was once used as floating runways during World War II and the Cold War by the United States. Norwegian Arctic explorer Storker Storkersen may have lived on a piece of sea ice like the type seen in the foreground. (Source: U.S.A.F.)
Figure 1.11: The region of thick ice along the coast of northern Ellesmere Island in July 2002. Today, these features are rapidly disintegrating, but there is a great deal of evidence that they were once much more extensive and were a feature along the northernmost Canadian coast of the Arctic for much of the late Holocene. In this image, the region of thick ice occupies the left two-thirds of the image and is covered by evenly spaced meltwater ponds, while thinner (single/multiyear) sea ice appears on the far-right side of the image. The coast of Ellesmere Island is just visible in the distance in the upper left corner. (Source: Dr. C. Braun, 2002).

Vincent et al. (2001) estimated from historical accounts that, at the end of the 19th century, these “ice shelves” occupied an area of ~8,900 km². Today, these formerly extensive features are rapidly disintegrating (Mueller et al., 2017), but there is a great deal of evidence that they were once much more extensive and have been a feature along the northernmost Canadian coast of the Arctic Basin (at least) for much of the late Holocene (England et al., 2017). Even today, there exists small relic ice shelves 40-60 m thick, still
persistent along parts of the northern coast of Ellesmere Island, Canada (Crary et al., 1955; Jeffries, 1992) (Figure 1.12).

Figure 1.12: The ice shelves along the northern coast of Ellesmere Island, Nanavut, Canada. The dashed dotted line indicates the extent of the “Ellesmere Island Ice Shelf” described by the Nares and Peary in the late 19th century. This ice shelf was described as having large, prairie like swells described by Peary in 1907. The black shading illustrates some of the last remaining large ice shelves in the Arctic today, such as the Ward Hunt and Ayles Ice Shelves. (Source: England et al. 2008)

We consider that these historical accounts from explorers reaching the Arctic at the end of the late Holocene “Little Ice Age” provide a glimpse of what conditions may have been like during the much colder conditions of the last ice age, when vast areas of the Arctic Ocean were likely covered by even thicker and more persistent ice. Here, we hypothesize that ice similar to this would have been much more prevalent during cold glacial-times and could have covered a significant portion of the Arctic. Then, as the climate warmed and this ice became unstable it might have been ‘flushed’ out of the Arctic.
into the North Atlantic, with the resultant freshwater forcing from the melting ice being large enough to weaken the AMOC.

1.5. Sea Ice Export as a trigger for Abrupt Climate Change

Over 40 years ago, Mercer (1969) hypothesized that the Arctic Ocean could have been covered entirely by thick glacial ice fed by the Fennoscandian and Laurentide ice sheets during the last ice age similar to that of the Antarctic Ice Sheet. As the climate warmed and the Arctic could no longer support an ice sheet, Mercer suggested a surge of ice from the Arctic to the Nordic Seas from the disintegration of these ice shelves may have caused a climatic shift similar to the Younger Dryas (~11 kyr. BP). Hughes et al. (1977) and Denton and Hughes (1981) further expanded on Mercer’s idea by hypothesizing the “Late Wurm Ice Sheet” behaved as a single dynamic system fed by ice streams from grounded continental ice sheets.

More recently, Jakobsson et al. (2013) presented a similar hypothesis suggesting a 1 km-thick Arctic Ocean ice sheet may have covered most of the Arctic basin during Marine Isotope Stage 6 (MIS; ~160-140 kyr. BP) (Figure 1.13). In addition, a numerical model study by Gasson et al., (2018) corroborates Jakobsson’s hypothesis suggesting that a 1000m thick ice shelf could have grown in the central Arctic Basin during MIS 6 grounding somewhere near Lomonosov Ridge. Furthermore, Moore (2005) speculated that prior to the onset of the Younger Dryas, the Arctic Ocean might have contained a significant number of large tabular icebergs, on the order of 200-400 m thick, that calved from ice shelves protruding from the northern edge of the Laurentide Ice Sheet. In all of these scenarios, however, the ice in question originates from land-based ice sheets, much like the ice shelves observed today over the Ross and Weddell Seas of Antarctica.
In 2008, Bradley and England (2008) hypothesized for the first time that the break-up of thick Arctic sea ice, (aka *paleocrystic ice*), rather than ice shelves or ice sheets, might be capable of weakening the strength of AMOC. As climate warmed and sea level rose during delgaciation, the Bering Strait re-opened and the Barents Sea ice sheet collapsed (*Figure 1.14a*). This change in geography allowed warm waters to enter the previously isolated Arctic Basin from both the Pacific Ocean and through the Barents Sea. The transport of water through the Laptev Sea, and the reestablishment of a Transpolar Drift, were hypothesized to have caused any thick, multiyear, sea ice to be exported into the Nordic Seas (*Figure 1.14b*). The authors went on to imagine that if the Arctic Ocean contained sea ice with an average thickness of \(\sim 50\) m, then \(\sim 10\) Sv of freshwater would be discharged to the North Atlantic if all of the ice were released in 1 year. By comparison, this is twice as much freshwater as is thought to have been discharged from Lake Agassiz at the onset of the 8.2-kyr-event and is much larger than any other meltwater floods emanating from land-based glacial lakes during the last deglaciation (Barber et al., 1999; Carlson, 2009; Clarke et al., 2004; Teller et al., 2002). The same volumetric flux, if released over a 100-year period (\(\sim 0.1\) Sv), is still adequate to weaken the AMOC for several centuries in many climate models (Manabe and Stouffer, 1997; Stouffer et al., 2006).

1.6. Summary

The majority of research over the last three-to-four decades focused on terrestrial freshwater sources as the cause of abrupt periods of climate cooling during the last deglaciation. To-date, very little research has considered the role of Arctic sea ice in triggering these events. Numerical climate models tell us that an increased freshwater discharge to the North Atlantic will have at least some impact on the AMOC, although the
response of the overturning to a particular input varies between models and with the location of the discharge (Roche et al., 2010; Stouffer et al., 2006). In the next chapter, I discuss the methodology used to test whether an increase in Arctic sea ice export can weaken the AMOC enough to create abrupt changes in climate similar to periods of climate cooling that occurred during the last deglaciation.

Figure 1.13: Ice sheet reconstructions of ice shelves during glacial conditions. (A) LGM ice-sheet reconstruction by Hughes et al. that cover the entire Arctic Ocean. (B) A schematic of the ice shelf proposed by Jakobsson et al. 2016 shown as a white shaded area off the northern coast of Canada. Yellow arrows represent published evidence of ice shelf grounding. The brown lines represent ice flow lines. The hatched areas represent the northern hemisphere ice sheets. Labels correspond to: AB is the Amerasian Basin, EB is the Eurasian Basin and LR is the Lomonosov Ridge. (Source: Jakobsson et al. 2016)
Figure 1.14: Schematic diagrams showing Arctic circulation and sea ice during A) full glacial conditions and B) deglaciation. It is hypothesized that a rise in sea level and deglaciation of the Barents Sea Ice Sheet allowed warm waters to re-enter the Arctic, resulting in thick ice being carried into the Nordic Seas toward key areas of NADW formation. Red arrows indicate ocean circulation from the North Atlantic. (Source: Bradley & England 2009).
CHAPTER 2

METHODS

2.1. Numerical Modeling of Ocean Circulation

The circulation of the ocean can be modeled mathematically as a three-dimensional fluid on a rotating sphere with the dynamical properties of the ocean driven by the surface radiation balance and momentum (wind stress) of the atmosphere and constrained by the continental landmasses that border the oceanic basins. At its most simplistic, ocean circulation occurs as a way to create a new equilibrium between two adjacent water masses whose properties differ in terms of temperature and salinity.

2.1.1. Numerical Modeling of a Fluid

The movement of a fluid parcel through its environment can be described by the equations of motion which are derived from Newton’s Second Law of motion. For simplicity, the equations of motion are initially described here in an inertial reference frame that is not accelerating or rotating. This states that the acceleration \( \frac{Du}{Dt} \) of an object of infinitesimally small dimensions \((\delta x \delta y \delta x)\) and mass \((\delta M = \rho \delta x \delta y \delta x)\), is the sum of all the forces, \( F \), acting upon the object,

\[
\rho \delta x \delta y \delta z \frac{Du}{Dt} = F
\]  

(2.1)
where $\rho$ (rho) is the density of sea water and $\mathbf{u}$ is the parcels velocity. In the expanded form (i.e. one that follows the same parcel around) $\frac{D\mathbf{u}}{Dt}$ is the total derivative describing a fluid’s velocity in the $u$ (zonal), $v$ (meridional) and $\omega$ (vertical) as:

$$\frac{D\mathbf{u}}{Dt} = \frac{\partial \mathbf{u}}{\partial t} + u \frac{\partial \mathbf{u}}{\partial x} + v \frac{\partial \mathbf{u}}{\partial y} + \omega \frac{\partial \mathbf{u}}{\partial z}$$ (2.2)

where $t$ is time. The three major forces acting on the parcel of fluid are gravity, the pressure gradient force and frictional forces. For gravity, this is described simply as:

$$\mathbf{F}_{\text{gravity}} = -g \rho \hat{z} \, \delta x \, \delta y \, \delta z$$ (2.3)

where $\hat{z}$ is the unit vector in the vertical direction and $g$ is the gravitational acceleration presumed constant at 9.8 m s$^{-2}$. Secondly, the pressure gradient force (PGF), created by the pressure force within the fluid, can be written as:

$$\mathbf{F}_{\text{pressure}} = -\nabla p \, \delta x \, \delta y \, \delta z$$ (2.4)

where $p$ is pressure. Here, the RHS of the equation states that the direction of acceleration is in the opposite direction of the largest increase in pressure so that a parcel will flow down gradient, from regions of high to low pressure. The $\nabla$ (del operator), indicates the gradient change of the pressure.
\[
\n\nabla = \left( \frac{\partial}{\partial x}, \frac{\partial}{\partial y}, \frac{\partial}{\partial z} \right)
\]

(2.5)

And, the final force acting on a fluid parcel is that from friction:

\[
F_{\text{friction}} = \rho F \delta x \delta y \delta z
\]

(2.6)

where \(F\) is the friction force per unit mass. Generally, frictional forces are negligible except near ocean boundaries where the fluid comes into contact with the sea floor and at the air-sea interface. The sum of these forces gives the Equation of Motion as:

\[
\rho \delta x \delta y \delta z \frac{Du}{Dt} = F_{\text{pressure}} + F_{\text{gravity}} + F_{\text{friction}}
\]

(2.7)

which, following substitutions from equations (2.3, 2.4, and 2.6), dividing through by the mass of the parcel (\(\rho \delta x \delta y \delta x\)), and rearranging gives the acceleration of a fluid parcel as:

\[
\frac{Du}{Dt} = -\frac{1}{\rho} \nabla p - g\hat{z} + F
\]

(2.8)

The equation considers a rotating reference frame where the Coriolis Force included, which leads to:
\[
\frac{D\mathbf{u}}{Dt} = -\frac{1}{\rho} \nabla p - 2\Omega \times \mathbf{u} - gz + \mathbf{F}
\]  

(2.9)

where \((-2\Omega \times \mathbf{u})\), is the Coriolis acceleration, and \(\Omega\) is the angular velocity \((7.2921 \times 10^{-5}\text{ rad s}^{-1})\). The Coriolis force is unitless and describes a fluid’s tendency to turn to the right (left) in the Northern (Southern) Hemisphere. Finally, to further calculate ocean circulation, it is important to consider the conservation of mass (i.e. that mass is conserved) and the law of thermodynamics, to describe the thermodynamic state in which the motion takes place.

### 2.2. Numerical modeling

To allow a time evolving state of ocean circulation to be derived, 3-dimensional numerical models are used to calculate the equations of motion at multiple locations at a prescribed time frequency. Solving them otherwise would be extremely difficult and time consuming. In this thesis, the equations of motion were solved using the Massachusetts Institute of Technology (MIT) General Circulation Model (MITgcm; Marshall et al., 1997). Like many numerical models, computational performance is increased by assuming that vertical motion due to acceleration and friction are negligible and can be described by the hydrostatic balance.

The majority of numerical experiments I discuss were performed on a global model domain projected onto a cube-sphere grid (Adcroft et al., 2004; Condron and Winsor, 2011, 2012) that evenly projects the Earth’s sphere onto a six-faced cube such that a relatively even grid spacing throughout the globe is permitted and polar singularities avoided (Figure 2.1). The ocean grid has a mean horizontal spacing of 2.8° (280-km) with 15 vertical levels,
ranging in thickness from 50m near the surface to approximately 690m at the maximum model depth. In particular, the MITgcm has been shown to accurately capture the vertical structure of the water column in the Arctic very well, including the physical properties of the cold upper halocline and intrusion of warm Atlantic intermediate water (Condron et al., 2009) making it well suited for our proposed Arctic sea ice experiments. Additional simulations studying the impact of meltwater outburst floods on Arctic sea ice export were also performed at an eddy-permitting (1/6°; 18-km) spatial resolution using a limited-area Arctic/North Atlantic grid configuration with 460 by 400 grid points in the north-south and east-west direction. Here, open boundaries at ~50° N in the North Atlantic provide ocean velocity, salinity, and temperature, and coincide with grid cells in a high-resolution global configuration (Condron et al., 2009; Hill and Condron, 2014).

Figure 2.1: The MITgcm cube-sphere (32x32) grid overlain bathymetry and topography (left) and the high-resolution global grid simulation (460x400) (right) with sea ice, land ice and ocean temperatures. The red arrow in the high-resolution grid (right) indicates the position of the Mackenzie River delta, where significant meltwater is thought to have entered the ocean during deglaciation. The white-grey shading is land-based ice whereas
the pure white shading in the ocean represents areas covered by sea ice. Finally, the red-blue colors represent sea surface temperatures, from warm to cool.

2.3. Sea Ice Model

To perform the numerical experiments integrated as part of this thesis, the MITgcm ocean model was coupled to the thermodynamic sea ice model of Winton, (2000) which simulates ice growth and melt using a three-layer, enthalpy conserving scheme capable of parameterizing brine pockets. The internal sea ice stresses were calculated using a non-linear viscous-plastic (VP) rheology with an elliptic yield curve (Hibler, 1979). This scheme is suitable for simulating the rheology of ice that grows to several tens-of-meters thick over the Arctic basin in our model. By using a VP rheology, the plastic behavior of sea ice can be maintained without being too computationally expensive (Hibler, 1979). The top of the simulated sea ice employs a zero-heat-capacity snow layer with a fixed temperature at the snow-ice interface, whereas below, there are two equally thick ice layers: an upper variable heat capacity layer with brine pockets parameterized and a lower fixed heat capacity layer (Bitz and Lipscomb, 1999; Winton, 2000). At each time step, the temperature of the upper layer is determined by the heat flux from the underlying layer, based on a conductive coupling, while the lower (i.e. basal) ice layer temperature considers the heat flux at the ice-ocean interface, termed $F_b$ in Figure 2.2. While values of $F_b$ are close to ~2.0 Wm$^{-2}$ for modern-day, values would have been much lower during the LGM.
Figure 2.2: A schematic showing the three-layer thermodynamic sea ice model of Winton (2000). A brine pocket is located in the upper ice layer to absorb and release insolation energy (I) during refreezing. At the base of the ice, the heat flux from the ocean to the ice ($F_b$) is not prescribed (as is common in many sea ice models), but instead calculated based on the temperature of the ocean surface in the model and the temperature at the base of the ice, the latter being fixed at the sea-water freezing point, $T_f$. Typical values for $F_b$ in the Arctic performed under glacial boundary conditions are found to be $<0.40 \text{ Wm}^{-2}$.

Sea-ice motion is driven by the stress on the ice from the wind, ocean, Coriolis force and the horizontal surface elevation gradient of the ocean (i.e. sea-surface height), as well as from internal stresses, as described in Losch et al., (2010) and Zhang and Hibler, (1997), and described as:
\[
\frac{\text{Du}}{\text{Dt}} = -mfk \times \mathbf{u} + \tau_a + \tau_w - mg \nabla h + F
\]

where \( m \) is the ice mass per unit area, \( \frac{\text{Du}}{\text{Dt}} \) is the acceleration, \( -mfk \times \mathbf{u} \) is the Coriolis force where \( f \) is the Coriolis parameter \((2\Omega \sin \theta)\), \( \tau_a + \tau_w \) are the atmospheric wind and oceanic stresses, respectively, \( -mg \nabla h \) is the mass flux associated with the slope of the sea surface height \((h)\) from pressure loading and \( F \) is the frictional force due to internal ice stresses resolved by the viscous-plastic rheology. Atmospheric wind stress \((\tau_a)\) is calculated by estimating surface wind from the geostrophic wind using the drag equation, and the oceanic \((\tau_w)\) stress is calculated from the oceanic currents below the sea ice (Hibler, 1979). When ice becomes several hundreds of meters thick (i.e. 10-fold thicker than the ice we simulate here), the horizontal flow of the ice due to internal deformation becomes significant (Goodman and Pierrehumbert, 2003; Pollard, 2005); this is not considered in our sea ice model however as the ice thickness never exceeded 80m over the Arctic basin.

Finally, the sea ice model allows for the differentiation between leads (open water) and very thick sea ice associated with divergence and convergence (ridging) of sea ice, while four sea ice albedo categories (wet/frozen snow-covered ice and wet/frozen bare ice) are used to simulate the different ice states (Perovich, 2002). An Ice Thickness Distribution formula (ITD) is used to calculate the probability of ice cover in a particular grid cell with thickness \( h_i \) (Thorndike et al., 1975) due to deformation and redistribution of ice. When air temperatures are below \( 0^\circ \text{C} \), liquid precipitation falling on the ice as snow accumulates to allow ice to grow in the vertical. This mechanism is especially important in the central
Arctic during the LGM for building thick masses of sea ice as temperatures were consistently below freezing.

2.4. Testing the hypothesis: Configuration and Key Experiments

A series of numerical Global Climate Model (GCM) simulations were integrated to test whether the export of significant volumes of sea ice from the Arctic to the subpolar North Atlantic might have weakened the strength of the AMOC during deglaciation. Particular emphasis was placed on i) determining the range of Arctic sea ice thickness during deglaciation, ii) understanding the main mechanisms that could have mobilized and exported thick, stagnant, multiyear Arctic sea ice out of the Arctic into the North Atlantic, and iii) quantifying the sensitivity of global ocean circulation to increased sea ice export from the Arctic to the subpolar North Atlantic.

The numerical model was configured to simulate LGM conditions by prescribing atmospheric forcing (radiation, humidity, precipitation, and wind speed) from monthly climatological output from the National Center for Atmospheric Research (NCAR) LGM Community Climate System Model (CCSM4) integration (Brady et al., 2013); sea level was set 60m lower than modern-day and the Northern Hemisphere ice sheets were prescribed at full glacial extent (Peltier, 2004) (Figure 2.3, 2.4). As a result, the only oceanic connection between the Arctic and the subpolar North Atlantic was at Fram Strait as the lower sea-level creates a land bridge between Siberia and North America and oceanic flow though Nares Strait and the Canadian Archipelago was inhibited by land-based ice over North America connecting to the Greenland ice sheet at this time. The initial 3-
dimensional tracer fields (ocean salinity and temperature) were created by re-gridding output from the CCSM4 simulation to the MITgcm cube-sphere grid.

Figure 2.3: Last Glacial Maximum study area. The position of the landmasses and major ice sheets during the Last Glacial Maximum (LGM) are shown by the off-white and green shading, respectively.
Table 2.1: Mean Winter (JFM) and Summer (JAS) forcing for an LGM (CCSM4) and modern day (ERA-40) atmosphere of the Arctic Ocean north of Fram Strait, assuming LGM coastlines.

<table>
<thead>
<tr>
<th></th>
<th>Modern</th>
<th>LGM</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>JFM</td>
<td>JAS</td>
</tr>
<tr>
<td>10m wind velocity (m s(^{-1}))</td>
<td>1.46</td>
<td>0.86</td>
</tr>
<tr>
<td>2-m air temperature (°C)</td>
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<td>-1.60</td>
</tr>
<tr>
<td>Precipitation (mm day(^{-1}))</td>
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<td>0.95</td>
</tr>
<tr>
<td>Downward Longwave (W m(^{-2}))</td>
<td>163.40</td>
<td>289.20</td>
</tr>
</tbody>
</table>
Figure 2.4: Winter (JFM) and summer (JAS) boundary conditions in the Arctic from the CCSM4 Last Glacial Maximum control simulation. Rows correspond to 10-m wind speed, surface temperature, precipitation and downward longwave radiation. The white contours in the stereographic view are the location of LGM ice sheets at full extent.
2.5. Modeling the Glacial Arctic Sea Ice

The MITgcm was spun up for 1500 years with glacial boundary conditions producing a mean equilibrium sea ice thickness of 45 meters and a volume of $2.50 \times 10^{14}$ m$^3$ suggesting the Arctic could grow massively thick sea ice during extremely cold conditions. Proxy studies (as previously discussed in Chapter 1) in the Arctic suggest thick, permanent sea ice existed during the Last Glacial Maximum (LGM), but they do not tell us much about the actual thickness. There are many factors affecting the growth and thickness of sea ice, the most important of these factors being the total downward longwave radiation (DWL) at the surface governed by total cloud cover. Simulating cloud cover in numerical models has always been a challenge due to the sub-grid scale processes that occur at the molecular level. Parameterizations governing cloud formation is an important area of research because these equations govern how much radiation is making it from the atmosphere to the surface, which have broad effects, especially on sea ice in the Polar Regions. Using analysis of 23 Intergovernmental Panel on Climate Change (IPCC) models used to simulate climate change for the 21st century in the AR4 report, (Eisenman et al., 2007) found there is a >20 Wm$^{-2}$ intermodal spread in downwelling longwave radiation incident to the surface of the Earth associated with simulated total integrated cloudiness (Figure 2.5). It is therefore likely that the actual downward longwave radiation over the Arctic during the LGM might have differed from the values in CCSM4 by at least this amount. Although most of the climate models in the latest IPCC Fourth Assessment simulate present day Arctic sea ice well, Eisenman et al. (2007) suggests this is due to compensation by tuning parameters such as albedo.
Figure 2.5: Simulated Arctic (70-90°N) 1980-1999 mean monthly vertically integrated “cloudiness” (A) and downward longwave radiation at the surface (B) from the IPCC AR4 modelling studies. Surface radiation can vary as much as ±20 Wm\(^{-2}\) between models. (Source: Eisenman et al. 2007)

To explore the sensitivity of equilibrium sea ice thickness to errors in solar radiation due to uncertainties in cloud cover, I performed four 1000-year integrations that varied the downward longwave radiation over the Arctic by -20, -10, +10, and +20 Wm\(^{-2}\) (Figure 2.6).
Figure 2.6: Mean annual sea ice volume and thickness of the spin-up integration and radiation perturbations. All simulations start from the end of the spin-up simulation. The blue line shows equilibrium thickness of the spin-up integration and the branching lines show the radiation perturbations.

Here it was found that a decrease (increase) in surface radiation by 10 Wm$^{-2}$ produces an increase (decrease) in sea ice thickness of 14.6% (12.1%) producing equilibrium thickness of 52m (39m), compared to the control integration of 45m (Table 2.2). Decreasing (increasing) the surface downward radiation further by 20 Wm$^{-2}$ increases (decreases) the equilibrium sea ice thickness by 22% (30%), while also increasing (decreasing) the mean volume of sea ice by 18% (32%) (Table 2.2). In other words,
changing the DLW by only +/- 20 Wm$^{-2}$ causes a 25m variation in equilibrium sea ice thickness.

Initially, to gain more insight into how realistic the MITgcm sea ice thicknesses are, I compared them to other similar LGM simulations (Figure 2.7) to find other LGM models exhibit mean sea ice thickness of 5.2m to 15.7m. Although these models simulate thinner sea ice compared to my simulations, reconstructions of past sea ice extent are difficult and it is possible that sea ice is artificially ‘capped’ at an upper threshold in many of these models to avoid model limitation.

Figure 2.7: Simulated Arctic Ocean Last Glacial Maximum (LGM) sea ice thicknesses from the PMIP2 (Paleoclimate Model Intercomparison Project 2) project amongst six climate models. The grey box indicates the range of sea ice thickness and the blue line and number denote arctic sea ice mean for each model. Data from Li et al., (2010).
For the glacial simulation (discussed next), a decision was made to increase the DLW over the Arctic by +20 Wm$^{-2}$ so that our results more closely resemble other LGM modelling studies (Li et al., 2010). It is thus likely that our results provide a lower bound for the volume of freshwater that could have been stored as sea ice in the Arctic prior to deglaciation, and the implications for this are discussed in the final chapter of this dissertation in terms of future directions.

Table 2.2: Downward longwave radiation (DLW) perturbation experiments with respective mean sea ice volume and thickness. Numbers in parentheses are the percent change from the Control integration.

<table>
<thead>
<tr>
<th>Radiation Perturbation</th>
<th>Mean sea ice volume (km$^3$)</th>
<th>Mean sea ice thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>2.50 x 10$^5$</td>
<td>45.30</td>
</tr>
<tr>
<td>+10 Wm$^{-2}$</td>
<td>2.13 x 10$^5$ (-14.8%)</td>
<td>39.81 (-12.1%)</td>
</tr>
<tr>
<td>-10 Wm$^{-2}$</td>
<td>2.77 x 10$^5$ (+10.8%)</td>
<td>51.89 (+14.6%)</td>
</tr>
<tr>
<td>+20 Wm$^{-2}$</td>
<td>1.68 x 10$^5$ (-32.8%)</td>
<td>31.53 (-30.4%)</td>
</tr>
<tr>
<td>-20 Wm$^{-2}$</td>
<td>2.97 x 10$^5$ (+18.8%)</td>
<td>55.32 (+22.2%)</td>
</tr>
</tbody>
</table>
CHAPTER 3
THE GLACIAL ARCTIC: SIMULATED FRESHWATER FLUXES AND SEA ICE STORAGE

3.1. Introduction

This chapter discusses the results of the control simulation of the Arctic under full glacial atmospheric boundary conditions. Results show that the glacial Arctic Ocean is extremely isolated from the major oceans, reducing any major heat flux into the basin limiting the basal ocean-ice heat flux to 0.4 Wm$^{-2}$. This creates an environment suitable for sea ice to grow tens of meters thick, especially in the Western Arctic Basin. Analysis of the freshwater budget of the glacial Arctic shows that the basin can store ~3-times the amount of freshwater as the modern Arctic and 14-times the amount of sea ice.

3.2. Simulated Glacial Arctic Conditions

Results from the numerical models show the glacial Arctic was much colder than the modern Arctic. The relatively warm intermediate Atlantic Layer, a pervasive feature across the entire modern Arctic Ocean, is ~3°C cooler, 500m deeper, and restricted to the eastern Arctic basin (Figure 3.1b-c). For reference, the MITgcm simulates the deep warm Atlantic water penetration associated in a modern Arctic depth profile rather well and is support by Jakobsson (2013). In Figure 3.1, the warm Atlantic Layer is shallow and in agreement with that depicted in the schematic of Jakobsson et al., (2013) (Figure 3.2), but is confined south of the Fram Strait and does not penetrate deep into the central Arctic Ocean, limiting heat flux into the Arctic basin. Most of the interior basin ocean-ice heat
flux values approached 0.4 Wm$^{-2}$, which equates to an ~80% reduction in heat flux ($F_b$) compared to modern.

Sea ice patterns in the Arctic Ocean and north Atlantic Ocean are comparable to those derived from oceanic proxies by the GLAMAP 2000 working group (Pflaumann et al., 2003) (Figure 3.3). The thickest sea ice is found in the Western Arctic Basin, while thinner sea ice resides in the Eastern Arctic Basin. The reason for this dichotomy is due to the eastern Arctic basin being influenced by warm water intrusion from the north Atlantic Ocean, and a slightly stronger atmospheric circulation over Fram Strait making the ice more mobile. Here, sea ice moves >1.8 ms$^{-1}$ or 5.7 x 10$^7$ myr$^{-1}$ while in the western Arctic sea ice motion is largely stagnant, creating an environment suitable for ice to thicken up in some areas (Figure 3.1a). In fact, in this location a ~306,000 km$^2$ region of ice, up to 80m thick in places, stretches ~300-km offshore along the Canadian Arctic in good agreement with evidence from sediment cores that indicate a long period of minimal, or even non-deposition in this region between ~23-17 kyr ago (Jakobsson et al., 2013; Nørgaard-Pedersen, 2003; Polyak et al., 2009) (Figure 3.2). This is important, as evidence from Jakobsson et al. (2016) suggests there may have been a >1-km thick ice sheet covering this area during MIS 6. This evidence also suggests that an ice-sheet like formation could have formed from sea ice during the LGM, although an ice-sheet >1-km thick during this time is unlikely.
Figure 3.1: Study Area. (A) Map of the Arctic Ocean and subpolar North Atlantic during the Last Glacial Maximum (LGM) showing sea ice extent and thickness (gray-orange/red shading) simulated by our model. The red line indicates the location of the cross sections displayed in panels B-C and the labels (Fs, Lr, and Bs) represent the location of Fram Strait, Lomonsov Ridge, and Bering Strait, respectively. The position of the landmasses and major ice sheets during glaciation are shown by the off-white and green shading, respectively. (B-C) Cross sections of absolute water temperature (°C) through the Nordic Seas and Arctic Basin for (B) the LGM and (C) modern-day are shown. The orange triangle highlights the location used to construct the vertical salinity and temperature profiles displayed in Figure 3.3.
Figure 3.2: Conceptual model for the modern (top) and glacial (bottom) Arctic circulation pattern. The MITgcm simulates the modern Arctic depth profile rather well compared to the schematic here (top). However, in our model, warm water does not penetrate all the way through the Arctic basin like that shown here. Instead, warm Atlantic water is confined to the outermost areas of the Fram Strait, albeit very shallow. (Source: Jakobsson 2013).
Figure 3.3: During the LGM, our model indicated that the Arctic Ocean supported very thick slow-moving sea ice. The top panel (A) shows that the majority of the Arctic Ocean was covered by 50-80m thick sea ice that occupied an area of ~306,000 km². The bottom panel (B) shows that the thinnest ice (which is located in the eastern Arctic basin) would have been the most mobile while thicker ice (located in the western Arctic basin) would have been moving much slower. For example, ice >60m thick off the north coast of Canada was moving at <100 m yr⁻¹.

3.2.1 Comparison with the high-resolution simulation

The spatial extent of sea ice between the high-resolution and coarse resolution model is similar, although the high-resolution model resolves the seasonal changes of sea
in the Norwegian Sea and Fram Strait rather well. Most of the Arctic Basin in this simulation is filled with ice >20m. During the winter, sea ice in the Norwegian Sea is 2-3m thick, with the thickest ice (~6-7m) along the northern coast of Iceland. The lack of warm water intrusion from the North Atlantic keeps sea ice just north of the Fram Strait >10m thick. During the summer, warm water intrusion from the North Atlantic intrudes into the Norwegian Sea creating ice-free conditions, especially along the western coast of Svalbard. The thickest sea ice resides along the eastern coast of Greenland, transported via the East Greenland Current (EGC) from the Fram Strait.

Figure 3.4: Winter (left) and Summer (right) sea ice thickness in the MITgcm high-resolution regional configuration. During the summer, relatively warm, North Atlantic water enters the eastern Nordic Seas and the West Spitsbergen Current creating ice free conditions along the coast of Svalbard. The white areas correspond to area of no-ice (ice free conditions), the off-white shading illustrates the major Northern Hemisphere Ice Sheets and the green shading is ice-free land.
3.2.2 Comparison with an idealized sea ice model

An additional set of experiments to validate the thicknesses of sea ice produced by our low and high resolution MITgcm simulations in the western Arctic was subsequently undertaken using the 1-dimensional thermodynamic sea ice model developed by Bitz and Lipscomb (1999). This idea was motivated by the research of both Crary et al. (1955) and Walker and Wadhams (1979) who showed that given timescales of several hundreds of years and conditions colder than modern-day, ice could reach thicknesses of several tens to hundreds of meters thick.

When forced with modern-day climatic conditions, the model produces a realistic equilibrium sea ice thickness of ~2-3 m (Figure 3.5a-b). Configuring the model to simulate ice growth during the LGM (by forcing the model with the same CCSM4 LGM boundary conditions used to drive the MITgcm and setting the basal ocean-ice heat flux to 0.4 Wm$^{-2}$) led to ice reaching ~150m when summer and winter precipitation was accounted for (0.22 mm day$^{-1}$ and 0.52 mm day$^{-1}$, respectively) and ~80m when precipitation was set to zero (Figure 3.5c).

These results thus support the finding that thick sea ice could have grown in the western Arctic, given the right climatic conditions. The results from our MITgcm sea ice growth experiments were also corroborated by the idealized sea ice model of a 1-dimensional, non-dynamic (e.g. stationary) sea ice column model.

In addition to the 1-D sea ice model, the research into the Snowball Earth hypothesis (e.g. Pollard 2005) show that sea ice thickness can be determined by
\[ \Delta z = \frac{k \Delta T}{F_g} \]  

(3.1)

where \( \Delta z \) is the ice thickness, \( k \) is the thermal conductivity (2.5 Wm\(^{-1}\) K\(^{-1}\)), \( F_g \) is the thermal heat flux and \( \Delta T \) is the temperature difference between the atmosphere and the seawater beneath the ice. For our calculations, substituting \( F_g \) for the ocean-ice heat flux of our simulation (~0.4 Wm\(^{-2}\)) and calculating the difference between the air and water (~23°C), we get a thickness of 143.7m, similar to our results from the 1-d thermodynamic sea ice column model.

3.3. The large-scale freshwater cycle of the glacial Arctic

The modern-day Arctic has a complicated freshwater exchange with the North Atlantic and Pacific Oceans via the Bering Strait, Fram Strait and Barents Sea gateways and is controlled by a conjunction of latitude, geography and marine processes (Serreze et al., 2006). The storage and release of freshwater from the Arctic system may have played a major role affecting the stability of the Atlantic Meridional Overturning Circulation (AMOC) (e.g. Manabe and Stouffer, 1988) during Earth’s glacial climate periods.

Indeed, observations of the modern-day Arctic support this idea. For example, a increased freshwater and sea ice export at the Fram Strait during the 1960’s led to the formations of a low salinity anomaly (The Great Salinity Anomaly) that was traceable around the subpolar gyre for over a decade and may have weakened the strength of the AMOC by 1-3 Sv (Mysak et al., 2005). Other GSA’s may have also occurred during the 1980s and 1990s (Belkin, 2004; Belkin et al., 1998) decreasing North Atlantic salinity by
~0.4 psu. If changes in freshwater can happen so rapidly during the modern age (e.g. Great Salinity Anomaly), a mechanism like this could have happened during glacial events, perhaps on an even larger scale.

Figure 3.5: Idealized thermodynamic sea ice growth experiments. Under modern-day forcings, Arctic sea ice has an equilibrium thickness of ~2.5 m when precipitation is accounted for (A) and ~2 m when precipitation is set to zero (B). Configuring the model for the LGM (C), lead to an equilibrium sea ice thicknesses of ~80m (~150m) when precipitation was switched off (on). This result is very consistent with the solution found in our coupled dynamic/thermodynamic ice model where the most stagnant ice (i.e. that along the edge of the western Arctic) grows to ~80m thick.
3.3.1. Calculating Freshwater Flux, Freshwater Storage and Sea Ice Flux

In order to study the freshwater cycle of the glacial Arctic, I calculated an Arctic Ocean freshwater budget for the Last Glacial Maximum. It is important to note that this is a novel analysis that to-date has only been performed on the modern-day Arctic. In this study I define the Arctic using the same domain as Serreze et al. (2006) i.e. the area north of the Fram Strait, the Barents Sea, and the Bering Strait. By using a combination of reanalysis data (i.e. ECMWF ERA-40) and observational data on a land-sea model, Serreze et al., (2006) estimates the modern Arctic Ocean can store over 74,000 km$^3$ of liquid freshwater, with 60% of this water stored in the Beaufort Sea and is visible by a 20m thick vertically integrated column of freshwater (Figure 3.6). In addition, ~10,000 km$^3$ of freshwater is stored as sea ice (~10,000 km$^3$).

The freshwater content (m) of the Arctic can be calculated as:

$$FW_{storage} = \int_{Depth}^{0} \left( 1 - \frac{S}{S_{ref}} \right) dz$$

(3.2)

where $S$ is the salinity of the water, $S_{ref}$ is the reference salinity, 34.5 psu, and $dz$ is the vertical thickness (Serreze et al., 2006). In keeping with the work of Serreze and others, water that has a salinity value higher than 34.5 psu is considered salty and is not included in this calculation.

To determine the freshwater flux through each of these major Arctic gateways (i.e. Fram and Bering Straits and the Barents Sea), I assume that water with a salinity of >34.5
to be a freshwater source, whereas values greater than the reference salinity are considered freshwater sinks. In the model, a series of gates were placed in the Fram Strait, Barents Sea and Bering Straits to allow liquid freshwater export to be calculated as:

$$FW_{\text{flux}} = \int_{L}^{0} \int_{Depth}^{0} \left(1 - \frac{S}{S_{\text{ref}}} \right) U_{n} d_{z}$$

(3.3)

where $S$ is the surface salinity, $S_{\text{ref}}$ is the reference salinity, $d_{z}$ is the depth of the vertical grid cell, $L$ is the length of the section, $U_{n}$ is the ocean velocity, and $dl$ is the cell width,

Figure 3.6: Arctic Ocean integrated freshwater content for the Arctic domain, as defined by Serreze et al. (2006). Box 1 represents Lomonosov Ridge. Most of the liquid freshwater is stored in the Beaufort Sea (box 2) as denoted by the 20m maximum thickness of freshwater. (Source: Condron et al. 2009)
And sea ice transport as:

$$SI_{\text{flux}} = \int_0^L (U_i h \, dl) \, dl$$  \hspace{1cm} (3.4)$$

where $U_i$ is the sea ice velocity in m$\text{s}^{-1}$, $h$ is the sea ice (ice+snow) thickness in meters and $dl$ is the horizontal grid length in meters.

3.4. Glacial Arctic Freshwater Budget Results

In this section, the glacial Arctic freshwater system is compared to the modern Arctic freshwater system as viewed in Figure 3.8. The geography of the Arctic Basin plays a major role in the freshwater storage and export of the entire system. Unlike the modern Arctic previously discussed, most of the major straits that are essential to freshwater export (Condron et al., 2009) are closed during the LGM, limiting freshwater fluxes in and out of the Arctic Basin. In my calculations, I find there is no liquid freshwater export out of the Arctic, but instead that it is primarily composed of sea ice export. The model indicates that the total freshwater content of the glacial Arctic is 83,006 km$^3$; spatially, the freshwater was evenly spread across the entire Arctic Ocean, rather than being confined to the Beaufort Sea, as observed for modern-day conditions (Figure 3.7). Overall, there is 10,000 km$^3$ more liquid freshwater in the Arctic Ocean than modern day.

More importantly, the Arctic Ocean stores 142,230 km$^3$ of freshwater as ice, which is 14-times the modern Arctic. This result is extremely significant for assessing the role that the storage and release of Arctic sea ice might have played in modulating deglacial
climate. In short, it begs the question of whether the release of this sea ice would have created “Mega Great Salinity Anomalies” (MGSA’s) in the subpolar North Atlantic.

Further analysis of the glacial Arctic freshwater budget indicates that ~67% of the freshwater entering the Arctic Ocean is from land runoff, which although greater percentagewise than modern (where it is 35%), it is less than modern in terms of volume as a result of the Arctic being smaller. The data also supports a very dry environment, where LGM modeled precipitation (e.g. 539 km$^3$ yr$^{-1}$) is ~1/6$^{th}$ of the modern-day Arctic precipitation and LGM modeled evaporation is ~1/3$^{rd}$ the evaporation compared to modern-day Arctic precipitation (e.g. 392 km$^3$ yr$^{-1}$) (Figure 3.8).

![Sea Ice Thickness](image)

**Figure 3.7**: The liquid freshwater content of the glacial Arctic Ocean. Colors are vertically integrated freshwater content, calculated from the Equation 3.2.
Indeed, evidence supports thick sea ice, but the actual thickness remains an enigma. Studies indicate a low obliquity angle and frigid temperatures that suggest sea ice could have been extremely thick. Sea ice of this magnitude, along with a change in geography due to low sea level, may have disrupted normal Arctic Ocean circulation that we see in the modern Arctic. This leads to increased resonance time of freshwater in the basin itself causing the Arctic to store almost 3-times the amount of freshwater as the modern Arctic. Previous studies (i.e. Condron et al., 2009) illustrate that with the right conditions, freshwater can easily be exported out of the basin and into the North Atlantic very quickly in large volumes. Therefore, we imagine that with a change in heat flux or atmospheric pattern in the Arctic, sea ice may be exported quickly and in large volumes possibly causing abrupt climate change.
Figure 3.8: Mean freshwater budget for the glacial and modern Arctic. All fluxes and trends are in km$^3$ yr$^{-1}$ and for freshwater storage km$^3$. Arrows are proportionate to the size of the flux. Notice the glacial Arctic mainly transports freshwater in the form of sea ice whereas the different circulation regime in the modern Arctic allows transport of liquid freshwater via multiple routes.
CHAPTER 4
DEGLACIAL CLIMATE MODULATED BY THE STORAGE AND RELEASE OF ARCTIC SEA ICE

4.1. Abstract

This chapter presents the results from a series of numerical experiments designed to mobilize Arctic sea ice to study the potential impact of Arctic sea ice export on ocean circulation and climate. Here, I use a climate model to (Chapter 2 and 3) to show that the episodic break-up and mobilization of thick perennial Arctic sea ice during this time would have released considerable volumes of freshwater directly to the Nordic Seas where processes regulating large-scale climate occur. Massive sea ice export events to the North Atlantic are generated by the MITgcm model whenever the transport of sea ice is enhanced, either by changes in atmospheric circulation, rising sea level submerging the Bering land bridge, or glacial outburst floods draining into the Arctic Ocean from the Mackenzie River. In addition, several other deglacial mechanisms also exhibit some sea ice discharge when boundary conditioned are changed, such as flooding of the Barents Sea. I find that the volumes of freshwater released to the Nordic Seas are similar to, or larger than, those estimated to have come from terrestrial outburst floods, including the discharge at the onset of the Younger Dryas. My results provide the first evidence that the storage and release of Arctic sea ice helped drive deglacial climate change.
4.2. Introduction

A series of perturbation experiments were performed to explicitly test the sensitivity of Arctic sea ice break-up and mobilization to climatological, geographical, and hydrological changes that occurred in the Arctic during deglaciation (Table 4.1)

Table 4.3: A summary of the main model experiments performed using MITgcm to test climate sensitive to the mobilization of Arctic sea ice.

<table>
<thead>
<tr>
<th>Experiments</th>
<th>Model Configuration</th>
<th>Model Resolution</th>
<th>Details</th>
</tr>
</thead>
<tbody>
<tr>
<td>TPD_5.0_5yr</td>
<td>Global</td>
<td>2.8° (280-km)</td>
<td>TPD enhanced for 5yr (w. 5 m/s wind) followed by 50 yr relaxation. Cycle repeated 5-times</td>
</tr>
<tr>
<td>TPD_7.5_5yr</td>
<td>Global</td>
<td>2.8° (280-km)</td>
<td>TPD enhanced for 5yrs (w. 7.5 m/s wind) followed 50 yr relaxation. Cycle repeated 5-times</td>
</tr>
<tr>
<td>TPD_5.0_50yr</td>
<td>Global</td>
<td>2.8° (280-km)</td>
<td>TPD enhanced for 50yrs (w. 5 m/s wind) followed by 100 yr relaxation. Cycle repeated 2-times</td>
</tr>
<tr>
<td>TPD_7.5_50yr</td>
<td>Global</td>
<td>2.8° (280-km)</td>
<td>TPD enhanced for 50yrs (w. 7.5 m/s wind) followed by 100 yr relaxation. Cycle repeated 2-times</td>
</tr>
<tr>
<td>BS_open</td>
<td>Global</td>
<td>2.8° (280-km)</td>
<td>Bering Strait opened to 50 m depth</td>
</tr>
<tr>
<td>Mack_flood</td>
<td>regional</td>
<td>1/6° (18-km)</td>
<td>Glacial outburst flood (of 5 Sv) released from Mackenzie R. (Canadian Arctic) for 1 yr.</td>
</tr>
<tr>
<td>Bsea_open</td>
<td>Global</td>
<td>2.8° (280-km)</td>
<td>Barents Sea opened to 250 m depth</td>
</tr>
<tr>
<td>BSBsea_open</td>
<td>Global</td>
<td>2.8° (280-km)</td>
<td>Barents Sea and Bering Strait opened</td>
</tr>
</tbody>
</table>

Notes: “TPD” refers to the Transpolar drift of Arctic sea ice. In the model ice movement was enhanced by increasing the near-surface wind speed over the region 80°N-90°N by 20°W-20°E. The middle values of the experiment name ‘_X.X_’ denote the wind speed used in the perturbation, in this case 5.0 ms⁻¹ and 7.5 ms⁻¹, while the final values of the name refer to the number of years the perturbation was applied for.
4.2.1. Changes in Atmospheric Forcing

During the LGM, the pressure gradient between the Greenland Ice Sheet and the Nordic Seas was significantly enhanced (Figure 4.1), producing mean annual southerly wind speeds across Fram Strait of 5.6 ms\(^{-1}\) that peaked over the East Greenland Current (to the west) at ~8.9 ms\(^{-1}\).

![Figure 4.1: Mean annual sea level pressure. The surface pressure over the Arctic and North America is shown here for the Last Glacial Maximum climate simulated by CCSM4 (A) and the Modern-day climate derived from the NCEP reanalysis dataset through the period 1979-1999 (B). During the Last Glacial Maximum, there existed an enhanced pressure gradient between the Nordic Seas and the central Arctic Basin denoted by tighter isobars. This may have played a role in significantly enhancing southerly winds out of the Arctic and into the North Atlantic increasing sea ice export during the Last Glacial Maximum. Note the landmasses are shaded in green and the pressure contours are each 1 hPa.](image)
To study the impact of changes in atmospheric circulation on Arctic sea ice transport, winds over the region 80°N-90°N by 20°W-20°E were artificially perturbed to invigorate the transpolar drift and simulate an increase in the pressure gradient between the Nordic Seas and the Arctic Ocean (Figure 4.2). In each case, winds in this region were set to blow in a consistently southerly direction, i.e. from the central Arctic towards Fram Strait, at speeds of either 5 ms\(^{-1}\) or 7.5 ms\(^{-1}\). In the first experiment, the wind was set to blow constantly for 5 years, after which time the winds were returned to those in the Control integration for 50 years. This 55-year cycle was then repeated 5 times. In the second experiment, the wind was set to blow steadily south for 50 years before being returned to conditions in the Control simulation for 100 years. This 150-year cycle was repeated twice. An additional experiment was conducted to study whether a negative phase of the North Atlantic Oscillation (NAO) would increase sea ice export from the Arctic Ocean. Here, the glacial simulation was forced with modern winds from a negative phase of the NAO simulated by CCSM4.
Figure 4.2: The region (80°N-90°N by 20°W-20°E) described in Section 4.3.1 where the atmospheric winds were perturbed at 5 and 7.5 ms\(^{-1}\). This experiment was designed to test a strengthening of the Transpolar Drift and the subsequent sea ice export.

Other atmospheric forcing mechanisms explored during our simulations included studying the effects of a modern positive North Atlantic Oscillation (NAO+). Past studies have shown that positive NAO wind forcing accelerates the export of liquid freshwater into the North Atlantic from the Arctic Basin (Condron et al., 2009). Here, the MITgcm wind fields were forced with positive NAO wind forcing from the European Center for Medium Range Forecasting (ECMWF) ERA40 reanalysis data.
4.2.2. Opening the Bering Strait

Based on fossil evidence from Banks Island in the western Canadian Archipelago, the Bering Strait opened around the time of the Younger Dryas. Radiocarbon dates suggest the Bering Strait may have been opened between 13.2-13.4 kyr BP, which is at least as old as the Younger Dryas (England and Furze, 2008; Hughen et al., 2004). To study the impact of rising sea level during deglaciation, I opened the Bering Strait land bridge in the model by artificially modifying the model land mask and bathymetry. The land grid points corresponding to the modern-day Bering Strait were changed to wet grid cells with a depth of 50m and the salinity and temperature through the Bering Strait was calculated by linear interpolated values between the North Pacific and western Arctic on either end.

4.2.3. Opening the Barents Sea

Opening of the Barents Sea represents an important process that occurred during deglaciation from the Last Glacial Maximum (LGM) to the Holocene. The Barents ice sheet covered a large portion of northern Europe and Scandinavia blocking warm, North Atlantic water from flowing north via the Norwegian Current through the Barents Sea to the Arctic basin. Data suggests the deglaciation of the Barents Ice Sheet started ~15 kyr BP, with most of the Barents Sea open water by ~12 kyr BP (Landvik et al., 1998). Here, we test whether the disintegration and disappearance of the Barents Ice Sheet may have allowed warm water flow into the Arctic basin, destabilizing sea ice and creating a major sea ice export event. Similar to the Bering Strait experiment, we removed the northeastern half of the Barents Ice Sheet by artificially modifying the model land mask. The land grid
cells were changed to “wet” grid cells, with a depth of 250m and a temperature and salinity were set similar to those around the region.

4.2.4. Holocene (open Bering Strait and Barents Sea) Arctic Basin

A Holocene-like Arctic Ocean is influenced by water from the North Pacific and Atlantic Oceans. Here, the Bering Strait and Barents Sea are opened to simulate a Holocene geography of the Arctic region. I took the two experiments mentioned previously and combined them into one simulation to determine if a Holocene Arctic geography could have played a major role in destabilizing and mobilizing thick sea ice from the Arctic into the subpolar north Atlantic Ocean. This model set-up was used to simulate a completely deglacial-like climate and whether opening both straits could invigorate the Trans-Polar Drift and export large amounts of sea ice, via the Fram Strait, into the North Atlantic.

4.2.5. Simulating glacial outburst floods

Using the high-resolution Arctic regional configuration of the MITgcm described in section 2.2, a glacial meltwater flood originating from the western Arctic Basin is simulated. Here, I released 5 Sv of freshwater from the mouth of the Mackenzie River (~69.3° N, 133.8° W) for 1 year to simulate a catastrophic flood caused by a sudden drainage/rerouting of glacial Lake Agassiz to the Arctic. The flux released here is comparable to the discharge estimates around the time of the 8.2 kyr event (Teller et al., 2002). Meltwater entering the ocean was released from the closest 5 grid cells to the mouth of the Mackenzie River to simulate the exit of freshwater from the river. The water entering these cells has an initial salinity of 0 psu and temperature of 0°C.
4.4. Results

In this section, I describe the results of the initial perturbation experiments (section 4.3). We qualitatively and quantitatively analyzed the effect of increased wind on sea ice transport, the flux of sea ice when the Bering Strait is opened and the changes in sea ice flux with a large outburst flood of freshwater into the Arctic basin.

4.4.1. Atmospheric wind forcing

In perturbing the models wind field, I find changes in atmospheric circulation played a major role in transporting large amounts of freshwater (as ice) out of the Arctic Ocean and into the North Atlantic, mimicking a large freshwater outburst flood. In the simulation involving increased winds through the Fram Strait, repeating the wind ‘perturbation-relaxation’ pulse five times showed that each time the near-surface wind speed over the transpolar drift reached 7.5 ms\(^{-1}\), ice from the central Arctic was rapidly transported through Fram Strait to the Nordic Seas with fluxes peaking at \(~0.15\) Sv (Fig. 4.3a). Each period of enhanced sea ice export resulted in \(~18,500\) km\(^3\) of freshwater being discharged into the Nordic Seas, which is roughly double the freshwater discharge (\(~9,500\) km) estimated to have been released from glacial Lake Agassiz at the onset of the Younger Dryas cold episode (Teller et al., 2002). Still, the cumulative volume of ice exported to the Nordic Seas by the five periods of enhanced sea ice transport (\(~110,800\) km\(^3\)) was \(~70\)% of the estimated volume of freshwater released into the North Atlantic by the catastrophic outburst flood from Lake Agassiz around the time of the 8.2-kyr-event (Clarke et al., 2004; Teller et al., 2002).
Figure 4.3: The response of Arctic sea ice export at Fram Strait and ocean circulation to a mobilization of thick perennial sea ice in the central Arctic basin. Each time the transpolar drift is enhanced, sea ice export at Fram Strait (A) rapidly increases such that ice fluxes peak at ~0.15 Sv after 6 months. As this ice melted, it caused the central Nordic Seas to freshen by ~0.5 psu (B), the strength of the AMOC (C) to weaken by 0.5-0.7 Sv. Each weakening caused a 0.1 Pw decrease in heat flux entering the Arctic Basin (D). Gray shading corresponds to periods when the transport drift was enhanced in the model to mobilize the pack ice.

Compelling evidence that the discharge of sea ice to the subpolar North Atlantic directly impacts ocean circulation comes from an inspection of the salinity of the central
Nordic Seas and the strength of the AMOC (Fig. 4.3b-c). Here, I find that each period of enhanced sea ice export led to a decadal-length freshening of the Nordic Seas of ~0.5 psu and a 0.5-0.7 Sv slowdown in the strength of the AMOC. Each of these weakening produced a 0.1 Pw (1 x 10^{14} Watt) decrease in heat transport into the Arctic Ocean via the Fram Strait, signifying a weakening of the northern branch of the Gulf Stream. While the greatest slowdown of AMOC coincided with the peak Arctic sea ice export, a full recovery of the overturning cell and heat transport took ~30-40 years.

The robustness of changes in the strength of the transpolar drift influencing Arctic sea ice export and the AMOC during deglaciation is corroborated by an additional set of experiments that perturbed the near-surface winds over the eastern Arctic for 50 years, followed by a 100-year return to LGM conditions (Fig. 4.4). Again, the transport of sea ice to the Nordic Seas peaked at ~0.15 Sv, but in this experiment the flux of ice remained sustained at ~0.1 Sv for ~45 years, which has been shown to weaken the strength of the AMOC in numerous climate model simulations (e.g. Stouffer et al., 2006). It is also important to note that the total volume of freshwater discharged as ice in this experiment is equivalent to the drainage of glacial Lake Agassiz ~8,200 years ago, but rather than being a one-time event, my model was able to produce a discharge event of this magnitude each time the transpolar drift was invigorated. In addition, we found that a 5 ms^{-1} wind perturbation over the eastern Arctic resulted in an increase in sea ice export at the Fram Strait after ~20 years, suggesting that the circulation of the central Arctic may take some time to ‘spin-up’ before ice begins to mobilize. However, once this event began, sea ice fluxes of ~0.1 Sv were generated at Fram Strait and ~76,000 km^3 of freshwater was exported to the Nordic Seas.
Figure 4.4: Sea ice export and ocean circulation response to enhancing the transpolar drift of sea ice. Here, the atmospheric drag force exerted on sea ice in the Arctic Ocean (80°N-90°N by 20°W-20°E) was increased for 50 years (gray shading) before being returned to conditions in the Control simulation for a further 100 years. The experiment was performed using wind perturbations of 5 ms\(^{-1}\) (orange line) and 7.5 ms\(^{-1}\) (red line). Each time the transpolar drift was enhanced the export of sea ice to the North Atlantic at Fram Strait increased (A), which resulted in a freshening of the central Nordic Seas (B), reduction in the strength of the AMOC (C) and global heat transport (D). During the first two decades, ~50,000 km\(^3\) of sea ice was exported to the North Atlantic, which is more than 5-times the volume of meltwater released from glacial Lake Agassiz at the onset of the Younger Dryas.
Recently, Condron et al. (2009) found repeated modern negative NAO (North Atlantic Oscillation) forcing allows freshwater to be retained in the Arctic, whereas positive NAO forcing accelerates freshwater export into the North Atlantic. Using the same positive NAO dataset as Condron (2009), I tested the effect of a NAO+ on LGM sea ice export (Figure 4.5). Results suggest that the effect on sea ice is opposite to that of liquid freshwater; sea ice export from the Arctic Ocean reduced by 42% compared to the control integration under a NAO+ scenario. The overall evolution of the AMOC during NAO+ wind forcing included a sharp increase in AMOC strength followed by a 70-year decrease in strength, and finally, recovery back to control conditions. The abrupt atmospheric change to a strong NAO phase increases AMOC convection for 50 years before weakening and returning to control conditions within 100 years. Yet, sea ice export remains steady and less than the control for the entirety of the simulation.
Studies show there is considerable ambiguity between the NAO circulation and Fram Strait Arctic sea ice export (Tsukernik et al., 2010). Recently, it has been concluded sea ice export is more closely related cyclone frequency and strength (Maslanik et al., 2007). One cyclone has the potential to increase sea ice motion and ice velocity by a factor of three (Brümmer et al., 2003). I therefore suggest that a persistent and strengthened phase of the NAO (positive phase) may have actually created a negative feedback cycle on sea ice transport decreasing export into the North Atlantic. It is important to remember during the LGM, the jet stream and cyclone trajectory was suppressed further south inhibiting storms from entering the Fram Strait and mobilizing sea ice. A change to a positive phase
of the NAO does not seem to have a positive impact on sea ice export to create an export large enough to weaken the AMOC, at least in this model.

The results presented above illustrate that sea ice export, whether driven by atmospheric winds or the Transpolar Drift, can cause weakening of the AMOC and freshen the Nordic Seas although the NAO experiment did not agree with this. As illustrated above, these events may have happened multiple times in a short period causing AMOC weakening and reduced northward heat transport.

4.4.2. Flooding of the Bering Strait (BS_open)

The opening of the Bering Strait (BS_open) facilitates a flux from the North Pacific Ocean into the Arctic of 0.9 Sv with ~0.1 Sv (3210 km$^3$ yr$^{-1}$) of this being freshwater. The modelled mean annual volume transport is comparable to the observations of Woodgate and Aagaard (2005) whom calculated the Bering Strait mean annual transport to be ~0.8 Sv. The model indicates that the increased freshwater flux into the Arctic through this gateway increases total freshwater storage in the Arctic by ~10% from 225,236 km$^3$ to 248,900 km$^3$. The largest increase comes from the liquid freshwater component, which increases 67% from the control simulation, from 83,006 km$^3$ to 138,890 km$^3$. In contrast, opening the gateway causes a decrease in sea ice volume of 22% due to an increase in sea ice export at Fram Strait (Figure 4.6).

After the Bering Strait is opened (Fig. 4.7), ~0.8 Sv of Pacific Water entered the Arctic basin which caused invigoration of the upper Arctic Ocean circulation by 30% while also increasing the amount of ocean-heat transport into the western basin by 1-Terrawatt
(Tw = 10^{12} \text{ watts})(\text{Figure 4.8}). A warming of the intermediate depths (Fig. 4.7 C) is attributed to enhanced Atlantic Layer intrusion only when the Bering Strait is opened.

Figure 4.6: The simulated Arctic mean freshwater budget when the Bering Strait is opened (bs_open). Mean transports are represented by the bold arrows in km$^3$yr$^{-1}$. The width of the arrows is proportional to the size of the transport. Freshwater storage is in km$^3$.

This layer is significantly weakened and shallower in the LGM integration (Fig. 4.7 B). Accompanying this change, the intermediate depth Atlantic Waters warm by 5°C due to an increase in the northward heat transport from 4 to 13 Tw in the eastern Arctic basin (Fig. 4.8). Ocean-ice heat flux also increased 5-fold, from ~0.4 Wm$^{-2}$ to ~1.78 Wm$^{-2}$. This caused a ~30% decrease in sea ice thickness and volume. Taken together, these changes supported mobilization of the thick stagnant ice pack in the Canada Basin, such that after ~4 years the export of sea ice to the Nordic Seas at Fram Strait peaked at ~0.09 Sv and is coincident
with an AMOC weakening of ~23%. Sea ice discharge then showed a gradual reduction over a period of ~20 years but continued to be 66% higher than the control for the remainder of the simulation (Figure 4.9). During the first two decades, ~50,000 km$^3$ of sea ice was exported to the North Atlantic, which is more than 5-times the volume of meltwater released from glacial Lake Agassiz at the onset of the Younger Dryas.
Figure 4.7: Sea ice extent and cross section of LGM control and Bering Strait (open) integration. (A) Map of the Arctic and subpolar north Atlantic showing the sea ice extent and thickness simulated by our model during the Bering Strait (open) simulation. (B) A cross section of LGM and (C) Bering Strait (open) mean water temperatures (°C). The red line in (A) indicates the location of the cross-section area. The yellow arrow in (B) and (C) is the location of the temperature-salinity-depth plot in (D). Note the warm water intrusion in (D) denoted by the 5°C increase in 1000m temperature Ice sheet reconstructions (off-white patching) by Dyke et al. (2002). Note the open Bering Strait in the model land mask.
Figure 4.8: Mean annual northward Fram Strait heat transport for the Control, and the Bering Strait (open) simulations (blue and red lines, respectively). The yellow line represents Bering Strait northward heat transport from the northern Pacific Ocean. Initially, 4 Terawatts of heat energy are being transporting through the Fram Strait into the eastern Arctic basin. When the Bering Strait is opened (red line), heat transport increases to 13 Tw. Heat energy being transported through the Bering Strait from the North Pacific is only ~1 Tw indicating most of the heat is coming from the North Atlantic.

To determine the robustness of these changes, sea ice was artificially capped at 9m across the entire basin (hereafter known as “BS_cap”). Here, the BS_open simulation resulted in a slightly weaker AMOC compared to the BS_cap simulation. Within the first 5 years of the integration, sea ice export increases ~40% from the BS_open simulation (Fig. 4.9). After 5 years, the BS_cap simulation reaches an equilibrium sea ice export of 10,000 km³yr⁻¹ compared to 8,000 km³yr⁻¹ of the BS_open simulation. This is owed to a faster sea ice export due to thinner, more mobile sea ice. In addition, the liquid freshwater being exported into the North Atlantic between the BS_open and BS_cap simulations is also comparable to each other. After 5 years, the liquid freshwater flux from the Fram Strait...
in both the BS_open and BS_cap simulation is 3,700 km³ yr⁻¹ and 3,400 km³ yr⁻¹, respectively, and lasts for the entirety of the simulation as a new equilibrium. Although these results are analogous, the BS_open simulation still produces a weaker AMOC due to a larger liquid freshwater export than the BS_cap simulation. This could indicate liquid freshwater from the Arctic Ocean may play a larger role in weakening the AMOC than the export of sea ice.

Figure 4.9: Arctic sea ice export at Fram Strait in response to opening the Bering Strait. Rising sea level during deglaciation resulted in a spin-up of the Arctic Ocean circulation as the Bering Strait was flooded and warm Pacific waters returned to the Arctic. This resulted in sea ice export to the Nordic Seas peaking at ~0.1 Sv after ~4 years.

4.4.3. Opening of the Barents Sea

When the Barents Sea was opened and all other straits were closed (Fig. 4.10), the Arctic acts a freshwater sink storing 2.5-times the amount of liquid freshwater and >14-times (1.49 x 10⁵ km³) the amount of sea ice, compared to the modern Arctic. Liquid
freshwater enters the Arctic via the Fram Strait and Barents Sea at 2,336 and 179 km³ yr⁻¹, respectively (Fig. 4.11).

Figure 4.10: Comparison between the modern observed and the Barents Sea open (Bsea_open) mean freshwater budgets. Mean transports are represented by the bold arrows in km³ yr⁻¹. The width of the arrows is proportional to the size of the transport. Freshwater storage is in km³.

The increase of liquid freshwater entering the Arctic Basin causes a 130% increase of total liquid freshwater storage compared to the Control, and 37% increase of total liquid freshwater compared to the Bering Strait simulation. Intriguingly, sea ice is exported out of the Arctic via the Fram Strait and Barents Sea at 1,556 km³ yr⁻¹ and 798 km³ yr⁻¹ respectively, parallel to the liquid freshwater being imported into the basin. This result points out something interesting: sea ice is transported out of the Arctic whereas liquid
freshwater is drawn into the Arctic at the intermediate depths and stored near the ocean surface. This is important with respect to sea ice formation. Without the Bering Strait throughflow, the Beaufort Gyre and the Transpolar Drift are weakened, abating the export of water and sea ice from the Arctic leading to an overabundance of Arctic freshwater (i.e. freshwater sink). However, it is noteworthy that the total sea ice export from the Arctic at this time is comparable to the export when the Bering Strait is opened. This suggests a collapse of the Barents Ice Sheet may also have a profound effect on sea ice export into the North Atlantic due to natural mass balance dynamical effects.

When the Barents Sea was opened, the response time of the sea ice is quick, and we see a major decrease in AMOC strength. In the first year of the simulation, 1,183 km$^3$ of freshwater was transported into the North Atlantic, which is 12% of the freshwater that may have been discharged from glacial Lake Agassiz prior to the onset of the Younger Dryas (Fig. 4.12a). During this time, ice $>3.5$m leaves the Arctic via the Fram Strait at 0.08 ms$^{-1}$ (not shown). After 5 years, $\sim$6,777 km$^3$ of freshwater was transported to the North Atlantic freshening the Nordic $\sim$0.3 psu (Fig. 4.12b). This is equivalent to a sea ice flux of $\sim$0.05 Sv. Like the Bering Strait (open) simulation, sea ice flux via the Fram Strait and Barents Sea is elevated and constant for the entirety of the integration triggering a 10% AMOC weakening after the first 10 years of the simulation and a decrease in global net northward heat transport of 34%, indicating that sea ice export is having some effect on Atlantic Ocean circulation (Fig. 4.12c, d). This may also indicate the northern branch of the Gulf Stream has weakened enough to affect heat transport.
Figure 4.11: Sea ice extent and cross section of LGM control and Barents Sea (open) integration. (A) Map of the Arctic and subpolar north Atlantic showing the sea ice extent and thickness simulated by our model during the Barents Sea (open) simulation. (B) A cross section of LGM and (C) Barents Sea (open) mean water temperatures (°C). The red line in (A) indicates the location of the cross-section area. The yellow triangle in (B) and (C) is the location of the temperature-salinity-depth profile in (D).
Figure 4.12: Sea ice export, ocean circulation and Fram Strait heat flux associated with opening of the Barents Sea. The simulation switches from control boundary conditions to Barents Sea (open) conditions at year 0. When the Barents Sea is opened, sea ice export decreases from the Fram Strait, but increases through the Barents Sea (A), which resulted in a freshening of the Nordic Seas by ~0.3 psu (B), a decrease in the strength of the AMOC (C), and a reduction in northward global heat transport (D).
Although Gulf Stream heat transport has weakened, a ~1.5 °C warming of the Arctic intermediate water between 400-2,500 m depth (Fig. 4.11d) is observed, such that the ocean-ice heat flux (~0.83 Wm\(^{-2}\)) is twice that of the Control simulation. This is owed to a ~15 Tw intensification of northward heat transport through the Fram Strait in the first 5 years of the simulation (Fig. 4.13). After 70 years, Fram Strait heat flux decreases by 46% and 8 Tw of heat is transported into the Barents Sea from the North Atlantic, and 2 Tw of heat is being moved northward through the Fram Strait for a total of 10 Tw of heat energy being transported into the Arctic Basin (Fig. 4.13) at this time. In the Barents Sea, the northward heat flux remains at 2 Tw through 100 years, warming the intermediate depths of the Kara and Laptev Seas. It is important to note this warming never extends deep into the Arctic basin and is confined to the Barents Sea at 100m depth.

![Figure 4.13](image)

Figure 4.13: Mean annual northward Fram Strait heat transport for the Control, and the Barents Sea (open) simulations (blue and red lines, respectively). The yellow line represents the northward heat transport from the Norwegian Current into Barents Sea. However, the heat transport in the Barents Sea is too weak to make it into the deep Arctic Ocean. It is significant that there is an increase in both Fram Strait and Barents Sea
northward heat transport, indicating a collapse of the Fennoscandian Ice Sheet may have had a larger role in transporting heat and water northward.

4.4.4. Holocene Arctic Configuration

Finally, with a Holocene geography, the Arctic exports >6,500 km³ yr⁻¹ of liquid freshwater and ice from the Barents Sea and Fram Strait gateways. Compared to Serreze’s modern observations, this is an increase in total export of 48% (Fig. 4.14). Most of this increase is due to the fact the Arctic Ocean with a Holocene configuration holds 3-times the amount of total freshwater as the modern Arctic, and 1.1-times the amount of freshwater as the LGM control configuration. Still, it is important to note in a glacial system, the Arctic stores as much as 66% more freshwater and as much as 1.3-times the amount of sea ice compared to contemporary measurements. In general, liquid and sea ice exports were comparable to the modern observations noted by Serreze et al. (2009) suggesting our simulated Arctic circulation is close to a modern circulation regime. The Bering Strait exhibits a ~1.0 Sv throughflow from the North Pacific into the Arctic Basin in our Holocene simulation. As a matter a fact, the transport from the Pacific is very comparable with other modern observations (Woodgate et al., 2012), albeit 40% higher in our Holocene simulation compared to the modern derived observations.
Figure 4.14: The simulated Arctic mean freshwater budget with a near-modern Arctic geography. Mean transports are represented by the bold arrows in km$^3$yr$^{-1}$. The width of the arrows is proportional to the size of the transport. Freshwater storage is in km$^3$. Here, the Arctic is acting as a freshwater source.
When the Bering Strait is opened, 15 Tw of heat energy is passed through the Fram Strait increasing Arctic intermediate water temperature $>3.5$ °C between 200 and 1,200 m depth (Fig. 4.15A-D, Figure 16) while increasing the ocean-ice heat flux to $\sim 1.80 \text{ Wm}^{-2}$, which is 5-times the ocean-ice heat flux in the Control simulation. At year 12, peak Fram Strait heat flux approaches 25 Tw of energy entering the Arctic Basin and decrease to 18 Tw entering equilibrium for the rest of the simulation. The ocean heat flux entering the basin is comparable to observations of Schauer (2004) and their estimate of an annual Fram Strait heat flux of $16 \pm 21$ to $41 \pm 5$ Tw. A combination of the heat flux from the Fram Strait and Barents Sea gateway helps to decreases sea ice volume by 25% (20.4 m) compared to the control simulation and increase basal heat flux to $1.80 \text{ Wm}^{-2}$. Comparatively, modern ocean-ice heat flux is $\sim 2.0 \text{ Wm}^{-2}$ suggesting the Holocene geography allows for more warm water to enter the basin even under glacial conditions.
Figure 4.15: Sea ice extent and cross section of the LGM control and the near-modern (Barents Sea and Bering Strait open) integration. (A) Map of the Arctic and subpolar north Atlantic showing the sea ice extent and thickness simulated by our model during the near-modern simulation. (B) A cross section of the LGM control and (C) near-modern mean water temperatures (°C). The red solid line in (A) indicates the location of the cross-section area. Ice sheet reconstructions (off-white patching) by Dyke et al. (2002). Note the open Barents Sea and Bering Strait. Warming of the intermediate depths in (C) stretches into the Arctic basin past the Fram Strait, unlike the Control integration.
Sea ice export peaks at year 4 with a maximum sea ice flux of 0.05 Sv (1980 km$^3$) transporting ice $>3$ m thick at 0.05 ms$^{-1}$ through the Fram Strait into the North Atlantic (Fig. 4.17a). Sea ice moving out of the Barents Sea is transported into the North Atlantic at 0.03 ms$^{-1}$ with a max thickness of $\sim20$ m. Together, the total sea ice flux from both outlets is similar to the Bering Strait and Barents Sea simulation, which peaks at $\sim0.07$ Sv (Fig. 4.17a). Interestingly, the entire Arctic freshwater system has a net negative flow of freshwater out of the system ($\sim233$ km$^3$yr$^{-1}$) indicating the Arctic is now a freshwater source. The continuous export of sea ice and freshwater causes a freshening of 3 psu in the Nordic Seas (Fig. 4.17b) and a 21% decrease in AMOC strength by year 100 (Fig. 4.17c). It is significant that when the Barents Ice Sheet collapses, the most notable increase in heat flux is through the Fram Strait and not the Barents Sea. The “X” indicates when the simulation switched from Control to near-modern geography boundary conditions.

Figure 4.16: Mean annual northward Fram Strait heat transport for the Control, and the near-modern simulations. The yellow line represents the northward heat transport from the Norwegian Current into Barents Sea. The dotted line is the modern northward heat flux.
through the Fram Strait into the Arctic Ocean (from Schauer 2004). After 70 years, 20 Tw (~0.02 Pw) of heat is transported into the Arctic Ocean from the North Atlantic. Simultaneously, ~1 Tw (.001 Pw) of heat is being transported northward through the Barents Sea for a total of 21 Tw of heat energy being transported into the basin.

Figure 4.17: Mean annual total sea ice flux (Fram Strait and Barents Sea) (A), Nordic Seas surface salinity (B), AMOC strength (C), and global heat transport in Petawatts (D) for the first 100 years of the near-modern simulation. At year 0, boundary conditions switch from Control to a near-modern Arctic geography.
4.4.5. Freshwater flood integration

Lastly, I examined the break-up and mobilization of Arctic sea ice to a glacial outburst flood from Lake Agassiz draining into the western Arctic Ocean via the Mackenzie River. It has previously been suggested that during the Younger Dryas cold episode, sea ice over the Arctic may have been mobilized by a meltwater outburst flood from this region (Not and Hillaire-Marcel, 2012). To more accurately simulate the interaction between sea ice and glacial meltwater in the marine environment, we performed a series of numerical simulations at a higher, eddy-permitting, spatial resolution and released 5 Sv of meltwater at the mouth of the Mackenzie River for 1 year. In addition, I analyzed areas of deep convection defined by very deep mixed layer depths. I find deep convection occurs in the Labrador Sea and the Greenland Sea, which are observed in the modern North Atlantic (Marshall, 1999) (Figure 4.18).
Figure 4.18: The location of sites of deep open ocean convection in the North Atlantic in our high-resolution model. In agreement with observations, the main sites of deep convection (and North Atlantic deep-water production) are found in the centers of both the Labrador and Nordic Seas. Here, the models mixed layer depth is sued to show that the water column in these regions is homogeneous from the surface to more than 1000m depth as waters sinks from the surface to depth to form the return flow of the large-scale AMOC.

Here, we find that meltwater quickly mobilizes Arctic sea ice, such that after just 3 months, the export of ice at Fram Strait peaked at ~0.4 Sv. Ice as thick as 10m is transported from the western Arctic (Fig. 4.19) at speeds of up to 0.5 m s⁻¹, along the north coast of Arctic Canada and Greenland, and through the East Greenland Current and into the Nordic
Seas (Fig. 4.20). This flow ends in the Labrador Sea where it melts providing freshwater to areas of deep convection. The amount of freshwater released during this event is four-times the amount needed to weaken the AMOC in other climate models.

Figure 4.19: Arctic sea ice export at Fram Strait in response to a glacial outburst flood into the Arctic. The discharge of massive volume of meltwater into the Arctic ocean from the Mackenzie River, Canada, from a glacial outburst flood from Lake Agassiz caused sea ice fluxes at Fram Strait to peak at >0.4 Sv as ice 17 – 20m thick was transported into the subpolar North Atlantic.
Figure 4.20: Advection pathway of the sea ice released from the Arctic Ocean. The blue to red colors indicates the difference in sea ice thickness between the meltwater flood and Control simulations (i.e. flood – control). White areas indicate no change in ice thickness. After 112 days, the meltwater flood begins to break-up sea ice closest to Fram Strait (A). At day 158, sea ice along the edge of the Canadian Archipelago break ups and significant volumes of sea ice are transported into the subpolar North Atlantic by the East Greenland Current (B). After 257 days, the ice passes south through Denmark Strait and continues around the tip of southern Greenland into the Labrador Sea (C).

The increase in southward transport of meltwater in the East Greenland Current (EGC) increase the supply of meltwater to the Nordic Seas. After 5 months (Figure 4.21A, B), sea ice and freshwater freshen the surface of the central North Atlantic by up to 1 psu. Major freshening occurs along the eastern boundary current (EGC) between 0 and 200m depth. After 8 months (Figure 4.21C), significant freshening is evident along the eastern North Atlantic by $>5$ psu protruding to $\sim$300m depth due to deep boundary current eddies. At this interval, the surface of the entire Central North Atlantic is freshened up to 3 psu from East Greenland to the Western European Coast over the areas of modern deep ocean.
convection. The sea ice reaches the Grand Banks ~1 year after the initial onset of the sea ice and flood export just over the areas of deep convection. South of the Labrador Coast, the sea ice mixes well into the Gulf Stream and the freshening weakens considerably. Here, we see a large freshwater flood emanating from the Arctic does have the ability to export sea ice and freshwater to directly over the areas of deep ocean convection, not only freshening the surface water, but extending down into the Greenland-Norwegian-Seas >300m depth. This result suggests considerable weakening of the AMOC is possible by weakening deep convection.

Figure 4.21: The distribution of meltwater across the Nordic Seas. The orange to blue colors represents the difference in salinity (perturbation minus control) from the 5 Sv meltwater flood during 1 year of the model integration. The cross-section spans from the East Greenland coast to the European Ice Sheet across the entire Nordic Seas. After 112 days, evidence of sea surface freshening is evident in the East Greenland Current (A). Between 5 months and 1-year, enhanced freshening between 1 and 5 psu is observed across the entire basin (C) most notable in the East Greenland Current extending to ~300m depth.
4.5. Discussion

As discussed in depth in Chapter 1, the last deglaciation featured several prominent cold episodes (e.g. Younger Dryas, 8.2-ky event, Preboreal Oscillation, Intra-Allerød cold period, Older Dryas) coinciding with the release of meltwater from North American glacial lakes (Clark et al., 2001). Remarkably, however, there is little direct correlation between the duration of these cold episodes and the reconstructed volumes of meltwater discharged at their onset, despite glacial lake outburst floods frequently being advocated as the main drivers of past climate change. For example, the 8.2-ky-event was centennial in duration and cooled central Greenland by ∼2–3 °C, yet reconstructed lake levels suggest ∼10-times more meltwater was released into the ocean at the onset of this event than at the start of the Younger Dryas, yet the latter lasted over 1,000 years and cooled central Greenland by ∼7-10°C. A significant part of this disparity can now be explained if we consider that additional sources of freshwater were involved and that glacial lake outburst floods were not the sole trigger of these periods of cooling. Our model results indicate that thick stagnant ice formed in the Arctic Ocean could have been periodically exported to the North Atlantic in several ways, including changes in atmospheric circulation, sea level rise, and outburst flooding from glacial lakes. However, unlike the ‘one-time’ glacial lake outburst floods often attributed to periods of climatic cooling, the sea ice export events we describe would have occurred repeatedly during deglaciation and could have led to cumulative freshwater discharges to the Nordic Seas greatly exceeding the total volumes of freshwater stored in proglacial lakes.

It is important to understand the significance of Arctic geography and how that may have influenced sea ice export. Of the perturbation simulations I conducted, (e.g.
Atmospheric circulation, Bering Strait, Barents Sea, Holocene and freshwater flood) the sea ice and freshwater export from the atmospheric perturbation, the Bering Strait and Holocene simulations all had comparable or greater flux than what may have come from Lake Agassiz at the initiation of the Younger Drays. However, the exception was the Barents Sea simulation, where only 12% of the total freshwater thought to be from Lake Agassiz was exported into the North Atlantic. The results from the Bering Strait and Holocene simulations make it seem likely that opening of the Bering Strait plays a large role in producing an environment for moderate to large export events of freshwater in both liquid and solid form. In addition, if the Bering Strait was not opened and a large change in lower atmospheric circulation existed, either by a mid-latitude perturbation or geography, a significant flux from the Arctic can be generated producing roughly double the amount of freshwater released from Lake Agassiz. The same type of flux can be created when a vast amount of freshwater is released from the Mackenzie River into the northwestern Arctic Basin. Here, it is important to consider these mechanisms as major drivers of abrupt climate change.

The changes in ocean circulation depicted in my perturbation experiments could have been much longer than simulated because of feedbacks occurring with the atmosphere and ocean that are not simulated in our simulation due to forcing of the model atmosphere. In reality, increased sea ice cover would help perpetuate cold conditions by isolating the atmosphere from the ocean. This could create a positive feedback by increasing sea ice cover throughout the subpolar North Atlantic. This ice cover may last for several centuries until the ice melted naturally, or the release of heat accumulated in the subsurface of the North Atlantic (Li et al., 2010). Nevertheless, as this mechanism does not result in any
global sea level rise, it offers a tantalizing explanation as to why there was little, or no, change in sea level during the Younger Dryas. While it is unknown whether there were earlier episodes involving similar sea ice discharge events from the Arctic Ocean, we note that the Bering Strait was also probably breached during MIS 3 due to higher sea-level, the Barents/Kara Ice Sheet was absent during MIS 3 (Hughes et al., 2016), and other flood events from North America or Siberia may have periodically helped to mobilize paleocrystic ice from within the Arctic Ocean, leading to other disruptions of the AMOC circulation.

My results suggest that a collapse of the Barents Ice Shelf or rising sea level alone may not have released enough sea ice and/or freshwater to cause a major climatic shift. Rather, this mechanism combined with either a large, land-based freshwater perturbation or a change in atmospheric/ocean circulation could be strong enough to mobilize an enormous amount of sea ice into the North Atlantic. Finally, it was shown that thick sea ice could be transported into the North Atlantic from the Arctic Basin in a variety of ways causing major sea surface freshening and weakening of the Atlantic Meridional Overturning Circulation and global heat transport.
CHAPTER 5
CONCLUSIONS AND FUTURE RECOMMENDATIONS

5.1 Introduction

During the last deglaciation (~20,000 – 6,000 yrs. BP), periods of increased freshwater discharge to the North Atlantic and Arctic Ocean often coincided with the onset of centennial-to-millennial length periods of climate cooling. It has repeatedly been hypothesized that these warm-to-cold transitions were triggered by a weakening of the Atlantic Meridional Overturning Circulation (AMOC) in response to increased freshwater discharge to the ocean, inhibiting the formation of North Atlantic Deep Water (NADW), the main water mass modulating the strength of this large-scale overturning cell (Bond and Lotti, 1995; Broecker, 1994; Broecker et al., 1989; Clarke et al., 2004; Manabe and Stouffer, 1997. In this dissertation, I tested an alternative hypothesis to the classic glacial meltwater discharge idea by investigating whether the growth and export of thick, multiyear, sea ice out of the Arctic to the North Atlantic could have weaken the strength of the AMOC. Here, we used the MITgcm to simulate several mechanisms that may have increased sea ice export into the North Atlantic during the Last Deglaciation: i) opening of the Bering Strait and Barents Sea, ii) a change in atmospheric winds to a more southerly direction in the Arctic Basin, and iii) a large freshwater discharge from the Mackenzie River into the Arctic Ocean.
5.2 Equilibrium sea ice thickness

To determine the range of sea ice thickness in our model, we used the MITgcm to determine the possible range (i.e. lower and upper bounds) for sea ice thickness and distribution in the Arctic Ocean during glacial boundary conditions by varying downward longwave radiation. To account for radiational differences, we ran several simulations, each experiment with varying degrees of downward longwave radiation (e.g. -20, -10, +10, +20Wm$^{-2}$) creating a precessional change between the equator and the poles. The change in radiation produced a large spread of mean sea ice thickness ranging from 32m to 55m indicating the sea ice is exceptionally sensitive to incident radiation at the surface. Out of these results, I picked the simulation which most resembled other LGM modelling studies which happened to be the +20Wm$^{-2}$ radiation experiment.

In my model, the glacial Arctic was extremely cold and lacked many of the modern features that allow heat to enter the basin. The AMOC is weakened and large northern hemisphere ice sheets completely isolate the Arctic Basin from the rest of the world’s oceans limiting the amount of heat that is supplied to the Arctic. Sea ice thickness in our model is strongly influenced by ocean ice heat flux (or the lack thereof) entering the Arctic Basin and without a major heat source limiting sea ice growth, sea ice can grow 10s of meters thick. These results were corroborated by the idealized model of Bitz and Lipscomb (1999), using a 1-dimensional, non-dynamic sea ice column model. Major attenuation occurs when insolation penetrates the thick sea ice surface making it difficult to raise the internal temperature of the ice, creating a positive feedback for growing thick sea ice.
5.3 Sea ice export mechanisms

Here, we use a climate model to show that the episodic break-up and mobilization of thick perennial Arctic sea ice during this time would have released considerable volumes of freshwater directly to the Nordic Seas where processes regulating large-scale climate occur. Massive sea ice export events to the North Atlantic are generated whenever the transport of sea ice is enhanced, either by changes in atmospheric circulation, rising sea level submerging the Bering land bridge, or glacial outburst floods draining into the Arctic Ocean from the Mackenzie River. In addition, several other deglacial mechanisms also exhibit some sea ice discharge when boundary conditions are changed, such as flooding of the Barents Sea.

We find that the volumes of freshwater released to the Nordic Seas during an enhancement of the transpolar drift due to increased southerly atmospheric winds, are similar to or larger than those estimated to have come from terrestrial outburst floods, including the discharge at the onset of the Younger Dryas. My simulations concluded that a change in near-surface wind strength of 7.5 ms$^{-1}$ creates a significant sea ice export at the initiation of each wind event. A change in atmospheric wind has the ability to export 5-times the amount of freshwater (as ice) released from glacial Lake Agassiz at the onset of the Younger Dryas. Thus, it is important to realize that a change in atmospheric forcing could have happened multiple times in a short period owing to an abrupt climate shift. It is likely that the Barents and the Laurentide Ice Sheets may have influenced the pressure gradient between the Arctic and the North Atlantic. Additionally, enhanced katabatic winds coming off the Greenland Ice Sheet during the winter months could have enhanced sea ice export out of the Arctic basin (http://polarmet.osu.edu/PolarMet/paleonwp.php), as shown
by the Polar MM5 model. These two mechanisms working in concert may have increased
wind speeds and sea ice export some time during the LGM.

During the wind perturbation experiments, although the change in AMOC is only
~7% reduction in strength, it is important to stress this is only a 0.1 Sv export for 5 years,
which after, the wind was reverted back to pre-perturbation conditions and the AMOC
recovered. This result is consistent with what a numerical model would produce with a
freshwater perturbation of only 5 years. Typical freshwater hosing experiments release 0.1
Sv of freshwater over a 100 years period which would continue to weaken the AMOC over
that amount of time as the perturbation was applied. Once this perturbation is removed, the
AMOC recovers. This result was corroborated in a study by Stouffer et al. (2006) using 14
of the Climate Model Intercomparison Model/Paleo-Modeling Intercomparison Project
(CMIP/PMIP). Their results show a reduction of the AMOC of 9% to 62% relative to the
control. Thus, a 7% decrease in this short amount of time is very significant.

Other experiments, such as opening of the Bering Strait and Barents Sea produce a
large sea ice export and a decrease in AMOC strength. My simulations showed a collapse
of the Barents Ice Shelf alone may not have released enough sea ice and/or freshwater to
cause a major climate shift. Rather, this mechanism combined with flooding of the Bering
Strait, a land-based freshwater perturbation or a change in atmospheric/oceanic circulation,
is strong enough to mobilize an enormous amount of sea ice and transport it into the North
Atlantic. I observed opening the Bering Strait in either simulation helped increase heat
transport through the Fram Strait coinciding with an increase in ocean-ice heat flux. This
would allow for the breakup and mobilization of sea ice into the Nordic Seas along with
the liquid freshwater contained in the Arctic basin. The simulation that produced the most
similar results to the open Bering Strait simulation was the near-modern simulation, that includes an open Barents Sea. Here, sea ice is exported mainly from the Fram Strait and Barents Sea while freshwater is imported via the Barents Sea. Fram Strait heat flux almost reaches modern day levels.

It is imperative not to lose sight on the fact my simulations showed that a massive release of freshwater and ice from the Arctic can be created by opening the Bering Strait or increasing transpolar drift. The timescales that my simulations were able to create the exports form a good explanation as to why the initiation of the Younger Dryas was so abrupt. Finally, it was shown thick sea ice could be transported into the North Atlantic from the Arctic Basin in a variety of ways causing major sea surface freshening and weakening of the Atlantic Meridional Overturning Circulation and global heat transport. Our results provide the first evidence that the storage and release of Arctic sea ice helped drive deglacial climate change.

5.3 Freshwater flood

Tarasov and Peltier (2005) also showed (using a glacial systems model) that during the Younger Dryas episode, glacial meltwater from the North American ice sheets may have drained into the Arctic Ocean via the Mackenzie River Valley (Keigwin et al., 2018). Their claims are supported by evidence found in marine sediment cores from the central Arctic basin showing a freshwater flux emanating from the Arctic coincident with their study period (Not and Hillaire-Marcel, 2012). A freshwater flux of this magnitude released from the Mackenzie River might flush sea ice and freshwater stored in the Arctic Ocean through the Fram Strait. The transport of freshwater from the Arctic to the North Atlantic may have been large enough to weaken the AMOC by disrupting NADW overturning (Condron and
Winsor, 2012). Here, we show that if a large freshwater flux did originate from the Mackenzie River, it would have transported 0.4 Sv of freshwater into the Nordic Seas in less than 6 months. This is enough freshwater to weaken the AMOC (Manabe and Stouffer, 1997) and coincide with the timeline for abrupt climate change seen with climate cooling episodes such as the Younger Dryas.

A majority of the deep ocean convection takes place in the Nordic Seas and south of Greenland. When a 5 Sv flood is released from the northwest Arctic basin, I find these areas of deep convection are flooded with freshwater from melting sea ice and liquid freshwater. Although these initial results have been discussed in the scientific community for decades, my simulations are the first of their kind showing a large meltwater flood can mobilize sea ice and export it to the North Atlantic over sites of deep convection. The advantage these results have over other studies is that the freshening is not artificially created, such as hosing experiments.

5.4 Future Work

Look back at the research I performed for this thesis, I was able to develop, test and show that it is possible to create a major sea ice export event created by multiple mechanisms. However, future work is an important way of improving your research in many different ways. Here, future research might be considered creating a simulation with a Younger Dryas atmosphere allowing us to tailor the simulations exactly to our time period. Similarly, using the same NCAR CESM atmospheric conditions, simulating time slices from the LGM to modern day and incorporating the appropriate gateway and bathymetry changes like a trace study would be beneficial to studying other boundary
conditions associated with a YD to modern climate and how that could affect sea ice transport.

Additional studies using a fully-coupled model of intermediate complexity, for example, the Community Climate System Model (CCSM) version 4.0 (Brady et al., 2013). Testing the Bering Strait open simulation on a high-resolution grid in a fully-coupled system could provide better insight into the effect on the AMOC and how that would change northern hemisphere climate. By using a fully coupled model, the atmosphere and ocean can be directly compared. A study like this has been well documented and the results are presented in Hu et al. (2010, 2012, 2014) using the Community Climate System Model (CCSM) version 3.0. However, this study has not been reproduced during glacial-like conditions. Using the MITgcm at 1/6th degree resolution with high-resolution CCSM v4.0 atmospheric boundary conditions is an interesting way to analyze Bering Strait through flow at a finer scale while isolating the ocean from the atmosphere. However, it is important to note the limitations of this kind of model setup. Testing the sea ice export mechanisms in this thesis in a fully coupled model can cause oceanic and atmospheric instabilities inhibiting the model from completing the simulation. Using the current setup with a prescribed atmosphere, I am able to test different export mechanisms alleviating many model instabilities.

Additional experiments testing climate sensitivity to freshwater discharge from the European side of the North Atlantic would allow an assessment of the sensitivity of the AMOC to increased sea ice export from the Fennoscandian Ice Sheet (FSIS). Using a high-resolution iceberg and prescribing icebergs from the Norwegian Channel Ice Stream (Sejrup et al., 2003) (NCIS) can test whether ice would travel and melt over the important
areas of deep convection in the Nordic Seas. This same experimental setup could be initiated from the southern edge of the central Arctic Basin ice margin in the Fram Strait to simulate large iceberg-like pieces of sea ice similar to the ice mentioned in Moore (2005). A study testing other freshwater sources during glacial conditions, such as sources originating from Europe and Siberia on a high-resolution may shed light on other mechanisms during deglaciation that could cause abrupt climate change.

Improvements in model design, horizontal grid resolution and computational resources are progressing quickly with improvements in technology. Nonetheless, the effects of sea ice export on the global ocean’s is spatially and dynamically complex and the MITgcm is an effective tool for analyzing these factors (Marshall et al., 1997).

This thesis has demonstrated that large amounts of freshwater as ice can be transported from the Arctic to the Nordic Seas creating a large surface freshening. Better yet, this experiment was conducted without the usual “hosing experiment” that is typically used for these types of analyses. For abrupt climate change events such as the Younger Dryas, Arctic sea ice is normally not considered as a main freshwater source. Now, it has been shown that freshwater stored in sea ice can induced AMOC weakening and a reduction in northerly heat advection causing an abrupt climate cooling.
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