

Understanding Glacial Inception in Northeast Canada with a High-Resolution Regional Climate Model

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ABSTRACT

Over the past 0.8 million years, 100 kyr ice ages have dominated Earth's climate. Interglacials were brief, sometimes lasting only a few thousand years. Currently, state-of-the-art global climate models (GCMs) have difficulty simulating glacial inception, due in part to their coarse horizontal resolution and neglect of ice flow dynamics. In this thesis, I address the inception problem using a high-resolution regional atmospheric model and an ice flow model. I explore three stages and corresponding mechanisms for glacial inception: mountain glacier growth, downslope ice flow from these mountain glaciers, and the atmospheric circulation response to ice sheet growth.

I use the Weather Research and Forecasting model (WRF), a cloud-resolving atmospheric model capable of a realistic simulation of the regional mountain climate and therefore of surface mass balance. I focus this study on the mountain glaciers of Canada's Baffin Island, where geologic evidence indicates the last inception occurred at 115 kya. I examine the sensitivity of mountain glaciers to insolation (Milankovitch forcing), topography, and meteorology, all part of the first stage of glacial inception. In addition, I investigate the radiative effects of clouds which are resolved by WRF rather than being parameterized.

I also examine the possible role of ice flow dynamics missing in GCMs and the interaction of expanding ice cover and the atmosphere. Using the WRF-derived mountain glacier mass balance, I run the ice sheet model based on the shallow-ice approximation, capturing the ice flow that may be

critical to the spread of ice sheets away from mountain ice caps. The new ice surface elevation and extent are used as inputs in WRF. Through this iterated asynchronous coupling, I investigate how the regional climate and large-scale circulation respond to ice sheet changes, which are involved in the second and third stages of inception.

I find that ice would accumulate on mountain glaciers with anomalously cold meteorology, 115 kya, insolation, and realistic topography. Ice then proceeds to flow downslope, expanding the size of the ice caps. However, the changing surface elevation induces an atmospheric response that warms the inception site, causing ice sheet growth to stagnate in the model, and therefore preventing full inception from occurring.

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1

Introduction

For the past 0.8 million years, Earth's climate has been characterized by 100,000 year (100 kyr) glacial cycles (Petit et al., 1999; Augustin et al., 2004; Delmonte et al., 2004). The 100 kyr cycles began at the Mid-Pleistocene transition (Lisiecki and Raymo, 2005; Clark et al., 2006), before which glacial cycles occurred every 41 kyr. Ice ages lasted tens of thousands of years, during which the climate was colder by $\sim 15^{\circ}\text{C}$, ice sheets spanned continents, and sea level was as much as $\sim 130\text{ m}$ lower than today (Lambeck and Chappell, 2001; Bintanja et al., 2005). Eventually, the climate warmed

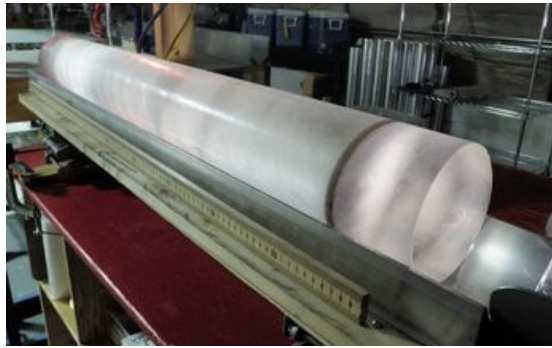


Figure 1.1: An Ice Core from the West Antarctic Ice Sheet Divide Project. The dark band represents volcanic ash settling on the ice ~21 kya. Courtesy: National Science Foundation and Heidi Roop.

abruptly and the ice sheets retreated, yielding an interglacial with temperatures similar to or warmer than present day (Kukla et al., 2002; Jouzel et al., 2007). Interglacials were brief, lasting only a few thousand years before the climate cooled again and ice sheets grew anew. The initial growth of ice sheets during an interglacial leading into an ice age is known as glacial inception.

Earth's climatic history has been recorded in ice and sediments around the globe, and by drilling through those layers of ice and sediment, cores (as pictured in Figure 1.1) are collected and analyzed for isotopes, gases, dust particles, and more. Isotopes in these cores, such as $\delta^{18}O$, serve as analogues or proxies for global temperature, ice volume, and sea level (Petit et al., 1999; Augustin et al., 2004; Shackleton, 1987). Ice volume and sea level are directly connected (Lambeck and Chappell, 2001; Camoin et al., 2001); when ice volume increases during glacial inception, sea level falls (and vice versa for deglaciation). Deep-sea sediment cores are used to reconstruct millions of years of climate history (Mix et al., 1995; Lisiecki and Raymo, 2005; Raymo et al., 1997), capturing the 41 kyr glacial cycles, the Mid-Pleistocene transition, and the recent 100 kyr glacial cycles. Ice and sediment cores contain global climate signals, but further analysis of the composition of the cores can reveal regional climate variability. Pollen found in cores (Tzedakis et al., 1997; Fréchette et al., 2006) identify regional vegetation changes. Most relevant for this thesis are the glacial deposits that appear in sediment cores.

These deposits are composed of rock debris that are too large to have been windblown; they must have been moved by ice. Using both glacial deposits and knowledge of landforms can help refine when and where ice appeared over land (Clark et al., 1993; Hughes et al., 2013).

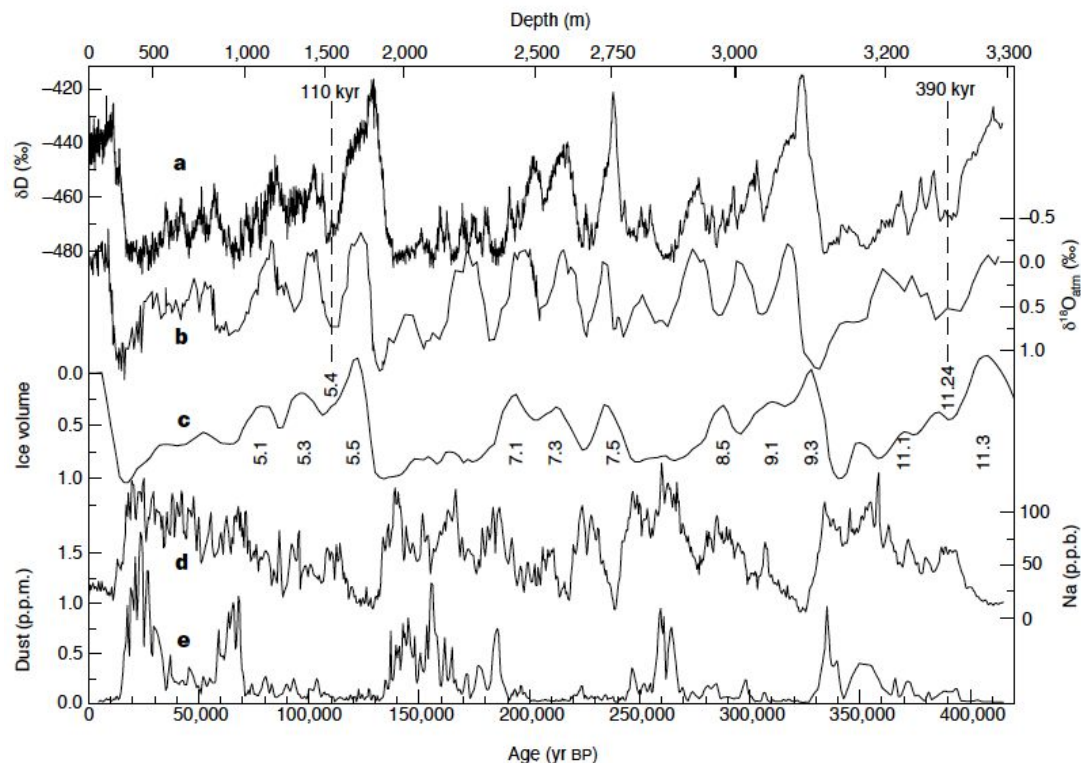


Figure 1.2: Glacial cycles over the past 400,000 years from the Vostok Ice Core from Petit et al. (1999). Note the saw-tooth pattern $\delta^{18}O$ and δD and the associated ice volume from Bassinot et al. (1994). The atmospheric dust is also pictured.

An example of the reconstructed climate record is presented Figure 1.2. The $\delta^{18}O$ and δD were recorded in the Vostok Ice Core for the past 400,000 years as proxies for temperature. The global ice volume was calculated based on $\delta^{18}O$ from Bassinot et al. (1994). Following the last interglacial, the Eemian, we see a significant build up of ice beginning about 115 thousand years ago (115 kya). The major ice sheets covered North America, Greenland, Eurasia, and Antarctica. There were multi-

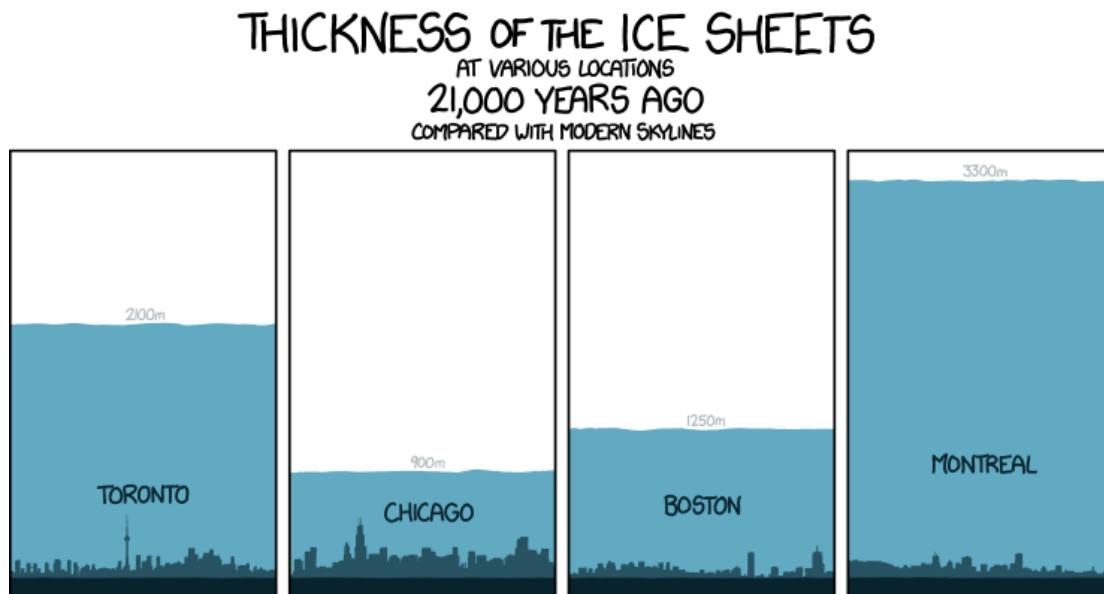


Figure 1.3: A comparison of modern skylines and the Laurentide Ice Sheet at LGM based on based on Dyke et al. (2002). Courtesy of Randall Monroe and xkcd.com/1225/

ple advances and retreats of ice sheets leading up to the Last Glacial Maximum (LGM, Clark et al., 2009) at about 21 kya, when the ice sheets reached their largest extent and sea level was at a minimum (Camoin et al., 2001). Rapid deglaciation followed at around 20 kya and 15 kya for the Northern and Southern Hemispheres respectively (Clark et al., 2009), leading into the current interglacial, the Holocene. This sawtooth shape pattern is a characteristic of Earth's glacial history (Broecker and Donk, 1970).

During the last ice age, the Laurentide Ice Sheet (the formation of which is the focus of this thesis) covered Northern Canada with a dome centered on the Hudson Bay area (Ives, 1978). During its build up, sea level is thought to have dropped rapidly. Waelbroeck et al. (2002) and Ganopolski et al. (2010) estimated sea level lowered 50 m in 10 kyrs but Lambeck and Chappell (2001) and Creveling et al. (2017) estimated a more modest 20 m and 10 m, respectively. The Laurentide has been thought to be the primary cause of the initial drop in sea level (Kleman et al., 2013), while the ice sheets in

Greenland, Antarctica, and Eurasia contributed later to the global ice volume increase. The southern margin of the Laurentide advanced into and retreated from Illinois multiple times during the last ice age (Clark et al., 1993). At its maximum extent during the LGM, the southernmost part of the ice sheet terminated at about 40°N in Illinois, just south of the great lakes (Willman et al., 1970; Ives, 1978; Kleman et al., 2010). In fact, Boston (Dyke et al., 2002) was covered by the Laurentide ice sheet until about 14 kya (Borns, 1973), and Walden Pond was formed by these ice sheets. The Laurentide Ice Sheet at the LGM was a major topographic feature, which a comparison to modern day skylines presented in Figure 1.3 helps put in perspective.

When asking how this colossal ice sheet formed, we must begin with where. From sediment cores off the coast of Baffin Island, glacial deposits are dated to about 115 kya, which coincides with the timing of the last glacial inception. In western Canada, glacial deposits do not appear until ~100 kya, meaning that glaciation occurred first in the east and later in the west (Clark et al., 1993). Initially, it was thought that the Laurentide ice sheet originated on the plateaus of Baffin Island and that ice flow from Baffin Island alone created a continental size ice sheet (Flint, 1952). This scenario has since been revised to that of small ice caps growing at high elevations on Baffin Island, causing regional climate cooling, and allowing snow to accumulate in areas previously unglaciated, such as the Keewatin and Labrador regions (Fulton and Prest, 1987). This process of ice flow and climatic cooling eventually allowed ice to coalesce into the Laurentide Ice Sheet (Ives et al., 1975).

Baffin Island has long been identified as a possible site for glacial inception:

The point I wish chiefly to make is that the conditions on Baffin Island are wonderfully near those which produce glaciation.

-Tarr (1897)

Baffin Island contains the Barnes and Penny ice caps, which are the hundreds of meters thick remnants of the Laurentide ice sheet (Andrews et al., 1972) that survived due to the ideal combination of

precipitation, high latitude, and high plateaus. Baffin Island's topography decreases in height gradually towards the west, allowing ice to flow easily from east to west accumulating ice (Ives et al., 1975; Williams, 1978; Oerlemans, 2002). It has long been noted that the glaciers of the Baffin Island region are associated with low pressure regions and winter cyclones (Hobbs, 1910; Keegan, 1958), normally associated with the significant precipitation needed for ice growth (Barry et al., 1975). The ice caps receive significant precipitation (Fisher et al., 1998), unlike the surrounding region, possibly creating a micro climate on Baffin Island (Zdanowicz et al., 2012; Andrews and Miller, 1972). Baffin Island is very sensitive to minor climatic changes, a possible characteristic of its mountain glaciers (Oerlemans, 1989). Around 300 years ago, the Little Ice Age left clues on Baffin Island of how sensitive the region is to cooling of about 1.5°C (Bradley and Jonest, 1993). Williams (1978) looked at the lichen-kill zones, which indicate that the snow line was about 300 m lower than at present day. Ives (1962) showed that 2% of the area north of the Barnes Ice Cap is glaciated currently, but in the recent past up to 70% was glaciated. Thus, Baffin Island appears an ideal place for inception to occur.

Now we must address the question of how Baffin Island would glaciade, and how ice growth would be maintained for thousands of years. A critical factor in causing glacial inception is thought to be changes in Earth's orbital parameters, i.e., obliquity, eccentricity, and precession. Earth's orbital parameter configuration affects the amount of sunlight, or insolation, that the Earth receives and thus they are an external influence on climatic changes. Less insolation at high latitudes during the summer would cause less melting, leading to ice sheet growth. This effect of orbital parameters on glacial cycles is the basis for the Milankovitch theory of glacial cycles (Milankovitch, 1941).

Each orbital parameter has an effect on the amount of insolation received by the Earth. The obliquity is the angle of the tilt of Earth's axis of rotation, ranging from 22 to 24 degrees and controlling how much sunlight the poles receive. Eccentricity is a measure of how much the orbit deviates from circular, controlling the seasonality in the Earth's distance from the sun. The solar constant depends on this distance, thus as Paillard (2001) noted the improperly defined solar constant changes

with eccentricity. Precession is the orientation of Earth's rotational axis. It affects the seasonality, and in particular the proximity to the sun during the critical Northern Hemisphere summer. Eccentricity and precession are linked: precession would not have a climatic effect in a perfectly circular orbit. These astronomically driven parameters vary over thousands of years and can be calculated using orbital mechanics (Berger and Loutre, 1991; Laskar et al., 1993), allowing the insolation to be known for millions of years, both past and future. In fact, before the identification of the strength of anthropogenic emissions, there were predictions of Earth going into an ice age in a few thousand years due to orbital variations (Kukla et al., 1981).

The orbital parameters have different periods of oscillation that appear in the climate record. The periods of precession, obliquity, and eccentricity, are ~ 20 kyr, ~ 40 kyr, and ~ 100 kyr. A spectral analysis of ice and sediment records reveals frequency peaks that coincide with those of orbital parameters (Hays et al., 1976). The glacial cycles occur every 100 kyrs, which is the period of eccentricity, but Imbrie et al. (1992) noted, that “the 100-kyr radiation cycle seems much too small and its phase too late to drive the 100 kyr glaciation cycle.” There appears to be a linear response of $\delta^{18}\text{O}$ to obliquity and precession (Shackleton, 2000). It has been suggested that orbital inclination or obliquity could be the driver of glacial cycles (Muller and MacDonald, 1997) with terminations occurring after 2 or 3 obliquity cycles, between 80 and 120 kyrs (Huybers and Wunsch, 2005).

It is useful to find an insolation-related quantity that is closely related with temperature and mass balance, and Huybers (2006) considered the integrated summer insolation, over a specified insolation threshold, as a measure of summer insolation that leads to melting. During the last glacial inception at 115 kya a minimum in integrated insolation indeed occurred. Leading into interglacials or warm periods, insolation increased coinciding with high obliquity (Kukla et al., 1981).

The timing of glacial cycles is thought to be set by nonlinear phase locking to Milankovitch insolation variations (Tziperman et al., 2006; Hyde and Peltier, 1987; Crucifix, 2013). In this sense, orbital parameters act as a pacemaker for glacial cycles. This requires the glacial cycle to be a non-

linear oscillation, and various climate processes, such as sea ice (Gildor and Tziperman, 2000) can provide the needed nonlinearity. One result of the nonlinear phase locking to Milankovitch forcing is that successfully simulating the timing of glacial inception by a given simple model does not necessarily mean the underlying physical processes are indeed the ones dominating the glaciation physics (Tziperman et al., 2006). Though when and where glaciation began is well documented, state-of-the-art global climate models (GCMs) have difficulty simulating the beginning of ice ages, which has become known as “the inception problem”. Models configured with the correct insolation from 115 kya still failed to accumulate perennial snow cover (Rind et al., 1989). Some models did get snow accumulation, but in areas that did not match observations (Dong and Valdes, 1995; Meissner et al., 2003). Recently Otieno and Bromwich (2009) extensively investigated the glacial inception problem with a land model. They found that meteorology is important to snow accumulation, in particular cold summers and wet winters. Combining the right meteorology and 115 kya insolation did not yield inception, though. Only artificially lowering the temperature allowed significant snow accumulation in areas that were previously unglaciated. Through model failures we learn that changes in climate from orbital parameters must be amplified by internal mechanisms. Other climate factors that have been considered are CO_2 , dust, sea surface temperatures, sea ice, ocean circulation, vegetation, meteorology, and atmosphere-ice interactions, as is discussed next.

1.1 SUGGESTED INCEPTION MECHANISMS

We turn our attention to internal climate mechanisms that can lead to inception. It has been suggested that CO_2 feedbacks (Petit et al., 1999) can participate in glacial inception, given the obvious correlation of temperature and CO_2 in ice cores (Cuffey and Vimeux, 2001). (Shackleton, 2000) posited that CO_2 leads ice volume changes, which Petit et al. (1999) noted as well. Fischer et al. (1999) argued that CO_2 lags changes in temperature, although they were concerned with deglaciation.

tion rather than inception. We also note that the uncertainty in dating gases in ice cores is at least 1000 years. Colder oceans can remove CO_2 from the atmosphere (Toggweiler et al., 2006) and therefore any initial cooling would lead to a CO_2 decrease and to an amplification of the inception, as has been explored in some modeling studies (Vavrus et al., 2011; Ganopolski and Calov, 2011; Vettoretti and Peltier, 2004; Bonelli et al., 2009).

Dust (Lambert et al., 2008; Delmonte et al., 2004) as seen in the Vostok core (Figure 1.1), can have multiple effects on climate, reflecting shortwave radiation and causing cooling, as is reflected in the correlation of dust and temperature in the ice cores (Lambert et al., 2008). Farrell and Abbot (2012) argued that dust impacts the hydrological cycle in a nonlinear manner: dust increases in the atmosphere causing an increase in precipitation, until a threshold when precipitation decreases. Though dust bands from volcanoes are noted in Figure 1.1, volcanoes are not the main source of dust. Terrestrial dust from strengthened westerlies may contribute greatly to cooling during glacial cycles (Basile et al., 1997). Dust can also decrease the snow albedo when it is deposited out of the atmosphere (Peltier and Marshall, 1995). Calov et al. (2005) simulated the sea level for the past glacial cycle, and they found that including dust was necessary to accurately recreate sea levels from sediment cores. Without the inclusion of dust, too much ice accumulates. Therefore, dust can impact two important aspects of glacial inception: temperature and precipitation. Interplanetary dust (Farley and Patterson, 1995) was also suggested to block solar insolation.

Pollen records (Fréchette et al., 2006) indicate a highly vegetated Arctic during the last interglacial, and the treeline moved southward during the last glacial inception. Thus, models using present day vegetation often fail to allow for sufficient snow growth (de Noblet et al., 1996). The use of dynamic vegetation in climate models, led to a simulation of the observed southward shift of the treeline and was critical to ice build up (Harvey, 1989; Crucifix and Loutre, 2002; Wang et al., 2005; Goñi et al., 2005; Mysak, 2008). The transition from forests to tundra at high latitudes during inception raises albedo and lowers temperature (Otterman et al., 1984), which would lead to snow

survival during the melting season.

Sea surface temperatures (SSTs), sea ice, and precipitation are all tied closely together through a temperature precipitation feedback (Källén et al., 1979; Ghil, 1994; Gildor and Tziperman, 2000, 2001; Li et al., 2005) which may play an important role during inception. Warmer oceans imply less sea ice cover and more evaporation and precipitation, which can increase accumulation and lead to inception (Stokes, 1955; Gildor and Tziperman, 2000). Further pollen-based evidence that the ocean was warm throughout inception was provided by Goñi et al. (2005), who found that SSTs remained warm until 105 kya, when substantial ice sheets had formed. The modeling study of Löfverström et al. (2014) similarly argued that sea ice did not extend significantly into the Atlantic until 80 kya. Ice cores in Greenland indicate that precipitation rates during the LGM (when the climate was much colder and sea ice cover extensive) were four times lower than present day (Cuffey and Clow, 1997).

Ocean circulation transports heat poleward, warming high latitude SSTs, affecting the atmosphere, and possibly inception (Broecker and Denton, 1989). Consistent with this idea, Böhm et al. (2015) simulated a strong overturning circulation up until the LGM. On the other hand, some simulations have suggested that high latitude SSTs may decrease during glacial inception. This decrease allowed sea ice extent to grow larger (Meissner et al., 2003), leading to cooling over land which aided glacial inception in their simulation (Yoshimori et al., 2002). These studies argued that cooling was more important to inception than increased accumulation. Other studies have found that precipitation increases even with colder SSTs Kageyama and Valdes (2000); Khodri et al. (2001); Wang and Mysak (2002).

The atmospheric circulation is a necessary component of the glacial inception process, through temperature and moisture fluxes. As noted before, Baffin Island has long been associated with low pressure systems and substantial precipitation (Hobbs, 1910; Zdanowicz et al., 2012). Ideal conditions for glacial inception involve winters with anomalously high precipitation and cold summers.

Bromwich et al. (2002) analyzed the present day circulation over Baffin Island, finding that the necessary wet winters and cold summers occur, but are rare. They classified only the year 1986 to represent meteorology that could lead to glaciation. A cyclone occurred during this year over Baffin Bay, leading to a positive mass balance Gardner and Sharp (2007). Generally cold summers on Baffin Island occurred when the flow came from the north (Bromwich et al., 2002). The model studies of Jackson and Broccoli (2003) and Kaspar et al. (2007) noted that storms increased in frequency and intensity during glacial inception. Thus, the increased precipitation deemed necessary (Barry et al., 1975) is feasible. Andrews et al. (1972) argued that precipitation on Baffin Island due to warm winters was more important than cool summer temperatures for a positive mass balance. In contrast, Otieno and Bromwich (2009) argued summer temperatures are more important.

As the ice caps begin to expand, the ice sheet interacts with the atmosphere, further influencing its own growth. Atmospheric circulation may change in response to the ice surface elevation (Roe and Lindzen, 2001a). However, the effects of mountains on the regional climate and large-scale circulation depends on the mean overlying flow and mountain size (Chen and Lin, 2005; Cook and Held, 1988). Topographic changes may have a stronger impact through wave superposition (Jackson, 2000), which were suggested to lead to rapid climate changes such as Dansgaard Oeschger events. The change in circulation has been noted to enhance glaciation by cooling the inception region (Roe and Lindzen, 2001b; Liakka et al., 2012). The European ice sheet may have influenced the formation of the Laurentide (Beghin et al., 2014), and Roe and Lindzen (2001b) noted that changes in the stationary wave pattern forced by one ice sheet would influence the downstream climate and affect the other ice sheet. Kubatzki et al. (2006) found that the presence of ice in Greenland did not significantly influence inception on Baffin Island.

Precipitation can be both suppressed and triggered by topography (Roe, 2005; Yamagishi et al., 2005). For instance, precipitation decreases at high elevations due to Clausius-Clapeyron, leading to the “elevation-dessert effect” (Weertman, 1976; Oerlemans, 1989). Antarctica is a present day ex-

ample; the dearth of precipitation allows it to be classified as a desert in some places. Inspired by Atnarctica, Budd and Smith (1982) included the elevation desert effect in an ice sheet model, finding the effect to be significant. For the Laurentide, the inclusion of the elevation-desert effect has proved critical for an accurate simulation of ice sheet growth (Marshall et al., 2002; Abe-Ouchi et al., 2007). However, orography can also trigger precipitation events, particularly on the windward side of the mountains (Roe, 2005). This same effect is seen on the margins of ice sheets (Bromwich et al., 2005).

The simulation of glacial inception may improve by using higher resolution simulations. Coarse resolution lowers critical topography significantly, artificially increasing temperatures and suppressing orographic precipitation. The importance of using realistic topography has long been noted by Marshall and Clarke (1999); Wang and Mysak (2002); Vettoretti and Peltier (2003). A recent success in simulating glacial inception (Jochum et al., 2012) involved using a higher resolution, but the model used had a significant cold bias of 5 degrees over Baffin Island, which may partially account for the successful simulation of inception. Vavrus et al. (2011) compared GCM simulations of inception at two different resolutions, finding that resolution increases favored inception by allowing for dynamical responses to other important climate factors such as greenhouse gases. Note that finer resolution GCM simulations are still coarse, at about 1 degree, which still lowers topography by 1000 m in critical areas such as Baffin Island.

1.2 THIS THESIS

I will think of glacial inception as consisting of three stages and corresponding mechanisms: positive mass balance on current ice caps, downslope ice flow from successive years of accumulation, and regional climate cooling as a result of the ice expansion allowing for further snow growth. In Chapter 2, I examine the impact of topography, insolation, and meteorology on inception on Baffin Island using a high-resolution regional atmospheric model (Weather, Research and Forecasting

model, WRF Skamarock et al., 2008). By focusing on these mountain glaciers, I investigate their sensitivity to these factors which leads to the first stage of inception. This area is integral to the eventual formation of the Laurentide Ice Sheet, and the island topography is a major part of why ice grows there to begin with. GCM resolution is typically coarse, which leads to an artificially low topography of the ice caps on Baffin Island. For instance, the Penny ice cap has an elevation of 2000 m, but even at a moderately high GCM resolution, about 1 degree, the topography is lowered to 400 m. The use of a high resolution regional atmospheric model allows a correct representation of the topography.

During glacial inception at 115 kya, the integrated insolation was at a minimum. I examine the effect of minimum insolation on glacier mass balances, and compare it to effect of the last maximum integrated insolation which occurred at 128 kya. I choose boundary conditions for the regional model based on Bromwich et al. (2002) and Otieno and Bromwich (2009), who identified cool summers and increased precipitation as important to inception. I investigate glacier mass balances using average, anomalously cold, and anomalously warm boundary conditions from present-day climatology. In the Arctic today, clouds play a significant role in the radiation budget during the summer (Zhao and Garrett, 2015). Jochum et al. (2012) found that clouds were a negative feedback on inception, and with a high resolution study I can investigate the cloud radiative effects without cloud parameterizations.

The second phase of glacial inception occurs when snow continues to accumulate on mountain glaciers and ice flows to lower altitudes. I implicitly investigate ice growth through an analysis that integrates the mass balance downslope and considers it as function of elevation. I also test the sensitivity to land cover (snow vs tundra) to identify the strength of the snow albedo feedback.

In Chapter 3, I use an ice model to more accurately capture the flow of ice. I analyze the impact of the snow albedo feedback and elevation changes on amplifying ice growth. I created a simple ice flow model, based on the shallow ice approximation. This ice model is asynchronously coupled to WRF, and is forced with the mass balance from WRF. I use the ice cover from the ice model as an

input WRF to calculate a revised mass balance, and repeat. As iterations progress, I examine how the regional climate changes and inception advances. Generally, as ice accumulates in the ice flow model, the surface elevation increases. This topography change impacts the atmospheric circulation. This reorganization of the circulation, in turn, affects the mass balance on the ice sheet. Other studies argued that an atmospheric circulation that amplifies the net accumulation is necessary for glacial inception and I examine this on the much smaller scale of the Baffin Island relevant to the initial inception process. I additionally test the sensitivity of the ice model to ice viscosity, realistic mass balances, and calving. I conclude my findings in Chapter 4.

2

The Role of Insolation, Meteorology, and Topography

2.1 INTRODUCTION

Since the mid-Pleistocene transition ~ 800 thousand years ago (kya), 100,000 year glacial cycles have dominated Earth’s climate variability (Petit et al., 1999). The transition into an ice age from an interglacial period (similar to today’s climate) is known as glacial inception. The area surrounding Hudson Bay, including Baffin Island, Labrador-Ungava and Keewatin has been identified as a likely inception site based on sediment cores (Clark et al., 1993), arguments pertaining to the orography (Oerlemans, 2002; Ives et al., 1975), and previous studies (Williams, 1979; Kleman et al., 2002; Otieno and Bromwich, 2009). According to Clark et al. (1993), the growth of an ice sheet in North America from Baffin Island westward occurred over 20,000 years. Ice volume reconstructions (Petit et al., 1999) confirm the timescale of this glacial inception process. We are interested in exploring the conditions necessary for the initial high-elevation mountain glacier expansion into the low-lying surrounding area, which occurs on a much shorter timescale. Using a high-resolution regional atmospheric model, we focus on the roles of complex topography and clouds on glacial inception.

According to Milankovitch theory, the driver or pacemaker of glacial cycles is changes in Earth’s orbital parameters (Milankovitch, 1941; Hays et al., 1976), perhaps by inducing wet winters and cool summers (Bromwich et al., 2002). A more updated view is that “pacemaking” is due to nonlinear phase locking of a self-sustained glacial cycle to the insolation forcing which therefore sets the timing of inceptions and terminations (Hyde and Peltier, 1987; Gildor and Tziperman, 2000; Tziperman et al., 2006; Crucifix, 2013). It has been hypothesized that weaker summer insolation (in particular weaker integrated insolation, Huybers, 2006) leads to decreased melting and thus glacial inception. Historically, state-of-the-art global climate models (GCMs) have difficulty simulating the transition from interglacial to glaciated due to Milankovitch forcing alone (Rind et al., 1989).

A successful glacial inception simulation in a model without ice flow is thought to include the expansion of perennial snow into initially non-glaciaded regions. Early GCM studies either failed to

simulate perennial snow cover or the pattern of snow expansion did not match geological records (Dong and Valdes, 1995; Meissner et al., 2003), accumulating snow outside of the Hudson Bay region. The introduction of SSTs from the time of the last glacial inception and ocean feedbacks to GCMs has been shown to lead to increased poleward moisture transport and snow accumulation over the Laurentide region (Khodri et al., 2001; Yoshimori et al., 2002). Vettoretti and Peltier (2004) considered CO₂ and orbital parameters, obtaining snow growth northward of inception sites inferred from geologic records. Simpler models have also been used to understand the relative importance of other climate factors that could influence glacial inception. These factors include the effects of nonlinear response to Milankovitch forcing (Le-Treut and Ghil, 1983), the impact of sea ice on snow accumulation (Gildor and Tziperman, 2000), vegetation feedbacks (Crucifix and Loutre, 2002; Wang et al., 2005), dust (Lambert et al., 2008; Calov et al., 2005) – particularly its ability to switch off the hydrological cycle (Farrell and Abbot, 2012), and the role of coupled Milankovitch-meridional ocean circulation feedbacks (Timmermann et al., 2010). Otieno and Bromwich (2009) thoroughly examined and nicely summarized the inception problem using a land surface model, noting that insolation, wet winters and cold summers were all important, but insufficient for inception, which necessitated an artificial 4 degrees of cooling. Jackson and Broccoli (2003) also illustrated that increased precipitation in northeast Canada at about 115kya is distinct possibility due to increased storm activity. Most recently Jochum et al. (2012) was able to simulate inception over the Baffin Island region by using a higher GCM resolution which allowed for a more accurate representation of the topography. The importance of using realistic topography was also noted by Marshall and Clarke (1999), Wang and Mysak (2002), and Vettoretti and Peltier (2003). Although the work of Jochum et al. (2012) was a success in many ways, the topography over Baffin Island was still at least 600 m below the observed height of the mountainous terrain in the region, and there was snow accumulation in areas that may not have been glaciated. Furthermore, the model they used has a substantial cold bias in its present climate state, of up to 5°C over Baffin Island in the annual mean –

raising questions about whether they simulated inception for the right reasons.

Given the demonstrated critical role of accurate topography over the inception sites, our objective is to study the inception problem using a model that represents this topography as accurately as possible. We, therefore, use the Weather Research and Forecasting model (WRF) in a high-resolution (4 km) regional configuration over Baffin Island, focusing specifically on the Penny Ice Cap, which is one of the two large ice caps on the island today remaining from the Laurentide Ice Sheet. Ice cores from the Penny Ice Cap (Fisher et al., 1998) illustrate that accumulation rates on the ice cap can range from 0.18 to 0.36 m year⁻¹, which is something our high resolution study will allow us to examine. The use of a high-resolution model also allows us to address the known important role of cloud radiative forcing (CRF) in the Arctic and sub-Arctic mountain glaciers. From measurements and modeling (Weller, 1972; Shupe and Intrieri, 2004; Zhang et al., 1996), we know that CRF has a dominant role in the mass balance in the Arctic. The CRF can range from $\sim 40 \text{ W m}^{-2}$ to -75 W m^{-2} (Zhao and Garrett, 2015), which translates to a positive CRF (warming) in the winter months and cooling during the summer of about -50 W m^{-2} . Low cloud feedbacks have been suggested to partially offset the insolation decreases (Jochum et al., 2012), meaning they cause less cooling in a glacial inception scenario. These issues are not well addressed by GCMs because their coarse resolution necessitates the use of inaccurate and uncertain convection and cloud parameterizations. Our model can be run in a cloud-resolving configuration without convective parameterizations, allowing us to study the important interactions between topography, orographic precipitation and cloud formation. Using this model we explore and analyze the impact of topography, clouds, orbital parameters, and meteorology on the melting and accumulation of mountain glaciers.

We view the process of glacial inception as consisting of at least three mechanisms, two of which we explore in this paper. First, ice caps may expand from their current locations in a way that may be very sensitive to climate, but is fundamentally reversible and linear for small-amplitude changes. Second, snow cover may alter regional climate and surface mass balance through the snow-albedo

feedback. This second mechanism may simply act as an amplifier but can also lead to instability and irreversible transitions in idealized models (Lee and North, 1995), although such instability is poorly understood in the context of glacial inception. Our main focus is the first mechanism (reversible ice cap growth), but we also address the second (regional snow-albedo feedback). A third mechanism underlying glacial inception, not dealt with here, is the development of a large-scale ice sheet whose growing elevation leads to further climate feedbacks including both cooling – which reduces melting – as well as the elevation-desert effect which reduces accumulation at large heights (Weertman, 1976; Oerlemans, 1989). This third mechanism, known as the height-mass balance feedback, can also lead to instability and irreversible transitions, but we cannot directly address it because our model simulations are only run for a few years without an ice sheet model for glacial flow.

Our main results show that an accurate representation of topography beyond what can currently be done in state-of-the-art GCMs is critical for correctly simulating the mass balance of mountain glaciers. We demonstrate and analyze the importance of orographic precipitation. The results indicate that clouds, in the presence of artificially lowered topography, respond by inducing cooling and reducing melting over the low topography. We, therefore, suggest that the results of GCMs – even if they find inception under the appropriate orbital forcing – need to be critically examined in view of this compensation between clouds and smoothed topography. Overall, with the correct topography, we find that a series of cold and wet years can lead to glacier expansion on Baffin Island, roughly doubling the areal extent of the ice cap. We find little evidence for irreversible snow-albedo feedback at the spatial scales of our simulation, but we do show that the snow-albedo feedback can amplify orbitally-forced changes in the surface mass balance and lead to further growth of perennial snow.

In the following sections, we describe our model setup (section 2.2.1) and experimental design (section 2.2.2), present and analyze our results (section 2.3) and conclude (section 2.4).

2.2 METHODS

2.2.1 WEATHER RESEARCH AND FORECASTING MODEL (WRF)

We use WRF (Skamarock et al., 2008), version 3.7, a fully compressible, non-hydrostatic model, which has been shown to reproduce reasonable values in the Arctic (Cassano et al., 2011) and over complex terrain (Kilpeläinen et al., 2011). Re-analysis data from the European Centre for Medium-Range Weather Forecasts (ECMWF, Dee et al., 2011) are used for setting the meteorological boundary and initial conditions, including sea surface temperatures and sea ice extent. Cassano et al. (2011) also noted that the Interim Re-Analysis (ERA-Interim) is a suitable choice of boundary conditions for WRF, leading to minimal bias in the Arctic, specifically in temperature. We set up a nested experiment with a 12 km resolution outer domain (Fig. 2.2a). Technically, there is some concern that this region may be sensitive to our domain size choice, but a test simulation with 20 km resolution produces similar results, giving us confidence in our 12 km parent domain. The inner domain has a 4 km resolution (Fig. 2.2b), and the actual elevation of the Penny Ice Cap, exceeding 2000 m, is well-resolved at a 4 km grid spacing. We use realistic land cover from the WRF database; note the extent of Baffin Island covered by ice land-use type is outlined in white in Fig. 2.2. Over these land grid points classified as ice, we prescribe a snow water equivalent (SWE) of 5000 kg m^{-2} — sufficient for maintaining snow cover for the duration of our runs. We focus our analysis on changes over this initial ice cap.

We use the WRF Single Moment 6 Class microphysics scheme, which allows for ice, snow and graupel processes (Hong and Lim, 2006). NOAH-MP (Niu et al., 2011) is used for the land surface model. It includes 4 layers of snow and allows for refreezing, which is an important process on the Penny Ice Cap according to historical trends (Zdanowicz et al., 2012). NOAH-MP caps the snow water content at 2000 kg m^{-2} by default, and we removed this limit. The radiation scheme is the Rapid Radiative Transfer Model for GCM applications (RRTMG, Iacono et al., 2008), which has

been found to have minimal bias in WRF (Cassano et al., 2011). In this scheme, we set the CO_2 to 290 ppm for our glacial inception scenarios (Petit et al., 1999; Vettoretti and Peltier, 2004). Since this is a high-resolution study at 4 km, we chose to allow WRF to be cloud resolving, turning off cumulus parameterization. The boundary layer is simulated with the Mellor-Yamada-Janic scheme (MYJ, Janjic, 1994). We do not nudge to observations in any way other than prescribed lateral boundary conditions, as our goal is to simulate a possible atmosphere 115 kya. With this WRF configuration we find a reasonable reproduction of the ECMWF boundary conditions when we simulate present-day climate. Furthermore, precipitation on the coast matches observations at values of about 300 mm per year, and the equilibrium line altitude is about 1400 m in our model, also agreeing with observations (Zdanowicz et al., 2012).

2.2.2 EXPERIMENTS

To better understand the importance of insolation (J), meteorology (playing a role via both temperature and precipitation rate, and referred to together as T), and topography (Z) on glacial inception, we conduct six experiments (Table 2.1). Plotting the integrated insolation (Huybers, 2006) over the past 140,000 years, which includes the last ice age and previous interglacial, indicates an insolation minimum at 115 kya (Fig. 2.1a). Ice and sediment core records (Petit et al., 1999; Clark et al., 1993) confirm that the last glacial inception began at this time. Therefore, we choose this orbital configuration as our base state. We also run the model for the insolation of 128 kya (time of maximum integrated insolation over the past 140 ka) and present-day (which coincidentally is a mean value for the past 140 ka) insolation, denoted $+\delta J$ and $+\delta J$ respectively (Fig. 2.1a). Our method for setting the chosen orbital configuration in WRF’s radiation module is detailed in the Appendix.

We examine the sensitivity of the Penny Ice Cap region to heat and moisture fluxes by using present-day interannual variability to supply boundary conditions for the regional model. We do not comment on how such altered boundary conditions would arise in a past climate simulation,

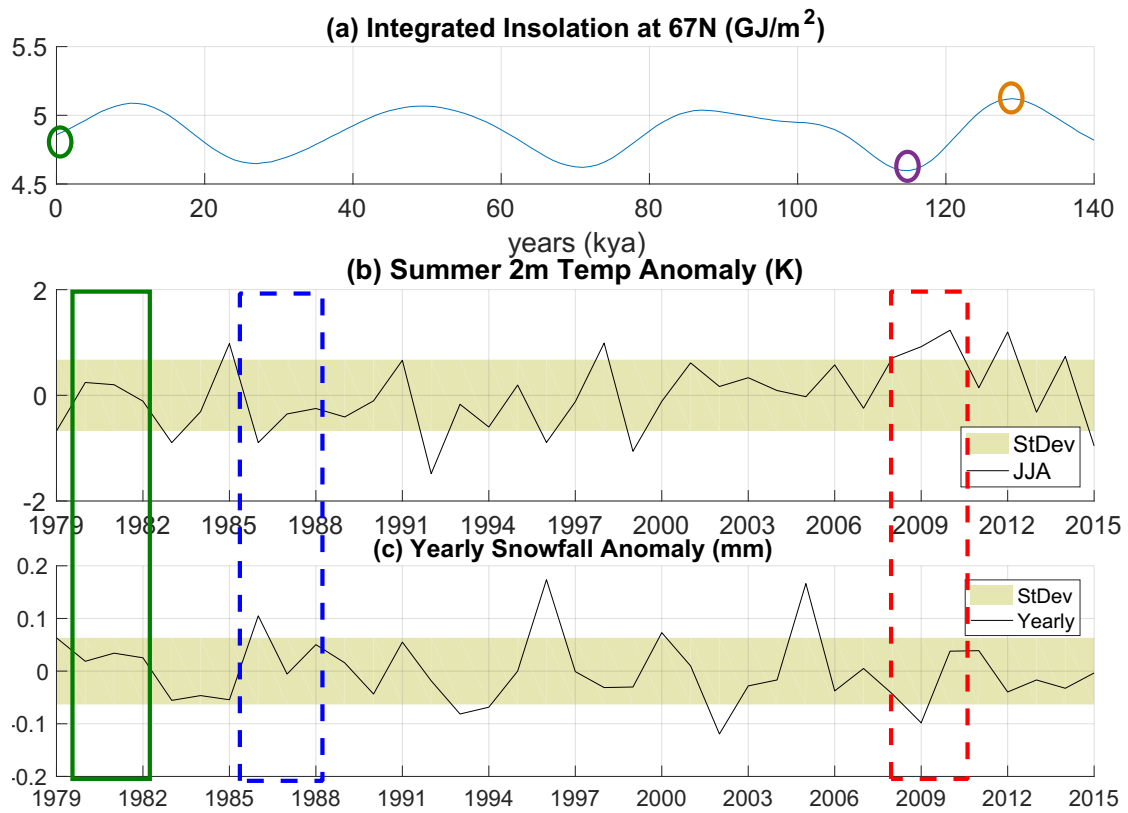


Figure 2.1: (a) Integrated Insolation for 128 kya (maximum insolation, orange), 115 kya (minimum insolation, purple), and 0 kya or present day (mean insolation, green). (b) Summer temperature and (c) Snowfall anomaly and standard deviation for "average" meteorology (green), "cold" meteorology (blue), and "warm" meteorology (red).

Table 2.1: Configuration of WRF simulations with perturbations to integrated insolation (J), topography (Z), and meteorology (T).

Run	T	Z	J	Perturbation Type
115 kya	1980-1982 (Average)	real	115 kya 4.6GJ m^{-2}	Control
Pres	1980-1982 (Average)	real	present 4.85GJ m^{-2}	$+\delta J$
128 kya	1980-1982 (Average)	real	128 kya 5.1GJ m^{-2}	$++\delta J$
Topo	1980-1982 (Average)	low	115 kya 4.6GJ m^{-2}	$-\delta Z$
Cold	1986-1988 (Cold, Wet)	real	115 kya 4.6GJ m^{-2}	$-\delta T$
Warm	2008-2010 (Warm, Dry)	real	115 kya 4.6GJ m^{-2}	$+\delta T$

but instead suggest that the present-day variability can be used as general proxy for regional warming or cooling – which could occur naturally or due orbital or CO_2 forcing. Therefore, using the ERA-Interim reanalysis dataset spanning the past 30 years, we determine average, cold, and warm boundary conditions. We also confirm that these anomalies are not unique to ECMWF, but are also consistent with NCEP reanalysis data (Kalnay et al., 1996). For our control run, we identify 1980-1982 as the three consecutive years that are the least anomalous (within one standard deviation and close to 0 anomaly) regarding summer temperature and yearly snowfall over the Baffin Island region (Fig. 2.1b,c). We run this simulation from October 1979 to October 1982 with 115 kya insolation and realistic topography. Model runs representing cold ($-\delta T$) and warm ($+\delta T$) perturbation simulations use boundary conditions based on 3 year periods of the most anomalous summer temperatures and yearly snowfall (with at least one year outside of one standard deviation). We identify 1986-1988 as anomalously wet with cold summers, and 2008-2010 as anomalously dry with warm summers (Fig. 2.1b,c). These simulations with cold and warm boundary conditions use insolation from 115 kya and realistic topography. We confirm that the atmosphere (heat and moisture fluxes) is strongly constrained by the boundary conditions, and for a regional study, these fluxes are mostly independent of the outer domain size.

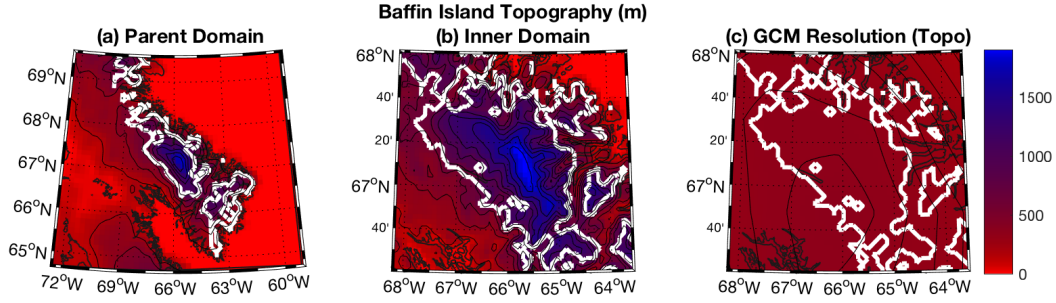


Figure 2.2: (a) Outer parent domain surface elevation (m) with 12 km resolution. (b) Inner nested domain with 4 km resolution. (c) Inner domain with low (GCM-like) topography. Present day realistic ice cap extent is outlined in white.

Table 2.2: Configuration of WRF simulations for ice cap sensitivity.

Run	T	Z	J	Threshold Elev.
Avg. 800 m	1980	real	115 kya 4.6GJ m^{-2}	800 m
Avg. 400 m	1980	real	115 kya 4.6GJ m^{-2}	400 m
Avg. All	1980	real	115 kya 4.6GJ m^{-2}	0 m (All Snow)
Avg. No Snow	1980	real	115 kya 4.6GJ m^{-2}	N/A (No Snow)
Cold 800 m	1986	real	115 kya 4.6GJ m^{-2}	800 m
Cold 400 m	1986	real	115 kya 4.6GJ m^{-2}	400 m
Cold All	1986	real	115 kya 4.6GJ m^{-2}	0 m (All Snow)

In order to test the role of topography in inception, we run a case with modified topography (run $-\delta Z$) using the topography from the 1-degree resolution Community Earth System Model (Hurrell et al., 2013) from CMIP5. Over the Penny Ice Cap, this configuration lowers the peak elevation to ~ 500 m (Fig. 2.2c), whereas the 4 km resolution is over 1900 m, just shy of the realistic 2000 m peak elevation (Fig. 2.2b).

We also test the sensitivity of the region to initial snow cover extent by imposing an initial ice cap above a threshold elevation, which is varied across experiments (Table 2.2). We impose of glacial inception scenario by using 115 kya insolation and run this set of experiments for one year with both

“average” (Oct. 1979-Oct. 1980) and “cold” (Oct. 1985-Oct. 1986) boundary conditions. We vary the initial ice cap extent by changing the land type to snow/ice and setting the SWE to be 5000 kg m^{-2} for all grid points above the chosen threshold elevations of 800 m (approximate Penny Ice Cap extent), 400 m and 0 m (all land covered). Additionally, we perform one simulation with average meteorology and no initialized ice cap, with tundra land type prescribed everywhere.

2.3 RESULTS

In the first subsection below, we study the first step of glacial inception – reversible mountain glacier growth over Baffin Island – by considering the sensitivity of the mass balance to insolation, meteorology, and topography (section 2.3.1). In the following subsection (2.3.2), we examine the second step of the inception process: the potential for regional snow-albedo feedback to amplify initial changes in glacier extent.

2.3.1 SENSITIVITY TO INSOLATION, TEMPERATURE, AND TOPOGRAPHY

Milankovitch theory explains glacial inception mostly through decreased summer insolation and therefore ablation (A), yet total precipitation (P) matters too, of course. In our simulations, there is very little liquid precipitation, and when it does occur, most freezes and contributes to the snow-pack (not shown). Thus, we find that year round all types of precipitation over Baffin Island are important to the mass balance and we define the net snow accumulation for each year as $P - A$. The difference in net accumulation between our control simulation and all other experiments is denoted $\delta(P - A)$, which varies based on our chosen perturbations of insolation (δJ), temperature (δT), and topography (δZ), see Table 2.1.

We begin our analysis of glacial inception by looking at average snow depth on the Penny Ice Cap (approximately 800 m and above) over model runs that are three years long (Fig. 2.3a). There is net

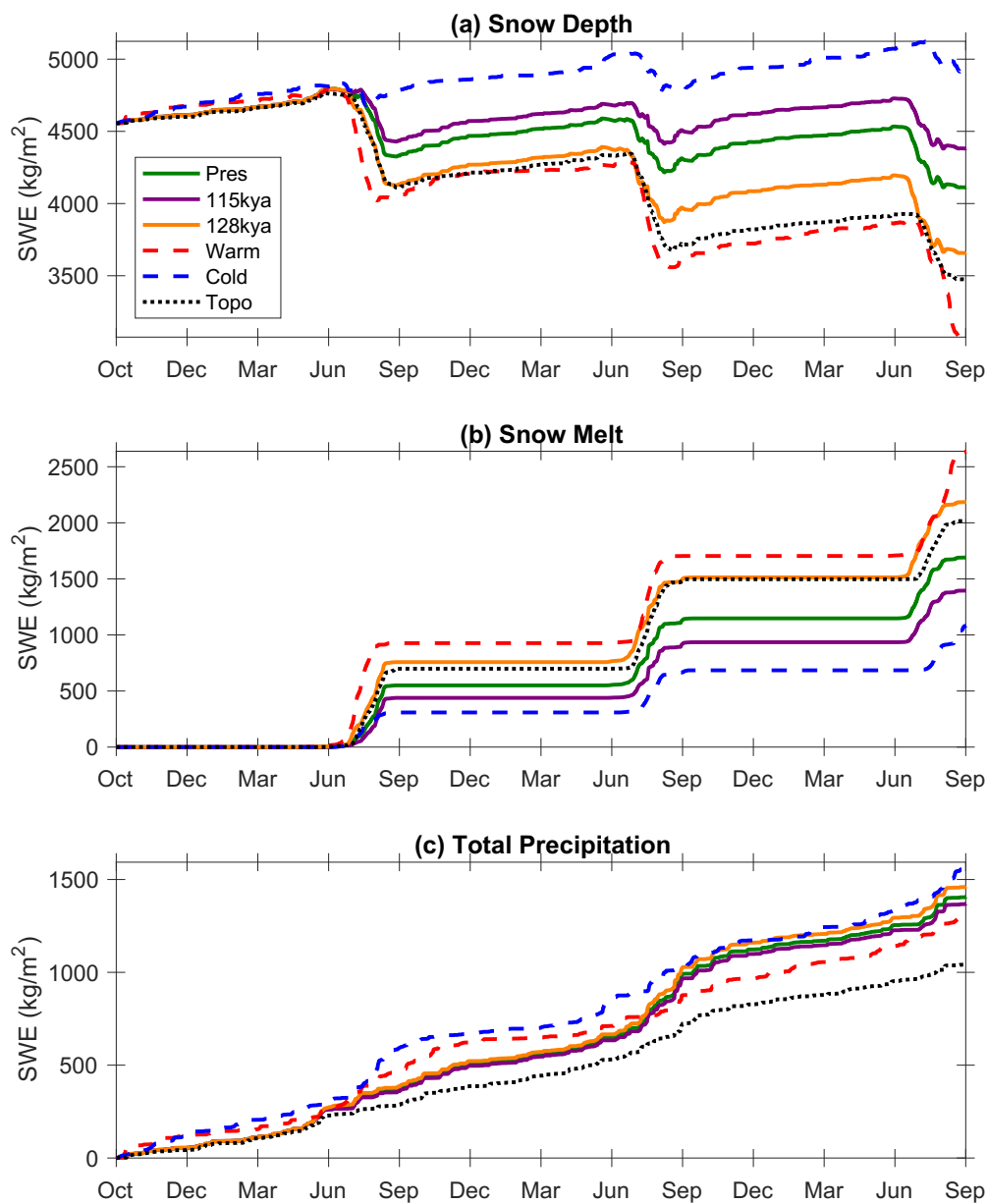


Figure 2.3: Time series over the Penny Ice Cap (above 800 m) for: (a) Snow depth, (b) Melting, and (c) Total precipitation. Simulation details are located in Table 2.1.

melting for all simulations, except for the “cold” ($-\delta T$) simulation (Table 2.1), which is 0.5°C colder and has about 7% more precipitation than the control simulation with average meteorology. In our control case with 115 kya insolation, the snow depth only decreases by 100 kg m^{-2} and appears to be approximately in equilibrium according to the area-integrated mass balance (Fig. 2.5b). Figs. 2.3b, 2.3c confirm that the simulation with cold meteorology has the least melting during the summer and the most precipitation over three years, leading to the only experiment with a positive net accumulation ($P - A > 0$) over the initialized ice cap. Our simulations with 115 kya insolation have shorter melting seasons than those using insolation from present day ($+\delta J$) and 128 kya ($++\delta J$), as expected. The only exception is when the experiment is forced with warm boundary conditions and 115 kya insolation, which leads to the longest and strongest melting seasons causing a loss of 500 kg m^{-2} per year on the ice cap. Thus, we conclude that meteorology, via its effects on temperature and accumulation, has the largest impact on mass balance in these experiments. The difference in summer temperature between the colder ($-\delta T$) and warmer ($+\delta T$) boundary conditions is about 1.5°C , and leads to a melting difference of $500\text{ kg m}^{-2}\text{ year}^{-1}$ (Fig. 2.3a). The yearly response of the melting to the imposed warm and cold boundary condition temperature perturbations is $\sim 300\text{ kg m}^{-2}$ per degree.

In agreement with Milankovitch theory, we find that stronger insolation ($+\delta J$ and $++\delta J$) causes ablation to increase, by $100\text{ kg m}^{-2}\text{ year}^{-1}$ and $200\text{ kg m}^{-2}\text{ year}^{-1}$ respectively, compared to our control simulation with minimum insolation from 115 kya. However, we can also see the effect of local insolation changes on precipitation by looking at Fig. 2.3c. Our simulations with 115 kya (purple), present-day (green), and 128 kya insolation (orange) have slightly different precipitation rates. Thus, Milankovitch forcing causes local changes in precipitation of $\sim 20\text{ kg m}^{-2}\text{ year}^{-1}$ with the 115 kya simulation having the least amount of precipitation and 128 kya having the most. For our glacial inception scenario, we find that 115 kya insolation causes cooler temperatures, less moisture in the atmosphere, and decreased precipitation. In particular, we find that the rainfall rates are re-

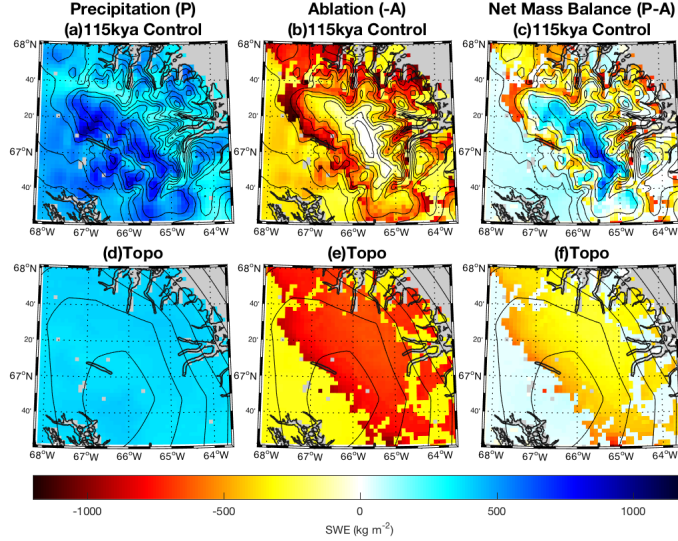


Figure 2.4: Maps of average yearly accumulation (a,d), ablation (b,e), and mass balance (c,f) for realistic (top row) and GCM-like (bottom row) topography.

duced when insolation decreases, but snowfall is fairly consistent across the runs (not shown). This result contrasts studies with coupled atmosphere-ocean models that find increased moisture flux and precipitation in the Arctic (Khodri et al., 2001; Jackson and Broccoli, 2003) during glacial inception and disagrees with Yoshimori et al. (2002), who found increased total precipitation but decreased snowfall. Thus, we are interested in examining how large-scale circulation is impacted by insolation changes, what is necessary for our ideal (cold and wet) conditions, and if the increased moisture flux found in earlier studies can be counteracted by regional temperature decreases.

The importance of realistic topography (Z) is demonstrated in Fig. 2.3a, which shows that using a low, GCM-like, topography ($-\delta Z$) causes snow loss equivalent to when the model is forced with anomalously warm weather (“warm”) or increased integrated insolation (“128 kya”). In only three years, lowered topography alone causes a decrease in snow depth of about 1000 kg m^{-2} , but this decrease in snow depth is not from melting alone (Fig. 2.3b,c). The low-topography simula-

tion experiences $\sim 500 \text{ kg m}^{-2}$ more melting and $\sim 500 \text{ kg m}^{-2}$ less precipitation than the control experiment, indicating that precipitation and melting play an equal role in the mass balance. Since the control experiment and the low topography experiment are forced with the same atmospheric boundary conditions, their differences in accumulation indicate that orographic precipitation may be important in this region. We confirm that convective precipitation is mostly inactive in this region, and we diagnose possible orographic enhancement of precipitation from the spatial pattern of accumulation in Fig. 2.4a. The highest rates ($1000 \text{ kg m}^{-2} \text{ year}^{-1}$) occur on the windward (southwestern) side of the Penny ice and lower ($200 \text{ kg m}^{-2} \text{ year}^{-1}$) on the lee (northeastern) side. In contrast, when the simulation uses smoothed GCM-like topography, the snowfall is mostly uniform over the domain due to the absence of significant topography (Fig. 2.4d). The low-topography simulation also has no snowfall during the summer and reduced snowfall during most of the rest of the year (not shown). Thus, realistic orography enhances the amount and alters the type of precipitation, which is not captured by GCMs, whose topography is substantially lower due to their coarse resolution.

The effect of topography on the mass balance is further explored using spatial averages (Fig. 2.4). Precipitation is enhanced by orography, but the topography also greatly affects ablation. In the lower topography case (Fig. 2.4e), ablation is fairly uniform, while in the realistic topography case (Fig. 2.4b) the highest amount of melting is localized on the edges of the initialized snow cover at about 800 m, and there is no melting at the top of the Penny Ice Cap. The pattern of mass balance for the realistic case (Fig. 2.4c) suggests that the upper ice cap would grow each year with realistic topography, while mass balance for the low-topography simulation indicates that the ice cap would be completely eliminated (Fig. 2.4f).

We investigate ice cap growth further by examining elevation-sorted mass balance in Fig. 2.5. We rank each land grid cell by the surface elevation and aggregate the points into bins representing 5% of the area each (Fig. 2.5a). Note that the ice cap lies above 800 m, which is in the 65th height per-

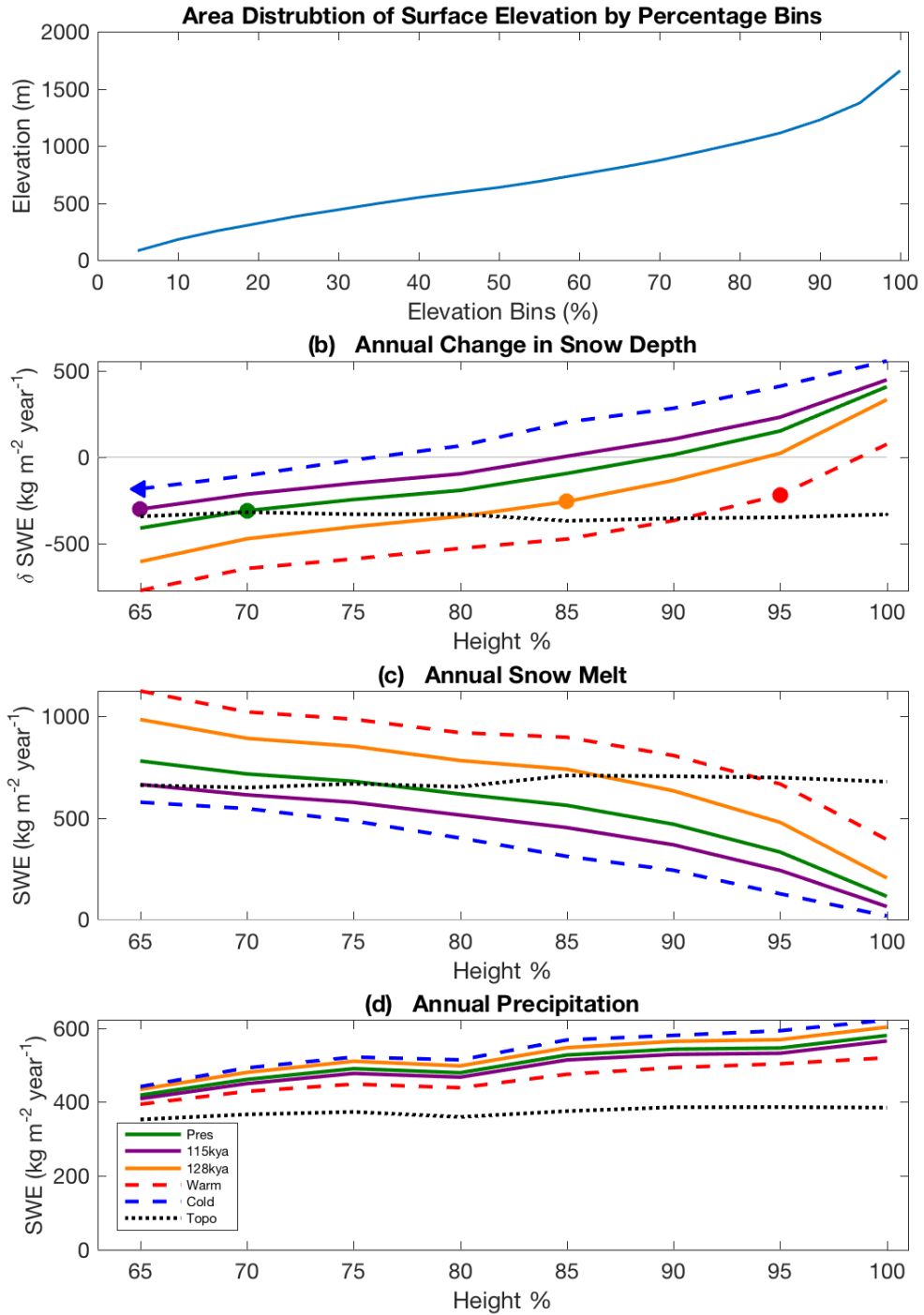


Figure 2.5: (a) Surface elevation area distribution, by percentile. (b) Annual change in snow depth, binned by surface height percentile. We then integrate the mass balance from the highest grid cells downward. A circle signifies our inferred equilibrium ice cap extent (integrated mass balance=0) and an arrow indicates the ice cap growth (positive integrated mass balance). (c) Annual snow melt. (d) Annual precipitation.

centile. In Fig. 2.5b, plotting the annual change in snow depth identifies the accumulation zone (values above zero), the ablation zone (values below zero), and the equilibrium line (the transition from net accumulation to net melting). In Fig. 2.5b, only the low-topography simulation (black dotted line) shows net melting over the entire domain. Increasing the insolation causes the equilibrium line to shift to higher altitudes, as seen by comparing the control 115 kya case (purple) with the increased insolation 128 kya case (orange). Similarly, the combination of cold meteorology and 115 kya insolation results in the lowest equilibrium line.

In an actual ice cap, accumulation at the higher elevations leads to ice flow to lower elevations where the ice is ablated, such that over the extent of the ice cap, total accumulation equals total ablation. Our model does not include ice flow explicitly and we, therefore, represent its effects indirectly. For that purpose, we integrate the yearly mass balance ($P - A$) on the ice cap (65th %ile of height and above) beginning at the upper part of the domain where $(P - A) > 0$, and integrating downslope until the integrated mass balance equals zero, indicated by large colored solid dots in Fig. 2.5b. This location, where the integrated mass balance vanishes, indicates the anticipated location of the equilibrium ice cap edge. We perform this and future integrations only over ice-covered area, where the mass balance can actually be evaluated. This, of course, should be viewed as a crude representation of ice flow and the feedbacks that it would induce. The present-day experiment shows that the ice cap extent is at the 70th %ile, corresponding to a height of ~ 900 m. This suggests that if run with an ice flow model, the ice cap in this run would equilibrate just above the current height of the Penny ice cap at 800 m (65th %ile). The 128 kyr and warm meteorology experiments would retreat further upslope to heights of 1100 m (85th %ile) and 1300 m (95th %ile), respectively. The low-topography experiment has no accumulation zone; the entire ice cap is therefore expected to disappear. The integrated mass balance of the control (solid purple line in Fig. 2.5b) goes to zero at the edge of the ice cap, signifying that the ice cap edge would be maintained at 800 m with 115 kya insolation, realistic topography, and average meteorology. Finally, for the cold simulation (dashed

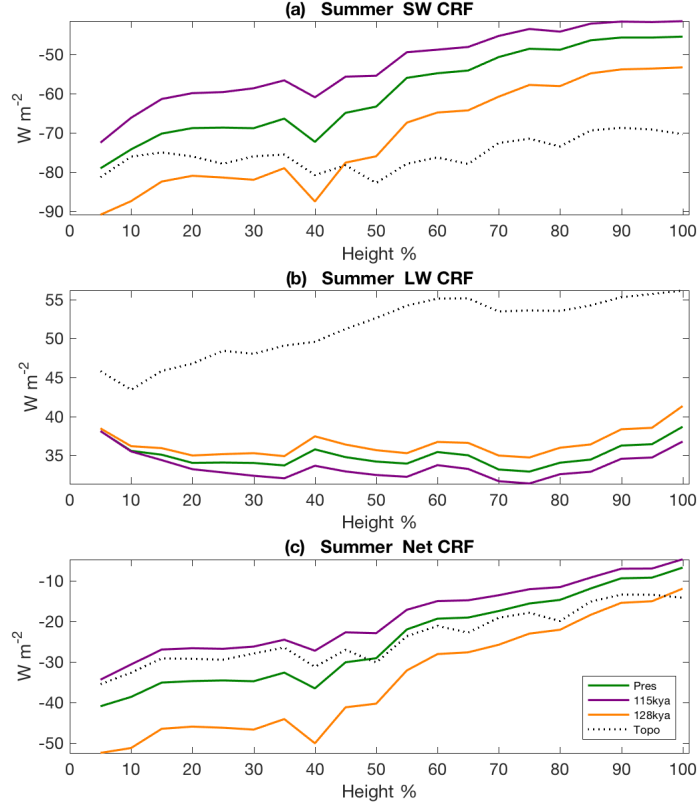


Figure 2.6: Surface CRF binned by surface height percentile from summer (JJA) of the first year of the simulation: (a) Shortwave cloud radiative forcing, (b) Longwave cloud radiative forcing, and (c) Net cloud radiative forcing.

blue line in Fig. 2.5b), the ice cap maintains a positive integrated mass balance as indicated by the blue arrow and is expected to grow — and perhaps develop into an ice sheet. The average grid cell of the initial ice cap in this simulation accumulates 0.2 m per year; over time, ice flow would allow the ice cap to expand downslope past its initial bounds. In the next subsection, we analyze the ice cap growth by considering simulations with different initial ice cap extents.

Our cloud resolving and high-resolution model allows us to examine the role of clouds in glacial inception and provide an independent test of GCM studies based on parameterized clouds and

convection. Jochum et al. (2012) found low clouds to be a negative feedback when it comes to Milankovitch forcing: weaker insolation leads to less low clouds and therefore to a weaker shortwave cloud radiative forcing (CRF). Clouds also lead to warming that may counteract the Milankovitch-induced cooling. Fig. 2.6 shows the shortwave (SW), longwave (LW) and net (SW+LW) CRF sorted by elevation for the summer (June, July, August) during the first year of the model run. For present day insolation (green line in Fig. 2.6c), the model CRF is -40 W m^{-2} at low elevations, comparing well with other studies on CRF in the Arctic (Zhao and Garrett, 2015). The figure shows only four of the runs because interannual variation in CRF for the warm and cold simulations is too large to allow us to deduce any clear temperature-dependence. Shortwave cloud forcing becomes less negative as insolation decreases (compare orange and purple curves in Fig. 2.6a), and this effect is consistent over the three years we ran the model, which is why we only look at the first year. Next, we consider the interplay of clouds and topography. The low-topography simulation forms more low clouds over the entire domain, and these clouds have a higher liquid water path and more negative CRF. The net CRF for low topography only becomes similar to that of realistic topography in the lowest part of the domain, where the two simulations have similar cloud properties. We conclude that clouds act as a negative feedback not only in response to Milankovitch changes but also in response to topography variations. This suggests that GCM studies of the inception problem that do not represent the topography accurately may suffer CRF biases, which may allow the development of ice caps due to cloud response to the modified topography.

2.3.2 SENSITIVITY OF MASS BALANCE TO INITIAL SNOW EXTENT

The previous set of experiments involves an essentially linear and reversible sensitivity to various inception forcing factors, and from our above analysis, we infer the location of the equilibrium ice cap edge by integrating the mass balance to the point where its net vanishes. As discussed in the introduction, the second stage of the inception process involves the (potentially nonlinear and irre-

versible) snow-albedo feedback and the effect larger areas of snow cover on the local climate, particularly the survival of perennial snow. This feedback is our focus here.

In order to examine this second inception mechanism, our next set of experiments explores the sensitivity of the mass balance to a specified snow cover that is beyond the present-day ice cap extent. We perform these runs for both “average” (1980) and “cold” (1986) meteorology boundary conditions — please note there are differences in mass balance between Fig. 2.5, which is the average of three years, and Fig. 2.7, which is only a year long simulation. As explained in section 2.32.2.2, we run such experiments for initial snow cover that is specified above threshold elevations of 800 m, 400 m, and 0 m.

For each of the two boundary conditions, we find that the equilibrium line (transition from accumulation to ablation) occurs at nearly the same elevation, although increased snow cover does cause a slight down-slope shift (compare same color curves in Fig. 2.7a). Note that the mass balance ($P - A$) may be estimated only over height ranges that are initially snow-covered, and therefore, we only plot it over these elevations. The “average” boundary conditions have an equilibrium line at about 1000 m (80th %ile elevation bin) and the cold boundary conditions have an equilibrium line at about 800 m (70th %ile). We again calculate the integrated mass balance over the ice cap in each experiment. We denote the location of the zero integrated mass balance with a solid circle, and if the integrated mass balance remains positive down to the edge of initial ice extent, indicating growth, we use an arrow. For the 800 m threshold experiment denoted by the dotted blue curve in Fig. 2.7a the integrated mass balance does not vanish over the specified initial snow extent, indicating that the 800 m ice cap would grow. This is different from the three year simulations in Figure 2.5b, because we only ran the model for one year, which has the least amount of melting.

With average meteorology, the ice cap is expected to expand to 700 m (60th %ile, purple circle) as identified by the 400 m and all-snow (0 m) simulations. This conclusion can only be reached by specifying larger snow cover than the present-day ice cap. Furthermore, the “cold” boundary con-

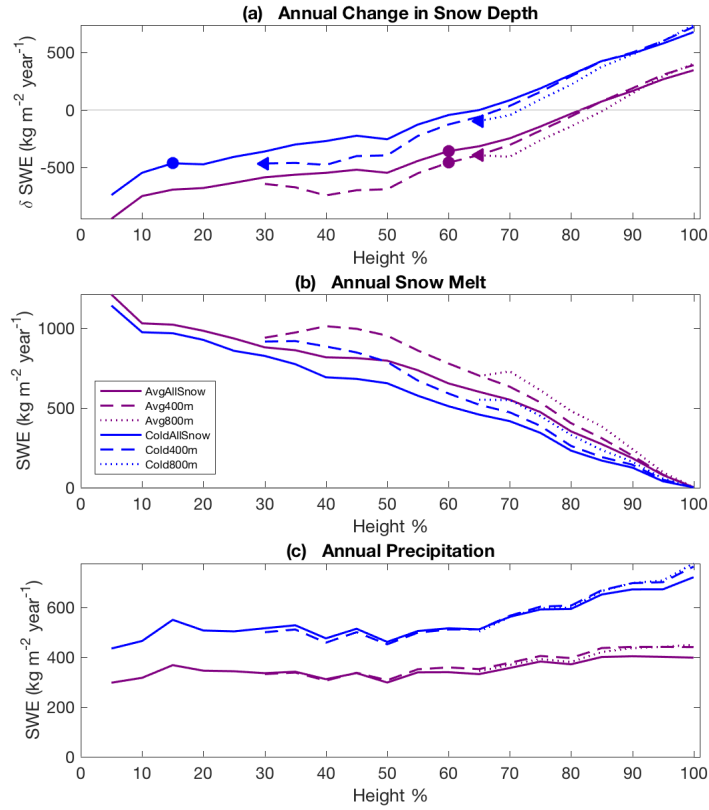


Figure 2.7: (a) Annual change in snow depth binned by surface height percentile with “average” and “cold” meteorology and varying initial ice cover. We then integrate the mass balance from the highest grid cells downward. A circle signifies our inferred equilibrium ice cap extent (integrated mass balance=0) and an arrow indicates the ice cap growth (positive integrated mass balance). (b) Annual snow melt. (c) Annual precipitation. Simulation details located in Table 2.2.

ditions result in a doubling of ice cap areal extent (relative to average meteorology), with the ice cap edge at about 300 m (15th %ile, blue circle). Such meteorology (cold summers, larger precipitation) persisting over several years can, therefore, lead to ice growth over much of the area surrounding the Penny Ice Cap.

We also examine ablation (A) and precipitation (P) separately in Figs. 2.7b,c. As expected, melting decreases at higher elevations where temperature is colder, but we see a slight fanning out of the yearly melt rates downslope for different ice cap extents. For instance, when we lower the threshold elevation from 400 m to 0 m (all snow), less melting occurs (compare solid and dashed curves in panel b), and we will show next that this is because the temperature is lower in the all-snow experiment, due to the snow-albedo feedback. The yearly precipitation does not differ greatly between the various ice cap extents, except above 1200 m (85th %ile), where the all-snow simulations have $50 \text{ kg m}^{-2} \text{ year}^{-1}$ less snowfall, colder temperatures, and less moisture in the atmosphere. Fig. 2.7c indicates that over the ice cap the “cold” boundary conditions cause about $300 \text{ kg m}^{-2} \text{ year}^{-1}$ of precipitation more than the average boundary conditions. Note that this precipitation rate is different from the simulations run for three years (1986-1988) because 1986 is the coldest and wettest of these three years. Thus, we find that the downward expansion of the ice cap edge from 800 m to 300 m is largely dependent on precipitation resulting from the large-scale circulation prescribed at the boundaries.

To confirm that the differences in melting between the different initial snow-cover experiments are due to the snow-albedo feedback, we examine the summer temperature over the domain for average boundary conditions (Fig. 2.8). Temperatures are below freezing where ice is present above 800 m in the control case (Fig. 2.8a). When we change the initial snow cover threshold elevation to 400 m, the 2 m temperature below 800 m decreases since snow is now present, and below 400 m, the temperature is only $\sim 1^\circ$ colder. When snow covers the entire domain, the snow albedo feedback causes colder temperatures than the 400 m simulation over the same elevations (note the darker blue

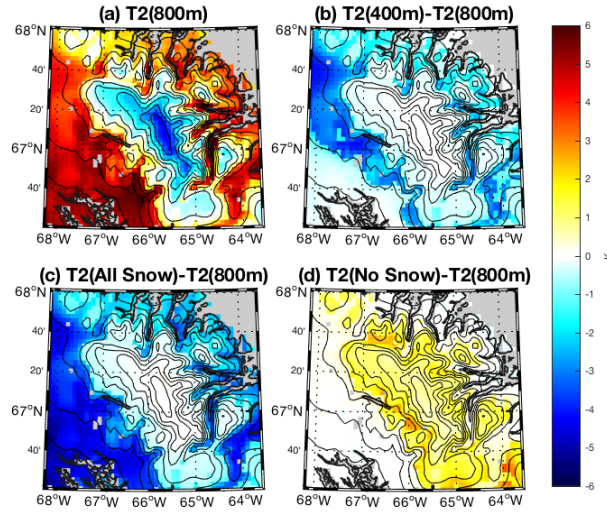


Figure 2.8: A map of the summer (JJA) mean 2m air temperature ($^{\circ}\text{C}$) for (a) 115 kya control with 800 m ice cap extent and average meteorology. The 2m air temperature differences: (b) 800 m ice cap extent minus the 400 m simulation (c) 800 m minus the all-snow simulation (d) 800 m minus the no snow simulation.

pattern). The temperature above 800 m does not change as we lower the ice cap threshold elevation from 800 m to 0 m, but when the simulation is initiated without snow cover, temperatures are up to 3 degrees warmer during the summer even though an ice cap is developing.

Thus, we identify the snow-albedo feedback to be significant in this region from decreased summer temperature and lessened melting with increased snow cover. For another perspective on the effect of the snow-albedo feedback, we consider a scatter plot of positive degree days (PDD) against melting for the simulations with average meteorology (Fig. 2.9). We use these plots to determine how changes in net accumulation $P - A$ relate to changes in temperature, and how topography impacts melt rates. The data points are color coded by grid-cell elevation, where blue is the highest (and coldest) part of the domain and red is the lowest. Melting increases where there are more positive degree days, and there are generally more positive degree days at lower elevations. For an initial snow cover threshold of 800 m, we observe a rather linear relationship between melting and PDD (Fig. 2.9a). As the threshold elevation is decreased, the snow-albedo feedback decreases melting rates

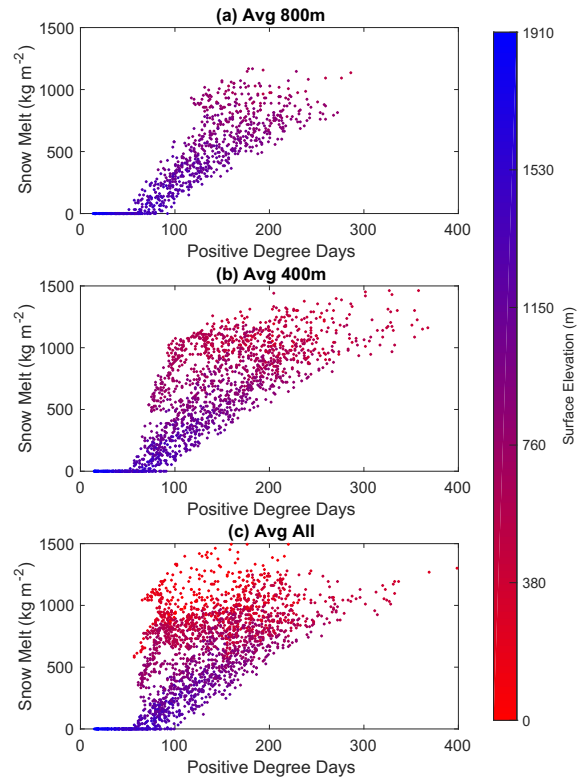


Figure 2.9: Positive degree days vs. melting for "average" meteorology, only including grid points that are initially ice-covered. The color bar represents surface elevation. (a) 800 m threshold elevation (b) 400 m threshold elevation (c) 0 m threshold elevation (all-snow simulation)

at a given elevation. For instance, the edge of the 800 m ice cap experiences $\sim 1000 \text{ kg m}^{-2} \text{ year}^{-1}$ of melt, but when the entire domain is covered with snow, only elevations below 300 m melt at that rate. We can observe this melt rate decrease in the downward shift of blue and purple points as we move from Fig. 2.9a to Fig. 2.9c (800 m elevation threshold to 0 m).

These same scatter plots show two additional interesting points. First, in all simulations, the highest elevations show positive degree days, but no melting. This occurs due to refreezing being allowed within the land model: snow melts during the day when the temperature is above zero but does not always have time to run off before nightfall when it refreezes. Second, note the spread of PDD values for a given melt rate: e.g., for $1000 \text{ kg m}^{-2} \text{ year}^{-1}$ melt rate, there is a PDD range of 100-300 (Fig. 2.9c). 100 PDD corresponds to lower elevations (more red, $< 300 \text{ m}$) at the ice cap edge, and higher values of 300 PDD have points corresponding to higher elevations (more purple, $\sim 500 \text{ m}$) closer to the center. Thus, use of PDD alone to represent melting does not capture terms in the surface energy budget that depend on surface elevation and tend to lead to less melting higher on the ice cap.

Finally, the cloud radiative forcing is shown in Fig. 2.10c and reveals that the initial snow cover extent substantially impacts the cloud radiative forcing, mostly through temperature and surface albedo changes. Changing the meteorology causes the magnitude of the net cloud forcing to change while leaving the functional dependence on elevation similar (compare blue and purple solid lines, for example, Fig. 2.10c). Where initial snow cover is prescribed, the SW CRF is less negative than over the tundra regions without an initial snow cover (Fig. 2.10a), which is consistent with observations showing the CRF becoming more negative when winter snow melts away (Shupe and Intrieri, 2004). The cloud properties are very similar for all simulations, and the SW CRF changes are as expected: increasing surface albedo reduces the reflection of shortwave radiation by clouds.

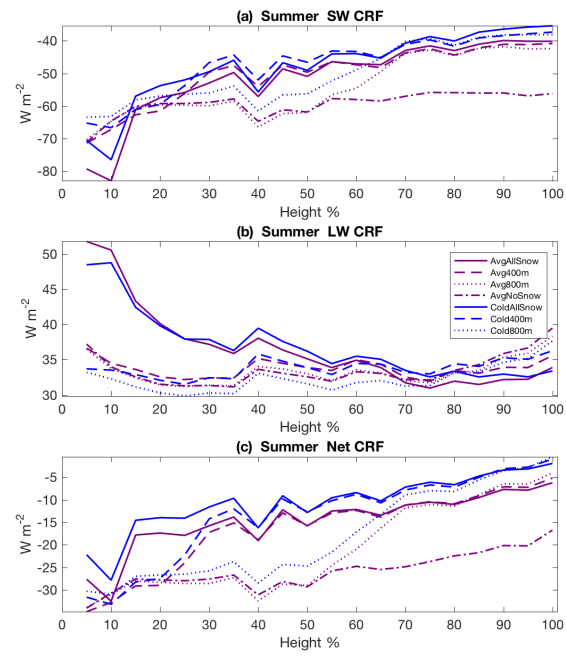


Figure 2.10: Surface CRF binned by surface height percentile during the summer (JJA) for the ice cap sensitivity simulations: (a) Shortwave cloud radiative forcing, (b) Longwave cloud radiative forcing, and (c) Net cloud radiative forcing.

2.4 CONCLUSIONS

We have studied the glacial inception problem using a high-resolution, cloud resolving atmospheric model, focusing on the importance of topography and clouds. Given present-day insolation and average meteorology, our model predicts the Penny ice cap is close to equilibrium (would retreat from 800 m to 900 m elevation with an equilibrium line altitude of ~ 1400 m), indicating that our model is appropriate for the study of the possible ice cap expansion in different climates. We find that a combination of Milankovitch forcing, realistic topography, and moderately cold and wet meteorology allows mountain glacier growth. This is in contrast to some previous studies (e.g., Otieno and Bromwich, 2009), suggesting that additional large cooling of 4°C or some additional feedbacks — for example vegetation — are needed. However, it is possible that such feedbacks would be needed in later stages of inception when snow cover spreads outside of Baffin Island. We find that our choice of boundary conditions heavily constrains the heat and moisture fluxes to the region, and taking advantage of interannual variability allows us to use modern climate as a proxy for the slightly cooler climate experienced during glacial inception.

Because our model does not include active ice flow, we attempt to represent that effect by calculating the glacier mass balance integrated from the top downward. While the model does not show a build up of perennial snow below the current level of the Penny ice cap, the accumulation over the ice cap increases under the appropriate inception conditions mentioned above, implying that simulations with 115 kya insolation and “cold” boundary conditions would allow the ice cap to expand downslope. Our “cold” boundary conditions are represented by the meteorology of years 1986-1988 and have a cooling of only 0.5° and precipitation increase of $\sim 7\%$, leading to the only simulation with a positive mass balance over the ice cap of 0.2 m per year. We find this sufficient for glacier area growth of $\sim 50\text{-}100\%$ — even without any height-mass balance feedback. This is a greater sensitivity of mass balance to climate than found in previous studies that required a much stronger cooling to

get substantial glacier area expansion.

The topography in state-of-the-art GCMs is characterized by ~ 500 m maximal elevations in Baffin island as opposed to the actual 2000 m topography which is well represented in our study. We find that GCM-like representation of topography is insufficient and leads to biases, despite allowing for some progress on the glacial inception problem (Jochum et al., 2012). In particular, the lower topography leads to increased ablation and decreased precipitation over the area crucial to glacial inception. When we run our model with the GCM-like topography, it does not produce any net accumulation zone for average present-day meteorology. The lowered GCM-like topography leads not only to warmer conditions and thus more summer melt, but also to reduced accumulation, as orographic precipitation – a critical process in this region – is eliminated. Smoothed topography also causes rainfall rather than snow during the summer, further contributing to inability of GCMs to produce inception.

Moving beyond the initial glacier expansion that marks the first step of inception, we expect increases in snow and ice cover to lead to a regional snow-albedo feedback which amplifies changes in mass balance. We explored this feedback by testing the sensitivity of the mass balance to changes in initial snow cover, setting the land cover to snow/ice everywhere above an elevation of 800 m, 400 m or 0 m, and also considering both “average” and “cold” boundary conditions. We find that the initial snow cover impacts temperature and melting, but changes precipitation very little. The mass balance deduced from these runs using “cold” meteorology indicates that once covered by snow, the area would support an equilibrium glacier extent above 300 m elevation, although an actual ice flow model would be needed to verify this conclusion and calculate the relevant response time scales of the ice caps.

Previous GCM work by Jochum et al. (2012) found that shortwave cloud radiative forcing acts as a negative feedback on Milankovitch forcing: when the insolation is reduced, low clouds let more SW radiation penetrate and weaken the tendency toward inception. We confirmed this conclusion

using a model that does not rely on parameterized convection and clouds. We also find that clouds act as a negative feedback to topography changes: when the topography is lowered from its realistic values to a GCM-like smooth topography, clouds respond by letting less SW penetrate to the surface. Thus, although lowering the topography leads to surface warming that tends to prevent inception from happening, the cloud forcing makes the state more favorable to inception. This cloud response may allow inception to occur without realistic topography, representing a model bias and suggesting that inception in GCMs with smooth topography should be examined carefully.

We conclude that with realistic topography and 115 kya insolation values, glacier growth (the first step of inception) may be possible. Furthermore, the snow-albedo feedback can amplify the effects of a mere half-degree of cooling (relative to present-day average), and cause substantial snow growth downslope to ~ 300 m (the second phase of glacial inception). Thus, inception may occur during an unusually long sequence of moderately cold and wet years, and inception commences with downslope ice flow rather than snow accumulating spontaneously in unglaciated regions. Given that this required cooling is not extreme and the precipitation anomaly is only 1 standard deviation above average, it is plausible that such a sequence could happen due to normal decadal and millennial climate variability in the interglacial period preceding the inception. A succession of many cooler years could also be a result of forcing by other mechanisms such as changes in CO_2 , ocean circulation, or Milankovitch forcing. It is also possible for precipitation to increase due to increased storm activity (Jackson and Broccoli, 2003), and we are interested in exploring the impact of insolation and ice sheets on large scale circulation. To test these conclusions, we employ an ice flow model in the next chapter, in addition to the realistic topography used here. The inclusion of an ice sheet model would also allow for the exploration of the height-mass balance feedback, the third step of glacial inception.

3

The Interaction of Ice Flow and Meteorology

3.1 INTRODUCTION

For the past million years, the climate has transitioned into an ice age with a periodicity of $\sim 100,000$ years. Ice sheets such as the Laurentide, eventually spanned continents, but they began in the moun-

tains of the Arctic in Eastern Canada (Ives et al., 1975; Clark et al., 1993). Orbital forcing provides a possible explanation: colder summers due to weaker insolation allow for snow survival (Milankovitch, 1941). Favorable meteorology, however, is another necessary component for explaining initial growth of continental ice sheets, which is called inception (Bromwich et al., 2002; Jackson and Broccoli, 2003; Birch et al., 2017). In this scenario, ice accumulates on mountain tops and flows downslope, causing an increase in ice area and elevation changes. The regional cooling occurring due to elevation changes and the snow-albedo feedback would allow for further net accumulation and expansion of the ice sheet. Additional feedbacks from ice elevation changes can impact the atmospheric circulation and therefore, net accumulation (Roe and Lindzen, 2001a).

Glacial inception of the Laurentide ice sheet is thought to have progressed fairly rapidly in eastern Canada (Clark et al., 1993). The sea level drop was estimated to be as high as 50 m sea level in 10 kyr (Waelbroeck et al., 2002; Ganopolski et al., 2010). The Laurentide ice sheet has been identified as the primary initial cause of the drop in sea level (Kleman et al., 2013) and it has been simulated to grow significantly within a few thousand years (Andrews and Mahaffy, 1976; Marshall and Clarke, 1999). However, this rapid growth required the precipitation to be tripled or the climate cooled by 2 degrees every thousand years, without an explicit explanation. A more modest sea level drop 10-20 m in 10 kyr has also been posited (Lambeck and Chappell, 2001; Creveling et al., 2017), which indicates a slower ice sheet growth rate and possibly ice growth relegated to Arctic Canadian Islands (Kleman et al., 2002).

Global climate models (GCMs) are used to simulate glacial inception, but often fail because snow fails to accumulate in unglaciated areas, nor do glaciers spread into unglaciated areas. One reason GCMs have difficulty simulating the glacial inception is their coarse horizontal resolution, which lowers the topographic features dramatically (Wang and Mysak, 2002; Vettoretti and Peltier, 2003). Snow then does not accumulate because the region is too warm due to artificially low elevations (Rind et al., 1989). Though topography is important other aspects of the climate system may also

be missing, like ocean feedbacks (Khodri et al., 2001), dust (Calov et al., 2005; Farrell and Abbot, 2012), sea ice (Gildor and Tziperman, 2001; Li et al., 2005), and vegetation (Wang et al., 2005; Goñi et al., 2005). However, increasing the model resolution has lead to moderate improvements in the simulation of glacial inception (Jochum et al., 2012; Vavrus et al., 2011), through the occurrence of areas of net snow accumulation.

Cool summer temperatures are critical for inception, and while Milankovitch forcing explains some summer cooling, the large scale circulation is an important driver of heat and moisture fluxes to inception areas. In the current climatology, large scale cyclonic circulation over Baffin Bay is associated with cool summers (Bromwich et al., 2002; Gardner and Sharp, 2007), which may be all that is necessary for inception. However, Jackson and Broccoli (2003) and Khodri et al. (2001) also noted increased storms and poleward moisture fluxes during inception. Otieno and Bromwich (2009) showed that in addition to favorable circulation and temperature, snow could accumulate in new areas only after artificial large-scale cooling was prescribed. The question remaining: how does this cooling occur? One possible answer is the interaction of ice sheets and the atmosphere.

As the ice sheet grows, it influences the atmosphere due to albedo and surface elevation changes. The large scale circulation can also be diverted by the changing topography. Anti-cyclones have been noted to occur over mountains (Smith, 1979), but ultimately, the effects of mountains on the regional climate and large-scale circulation depends on the mean overlying flow and mountain size (Chen and Lin, 2005; Cook and Held, 1988). Looking at ice growth in the context of glacial inception, Roe and Lindzen (2001a) argues using a simple model that slight changes in topography can influence the stationary waves, significantly changing the circulation to bring colder temperatures. The configuration of the topographic changes and geographic location may cause stronger impact on the atmosphere than others (Jackson, 2000).

To capture the interaction of ice and the atmosphere, models of intermediate complexity have been combined with ice sheet models. This type of model allows thousands of years of ice growth

due to changes in insolation and CO₂ to be simulated. Often, however, these models glacialize Alaska and Western Canada very rapidly (Kageyama et al., 2004; Bonelli et al., 2009; Ganopolski et al., 2010), which is not consistent with the geological record showing inception occurred first in eastern Canada (Clark et al., 1993). Ice sheet models and GCM studies (Dong and Valdes, 1995; Meissner et al., 2003) have a similar problem of not agreeing with the geological record. Combining GCMs and ice sheet modeling may allow for progress in understanding ice-atmosphere interactions (Herrington and Poulsen, 2011; Gregory et al., 2012; Löfverström et al., 2014). Coarse resolution GCMs, even if coupled with an ice sheet model, still suffer from the problem of biased topography.

The interaction of the ice and the atmosphere is critical to inception. Birch et al. (2017) used a high-resolution cloud-permitting regional climate model and found that ice would accumulate on the mountain glaciers of Baffin Island. The 4 km resolution of that study allowed us to capture both the ablation and accumulation zones of mountain glaciers. By integrating the mass balance from the top of the ice cap down slope, Birch et al. (2017) implicitly represented the effects of ice flow. As a result, the Penny Ice Cap was deduced to grow outside of its present day bounds due to decreased insolation and a sequence of cold summers and wet winters. As was found by previous GCM studies, no snow accumulated in unglaciated areas. Thus, the purpose of the current paper is to explicitly include ice flow from the growing mountain glaciers, allowing us to investigate the coupled interaction of ice flow and atmospheric feedbacks. For this purpose, we use a simple ice model based on the shallow ice approximation, asynchronously coupled with the Weather Research and Forecasting model (WRF, Skamarock et al., 2008). We find that the change to ice sheet topography induces an anticyclonic atmospheric response over Baffin Island. This anticyclone causes a flow of warm air over Baffin Island and Hudson Bay area, increasing ablation, stagnating further ice sheet growth, and thus indicating a negative feedback on inception at the regional scale. This is reminiscent to the findings of Herrington and Poulsen (2011) and Gregory et al. (2012) who noted the continued growth of the ice sheet is slowed by changes in large scale circulation that arise from atmosphere-

topography interactions. While the studies of Herrington and Poulsen (2011) and Gregory et al. (2012) found the negative feedback slowed ice growth mostly when the ice sheet was fairly developed, of a continental-scale, the negative feedback analyzed here occurs upon inception, as soon as the ice caps over the Baffin Island begins to grow. Previous studies (Herrington and Poulsen, 2011; Gregory et al., 2012) could not resolve this feedback their coarse resolution GCMs, while our coupled ice sheet - regional atmospheric model enables us to examine it.

The atmospheric and ice models used for the asynchronous coupling experiment are presented in Section 3.2. We present the results in Section 3.3, beginning with ice thickness and mass balances of our atmosphere-ice iterations. We then investigate the regional and large-scale impacts of expanding ice cover and increasing ice elevation. Sensitivity simulations exploring the importance of topography and insolation are also presented. Finally, the sensitivity of ice flow to ice viscosity and mass balance are examined. We conclude with a discussion in Section 3.4, including the limitations of this study and its implications.

3.2 METHODS

Our goal in this study is to better understand the interaction between changing ice cover and the regional climate of Baffin Island and the large-scale circulation over North America. We therefore use a nested configuration of the Weather Research and Forecasting Model (WRF Skamarock et al., 2008) with the outer domain covering all of North America with a 100 km resolution and the inner domain covering Baffin Island with a 20 km resolution, encompassing both the Penny and Barnes Ice Caps. To better understand the growth of these mountain glaciers during inception, we want to allow ice flow in the inner 20km resolution domain, but WRF does not contain an ice flow model, which means the ice expansion must be calculated offline. Based off of Oerlemans (1981), we developed a simple ice flow model that reduces the 3D Stokes flow to two dimensions using the shallow

ice approximation. The basal sliding velocity is assumed to be 0. This approximation results in a nonlinear diffusion equation in Equation 3.2 that solves for ice thickness and represents ice flow.

$$\frac{\partial H}{\partial t} = \nabla \cdot D \nabla H^* + G$$

H is the ice thickness and we solve for its evolution over time. G in Equation 3.2 is the yearly mass balance that is calculated from WRF simulations and assumes that the density of ice 900 kg m⁻³. $H^* = B + H$ is the height of the ice surface or the surface elevation, where B is the basal topography. D represents the diffusivity with

$$D = AH^{m+2} \left[\left(\frac{\partial H^*}{\partial x} \right)^2 + \left(\frac{\partial H^*}{\partial y} \right)^2 \right]^{(m-1)/2}$$

m for this model simulation is set to 3, which is a typical value for realistic ice simulations Cuffey and Paterson (2010). A in Equation 3.2 is the creep factor that influences the viscosity of the ice is based on the temperature of the ice, i.e. the colder ice flows more slowly, while warmer ice more quickly. We use the creep parameter A from Cuffey and Paterson (2010) evaluated at a temperature of -10°C. More details on this choice of creep parameter can be found in Appendix B. Also contained in Appendix B are the details of the numerical scheme used to integrate this ice model in time.

The ice model requires the specification of surface elevation (H^*), ice thickness (H), and mass balance (G) as inputs. The grid size and time step are also user specified, and for this study, taken from our WRF simulations. With these atmospheric and ice models, we perform iterations, running WRF, using the mass balance calculated by WRF to force the ice model, updating the ice thickness and extent calculated by the ice model, and then running WRF again with the newly calculated ice

cover as an input. This amounts to an asynchronous coupling of the two models, where a WRF run followed by the ice model is considered an iteration.

WRF (Skamarock et al., 2008), version 3.7, is a compressible, non-hydrostatic model, which has been shown to provide a reasonable simulation of Arctic regions (Cassano et al., 2011; Kilpeläinen et al., 2011). Reanalysis data from the European Centre for Medium-Range Weather Forecasts (ECMWF, Dee et al., 2011) are used for setting the meteorological boundary and initial conditions, including sea surface temperatures and sea ice extent. Cassano et al. (2011) also noted that the Interim reanalysis (ERA-Interim) is a suitable choice of boundary conditions for WRF, leading to minimal bias in the Arctic, specifically in temperature.

We use the WRF Single Moment 6 Class microphysics scheme, which allows for ice, snow and graupel processes (Hong and Lim, 2006). NOAH-MP (Niu et al., 2011) is used for the land surface model. It includes 4 layers of snow and allows for refreezing, which is an important process on the Penny Ice Cap according to historical trends (Zdanowicz et al., 2012). NOAH-MP caps the snow water content at 2000 kg m^{-2} by default, but we remove this limit. The radiation scheme is the Rapid Radiative Transfer Model for GCM applications (RRTMG, Iacono et al., 2008), which has been found to have minimal bias in WRF (Cassano et al., 2011). CO_2 is set to 290 ppm for a glacial inception scenario (Petit et al., 1999; Vettoretti and Peltier, 2004). The boundary layer is simulated with the Mellor-Yamada-Janic scheme (MYJ, Janjic, 1994). We do not nudge to observations in any way other than prescribed lateral boundary conditions. Note that these physics are the same as those used in our previous paper (Birch et al., 2017), except now we use the Grell convective scheme (Grell and Dévényi, 2002), which has been shown to give good results in present day (Hines et al., 2011) and LGM (Bromwich et al., 2005) Arctic climates.

With this WRF configuration we find a reasonable reproduction of the ECMWF atmospheric state when we simulate present-day climate. Furthermore, precipitation on the coast of Baffin Island matches observations at that point of about 300 mm per year (Zdanowicz et al., 2012). Greenland is

initialized with present day ice cover and elevation based on studies finding little influence of Greenland on the build up of the Laurentide ice sheet (Kubatzki et al., 2006).

These resolutions are too coarse for a cloud resolving simulation, necessitating the activation of convective parameterizations. The incoming solar radiation is set to values consistent with 115 kya; the process of modifying the insolation in WRF based on the orbital parameters is detailed in Appendix B. We use realistic present day land type from the WRF database as initial conditions for the ice extent. Over these land grid points classified as ice, a snow water equivalent (SWE) of 5000 kg m^{-2} is prescribed — sufficient for maintaining snow cover for the duration of our runs and allowing us to capture a representative yearly mass balance for the chosen climatology.

3.2.1 ASYNCHRONOUS COUPLING PROCEDURE

We begin our WRF-ice flow coupled model experiment with the WRF control simulation, which uses meteorology from October 1985 to October 1986. The circulation during this year brought the coldest and wettest extremes to Baffin Island (Bromwich et al., 2002) and causes the present day ice cover to expand when combined with 115 kya insolation (Birch et al., 2017). WRF is run for 5 years, but we only analyze the last 4 years for an average climatology. Accumulation occurs over areas initialized with ice cover at high elevations, and net mass balance is significant enough to allow for ice growth (Fig. 3.1a). The control simulation agrees with our 4 km simulation of the Penny Ice Cap: cold meteorology causes the positive mass balance needed for the ice cap to expand. This mass balance is then adapted for use in the ice model.

For the first iteration of the ice model, initial ice thickness is adapted from present day ice cover from the NASA Operation IceBridge Mission (Fig. 3.1a). We use the surface elevation from the 20km resolution control WRF domain. The mass balance needed to force the ice model is calculated from the WRF simulation. Figure 3.1b depicts the mass balance results from the control WRF simulation with cold meteorology, 115 kya insolation, and present day ice cover. The mass balance

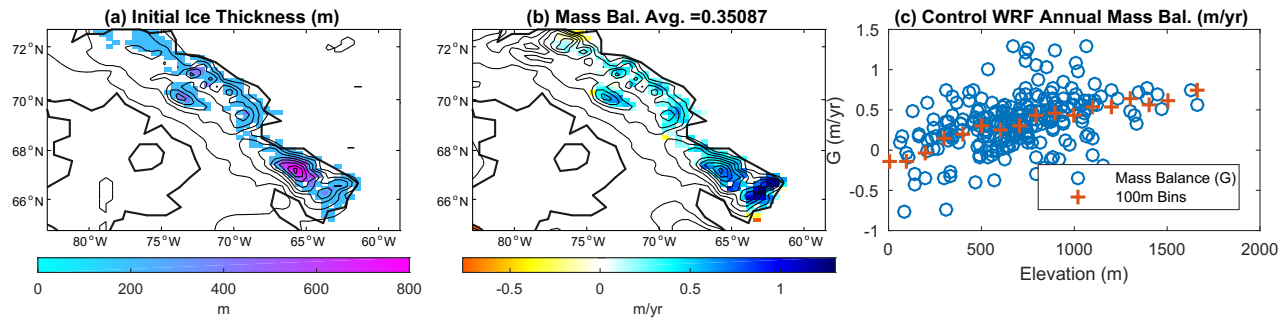


Figure 3.1: Ice thickness and Mass Balance for Ice Sheet Initialization: (a) Ice thickness was adapted from data collected by the NASA Operation IceBridge Mission (Kurtz and Farrell, 2011) with extent set by land use type in WRF. (b) Annual Mass Balance from WRF Control Run. (c) Binned Regression Visualization

over the ice caps is generally positive, except on the edges of the ice cap where there is net ablation.

Over the rest of the domain, WRF cannot calculate a meaningful annual mass balance because all snowfall received melts off during the summer. We characterize the mass balance as a function of elevation ($G(H^*)$), which allows the mass balance at each grid point to be updated as the surface elevation and ice thickness change every time step. This characterization involves sorting the domain by surface elevation and forming bins every 100 m. The average the mass balance in each bin is depicted by the purple crosses in Fig. 3.1c. Note the plateau in the mass balance at high elevations, which is expected from the elevation desert effect.

For our asynchronous coupling experiment, the ice model is run for 500 years, which was judged appropriate by Herrington and Poulsen (2011). After 500 years, the ice model has calculated a new ice extent and surface elevation (Fig. 3.2a) with which to force WRF. We import the surface elevation directly into the 20 km resolution WRF domain and use the ice extent to update the land type, i.e., the expansion the ice cap. Much like our control simulation, over areas classified as ice, we prescribe a snow depth of 5000. The surface elevation and land type of the 100 km resolution parent domain are adjusted through interpolation of the 20km resolution surface elevation and land type. Soil temperatures from the previous WRF run are used to initialize the next WRF iteration. Tests of spin up time for the soil temperatures lead us to agree with de Noblet et al. (1996) that present day soil tem-

peratures are too warm. Again using 115 kya insolation and cold meteorology, WRF is run for 5 years to characterize the mass balance on this new ice cap. This new run is the start of the next iteration. Thus, each iteration begins with a WRF run and ends after the 500 year ice model integration.

3.3 RESULTS

In the context of glacial inception, expanding ice cover interacts with the atmosphere, causing changes to both the regional climate of Baffin Island and the large-scale circulation. We first examine the progression of the WRF-ice model iterations, looking at the expansion of ice cover and the resultant changes in mass balance. Then, in Section 3.3.1, the changes in mass balance are explained by exploring changes in temperature, ablation, precipitation, and cloud cover. In section 3.3.2, we take a closer look at the large-scale circulation. Additional sensitivity simulations with the ice sheet model further clarify climate factors that could lead to Baffin Island glaciation in Section 3.3.5.

The progression of WRF-ice model iterations (5 years of WRF and 500 years of the ice model) is laid out in Figure 3.2. The ice thickness calculated by the ice model and then used in WRF is depicted in the left column, and the resulting annual mass balance from WRF in the right column. Also included in the right column is the spatial average of mass balance on the ice cap. Starting with the top left, we show the ice thickness resulting from running the ice model for 500 years using the control WRF mass balance. This new ice thickness, ice extent, and surface elevation computed by the ice model are used as inputs to the next WRF simulations. Then WRF is run for 5 years with the mass balance presented in the right column. After the second iteration of WRF, we bin the mass balance by elevation, and force the ice model with the new updated mass balance. A new ice cover is computed (left row, second column) and used as an input in WRF. Subsequent iterations pictured follow the same procedure.

The second iteration has the greatest ice extent, which is due to the rather high positive control

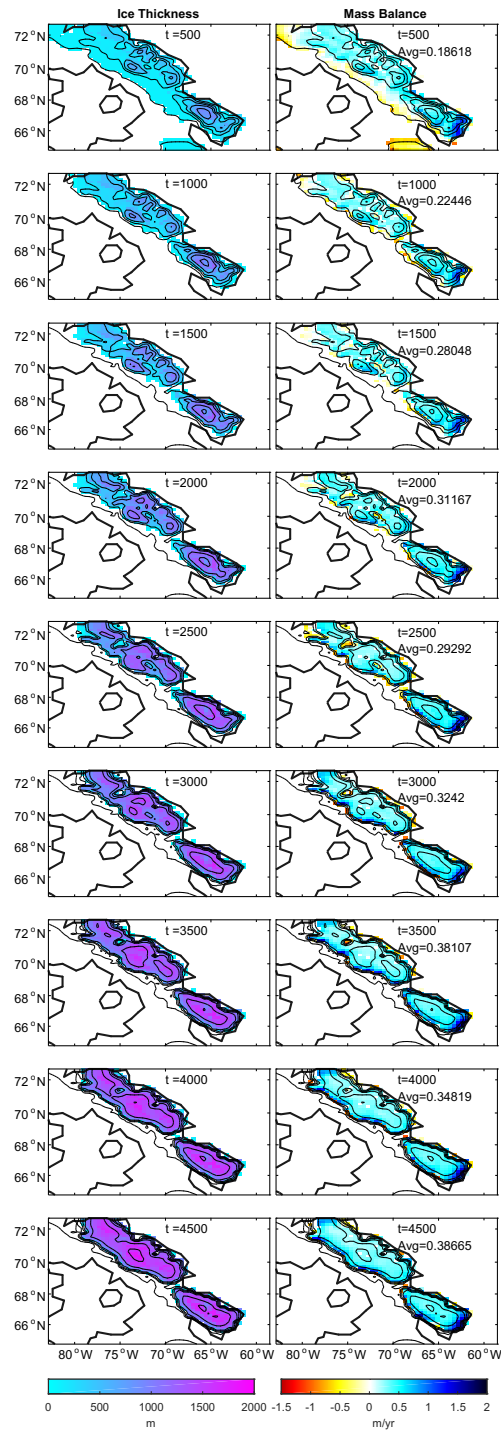


Figure 3.2: Ice growth due to WRF-ice model iterations. Left column is the ice thickness from the ice sheet model. Right column is the mass balance from WRF. The average mass balance on the ice cap is included.

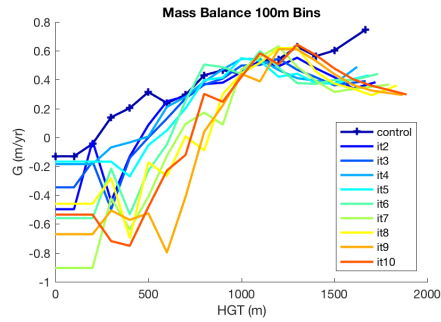


Figure 3.3: Characterizing annual mass balance vs. surface elevation. The domain is sorted into 100 m bins and within each bin the mass balance is averaged. Elevations falling within the range of the bin have the associated mass balance.

mass balance. In the third iteration, the ice retreats from its margins, though there is still growth in the highest parts of the domain. Additional iterations show that the ice cap grows in height, but expands slowly into new areas. After 5 kyrs, the ice accumulation on Baffin Island would only lower sea level by 0.6 m. The rest of our analysis focuses on why the ice does not expand quickly, though the average mass balance does seem to be generally increasing.

The binned mass balance used in the ice model is plotted in Figure 3.3 with annual mass balance on the y-axis and surface elevation on the x-axis. The control WRF run is the “+” blue line with the highest mass balance, particularly at the lowest and highest elevations. In subsequent WRF runs, the mass balance up to 1000 m becomes more negative at low elevations, which we find surprising. One might expect the equilibrium line to shift downslope as a result of the initial expansion of the ice cover, leading to a more positive mass balance. However, plotting the binned mass balance for all iterations shows a general trend of increasingly more negative mass balances at low elevations, with the equilibrium line altitude shifting upwards by several hundred meters over the course of ten iterations. As a result, the ice caps stagnate and fail to expand horizontally. The mass balance at highest elevations from all iterations also illustrate desertification, or the mass balance becoming less positive over time.

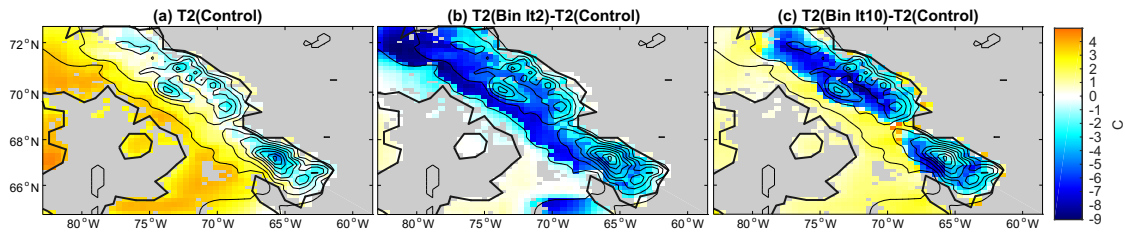


Figure 3.4: The Summer 2m Temperature in select asynchronous coupling iterations: (a) The temperature in control WRF. Differences in temperature between control and (b) second and (c) last iterations.

3.3.1 MECHANISMS OF CHANGES IN MASS BALANCE

A closer examination of temperature, melting, and precipitation can help us how the ice cap growth causes the changes to mass balance in the WRF simulations. We begin with the 2 m temperature on Baffin Island during the summer (June, July, August) when melting occurs. The differences in the temperature between the control run and subsequent iterations are pictured in Figure 3.4. In the control simulation, we note that most of the ice-covered land remains below freezing during the summer (blue). At the edge of the ice cap, temperatures generally remain near zero (white), and the low elevation tundra is above freezing (red). The largest expansion of ice cover occurs in the second iteration as noted above, which dramatically cools much of the region mostly due to the presence of snow and the albedo change. Also note that the upper part of the ice cap also cools, likely due to increasing elevation, since 500 years of ice growth increased the surface elevation by 200 m. In the third iteration, the ice has retreated somewhat, which makes the low elevations warmer than in the second iteration. The integrated mass balance becomes more positive as seen in Figure 3.2. As iterations of WRF and the ice model continue, we find that the area covered in ice cools, but the low-elevation tundra areas surrounding the ice cap appear to warm. The lapse rate effect due to increasing surface elevation progressively cools the ice cap, and below we explore why the low-elevation tundra warms in successive iterations.

The mass balance is made up of ablation and precipitation. When precipitation is larger than

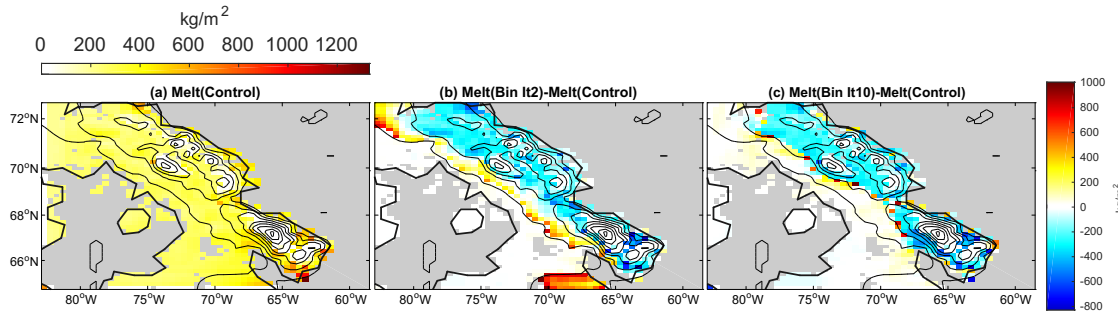


Figure 3.5: Melting in WRF in select iterations: (a) melting in the control WRF run. Differences in melting between control and (b) second and (c) last iterations.

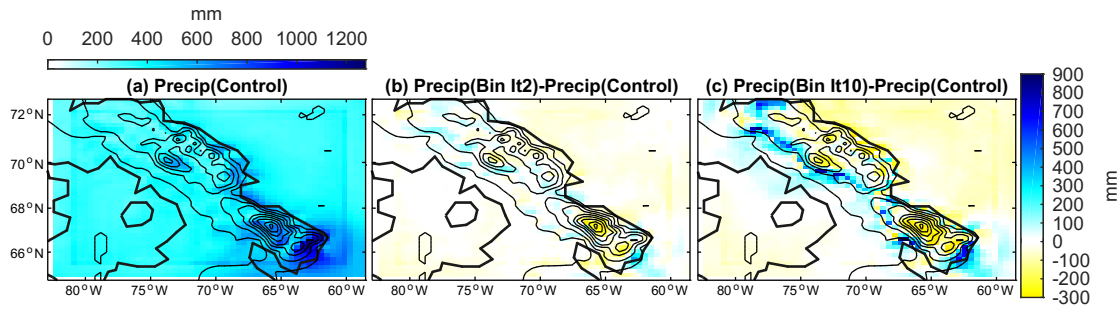


Figure 3.6: Annual precipitation in WRF in select iterations: (a) The precipitation in the control WRF run. Precipitation differences between control and (b) second and (c) last iterations.

ablation, the ice cap grows. We are interested in how both variables change over the iterations of the coupled WRF-ice sheet models. We start by examining the melting over the summer, and how it relates to the temperature discussed above. Figure 3.5a depicts the yearly melt in control simulation. In general, the high elevation parts of the region experience no melting during the summer, due to their sub-freezing temperatures. As the ice cover expands, the highest values of melting are located near the edges of the ice cap. As expected, melting decreases as the surface elevation increases.

Next, we look at the precipitation component of the mass balance. In this region, precipitation increases with elevation due to the orography (Figure 3.6). For glacial inception, continued snowfall

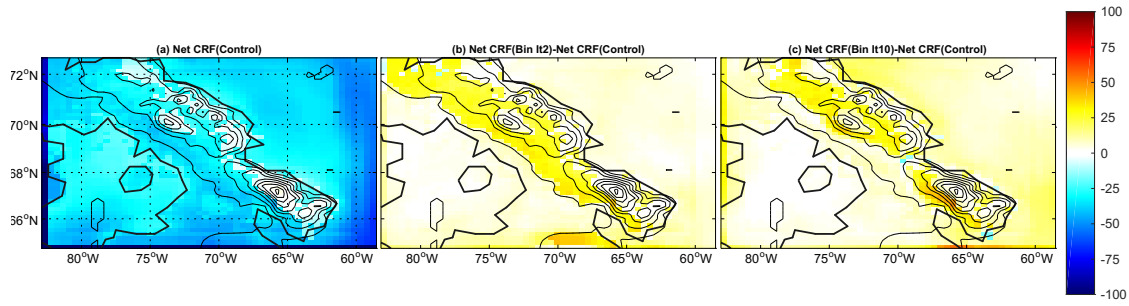


Figure 3.7: The Summer Cloud Radiative Forcing (CRF) in select iterations: (a) The CRF in the control WRF run. Differences in CRF in (b) second and (c) last iterations.

is critical in addition to low ablation. Figures 3.6b and c depict the differences in precipitation between the control simulation and ice model iterations. We see an elevation desert effect that reduces precipitation by 200 mm year^{-1} , which is $\sim 20\%$ of the annual precipitation. The highest elevations receive less precipitation as the iterations go on. However, at the edges of the ice cap, precipitation increases, because the change in elevation triggers orographic precipitation. It is interesting that the Barnes Ice Cap appears to be receiving increasing amounts of precipitation, which may be a necessary condition for glacial inception to occur. The accumulation increases on the western side of Baffin Island, which is also the windward side and as noted by Roe and Lindzen (2001b) a characteristic of the inception process. This general increase in precipitation near the margins agrees with the maps of the mass balance (Fig. 3.2) indicating the highest mass balances do not occur on the peaks of the island, but at about 1200 m elevation.

Finally, we are interested in the role that clouds may play in glacial inception. Jochum et al. (2012) and Birch et al. (2017) found clouds to be negative feedback during glacial inception. When insolation is changed from present day to that appropriate for the time of glacial inception at 115 kya, Jochum et al. (2012) noted that less low clouds form during glacial inception, while Birch et al. (2017) found that cloud properties change such that relative to present day, less shortwave is blocked from the surface. Our objective here is to examine this feedback during the integration of our asyn-

chronously coupled model. In Figure 3.7, the net CRF in the control simulation is negative during the summer because the clouds block shortwave radiation from reaching the surface. However, as the ice cover expands, the clouds block less shortwave radiation, illustrated by the positive net shortwave CRF difference between later iterations and the control run: the control simulation net CRF is more negative in all parts of the domain. The albedo of increasing snow cover accounts for less negative CRF over newly glaciated areas, but the change to the shortwave CRF in the rest of the domain is due to changes in cloud cover and properties.

3.3.2 INTERACTION OF BAFFIN ISLAND INCEPTION AND LARGE-SCALE ATMOSPHERIC CIRCULATION

We find the significant warming over Baffin Island to be surprising, as one might expect the expanded ice cover over the peaks of Baffin Island to lead to cooling (which would amplify the positive mass balance and lead to a larger ice sheet). We now analyze the warming by looking at the larger scale geopotential height field, temperature, and winds in the outer domain. First, the geopotential height and winds at 500 mbar for the control simulation during the summer are depicted in Fig. 3.8. The wind field over Baffin Island generally comes from the northwest with a low pressure trough covering the whole island. A weak cyclonic flow also exists over the island and Baffin Bay. Next we examine the subsequent iterations and their anomalies in the winds and geopotential height at 500 mbar in Figs. 3.8b-c. The pressure increases south of Baffin Island while it decreases over Baffin Bay to the north, and an anomalous anticyclone forms over the Hudson Bay region. Thus, the stronger southerly winds over East Canada due to this anticyclone response over the non-glaciated areas of Baffin Island bringing increased heat fluxes to the region, as previously observed in the warming of the inner domain. It seems that the larger ice caps on Baffin Island alter the circulation, and we will examine this hypothesis below.

The 2m temperature in the outer domain (Figure 3.9) confirms the findings based on the geopo-

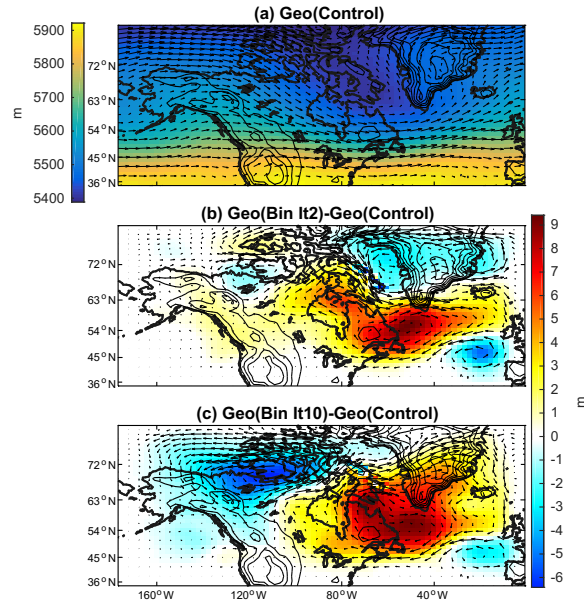


Figure 3.8: Changes in Geopotential in the Parent Domain in select iterations with wind vectors. Wind speeds are 6 m/s over Baffin Island. (a) The Geopotential at 500mbar for the control simulation with average largescale flow during the summer (b)-(f) Geopotential differences during the summer between control and iterations. Wind anomalies of 1 m/s are overlaid.

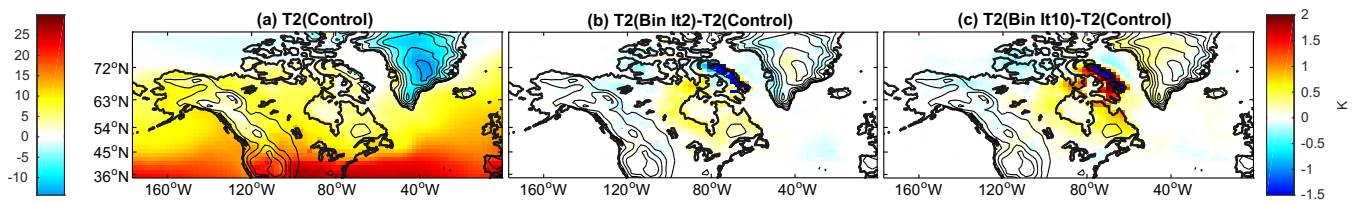


Figure 3.9: Summer 2m Temperature in the Parent Domain in select iterations: (a) The 2m Temperature for the control simulation (b)-(c) 2m Temperature differences during the summer between control and iterations 2 and 10.

tential height. The control simulation is depicted in the upper left with subsequent panels showing differences in the following iterations. On the ice caps of Baffin Island, the temperature generally grows colder because the elevation has increased, and newly glaciated areas cool due to both the presence of ice and the elevation increase. As iterations progress, the non-glaciated points of the island warm, which is also seen in the inner domain. Additionally, warming occurs outside of Baffin Island in the Hudson Bay region. These areas of warming agree with the increase in pressure found in the geopotential height and the anticyclonic motion with southerly warmer winds.

This warming appears to be a negative feedback on ice sheet growth, and it explains the mass balances becoming more negative at low elevations on Baffin Island through subsequent iterations (Fig. 3.3). With a region so close to freezing during the summer even a degree warming can prevent perennial snow from occurring during the critical inception period.

As a caveat, we note that with the topography in the coarser outer domain being generally lower than the 20km resolution domain, the temperatures on Baffin Island are much warmer than is likely realistic. For instance we do not see any summer temperatures significantly below freezing as found in the inner domain with more realistic topography.

3.3.3 THE OF IMPACT BAFFIN ISLAND ICE ELEVATION ON LARGE-SCALE CIRCULATION

Next, we diagnose the roles of ice and topography on circulation changes, by setting up sensitivity simulations based on the second iteration, which has changes in ice extent and surface elevation (Fig.3.2). For our “No H” scenario, we keep the surface elevation the same as the control simulation, but use the same ice extent as in the second iteration. In another scenario, “Tundra only”, the model is run with no ice anywhere to better understand the regional impact of ice. For an “H Only” scenario, we change the land type from ice to tundra, as in the “Tundra Only” simulation but set the surface elevation to the ice height in the second iteration. All of these simulations were run for 5 years with cold meteorology and 115 kya insolation and the analysis focuses on the last 4 years.

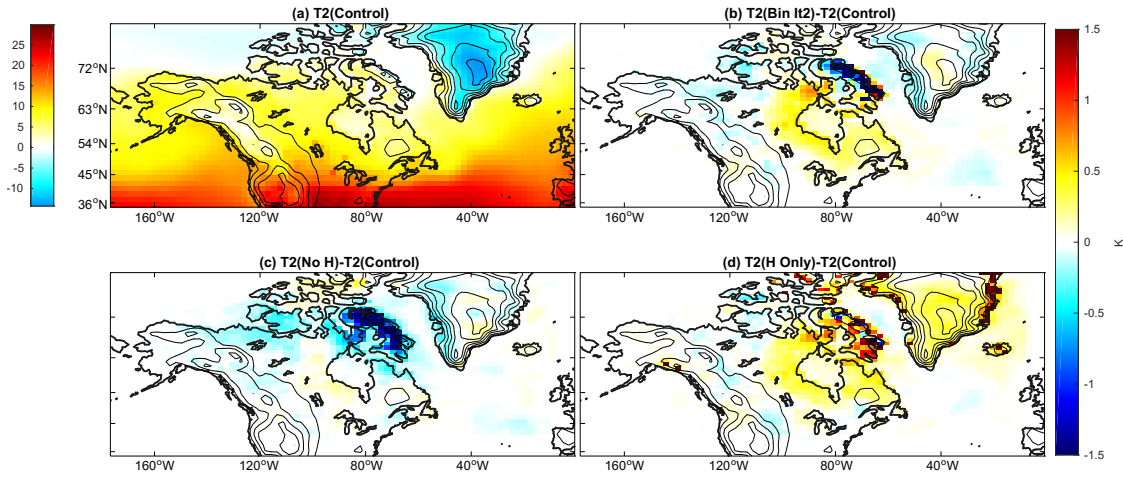


Figure 3.10: Changes in 2m Summer Temperature in the Parent Domain for topography and ice sensitivity simulations. (a) The 2m Temperature for the control simulation with averaged largescale flow during the summer (b) differences during the summer between control and second iterations. (c) differences between the control and No H. (d) differences between control and H Only. Wind anomalies are overlaid.

In Figure 3.10, we examine the temperature of the outer domain for 4 experiments. Again, the control simulation summer temperature and difference from the second iteration (having both ice and elevation changes) are shown, but we also include the “No H” and “H Only”, in c and d respectively. The “Tundra Only” is not shown, but as expected, Baffin Island is generally warmer. With “No H”, a significant regional cooling occurs, encompassing both Baffin Island and the Hudson Bay region, which is the pattern we had anticipated to cause rapid ice growth. However, with “H Only”, warming on both Baffin Island and around Hudson Bay occurs. Thus, the cause of the negative feedback due to large scale anticyclonic circulation and warming in our simulations is the change in topography. An interesting aspect of these last two simulations (“No H” and “H Only”) is that if we sum the difference between their response and the control simulation, the result is very close to the difference between iteration two and the control run, as seen in Fig. 3.11. This suggests that the responses to the ice topography and ice land type are linear.

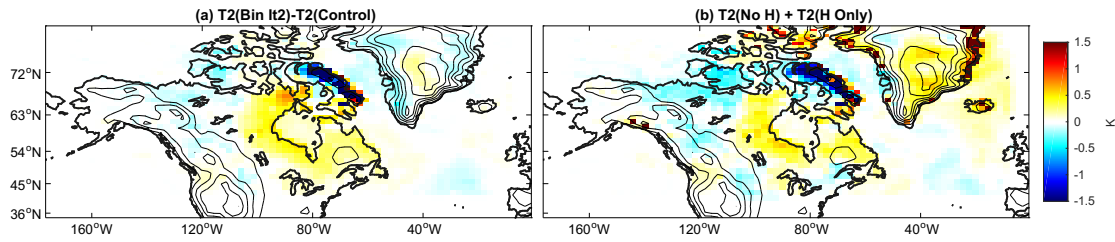


Figure 3.11: (a) Differences during the summer 2m temperature between control and second iterations. (b) Sum of temperature differences from “No H” and “Only H” scenarios, implying a linear response due to changes in ice cover and surface elevation

3.3.4 GLACIAL INCEPTION WITH CHANGING ORBITAL PARAMETERS

The WRF-ice model iterations performed previously used the minimum insolation from 115kya, which is useful for diagnosing the relationship between the ice cover and atmosphere. However, that is obviously not realistic. Figure 2.1a illustrates the insolation over the past 140 kyrs. Integrated insolation would have increased after its minimum at 115 kya, impacting temperature and glacial mass balance. We use the procedure from Appendix A to set the insolation in WRF for 115-111 kya. After each WRF run, the ice model is integrated for 500 years using the same mass balance procedure as detailed above. This set of simulations explores the mass balance that results from a combination of ice flow and insolation changes.

First, the progression of the ice cap every 500 years is pictured in Figure 3.12, beginning with the 3rd iteration at 114 kya. The first and second iterations with 115 kya insolation are pictured in Figure 3.2, as they are the same as our first set of iterations. Note that the spatial average of the mass balance is approximately the same for each successive iteration. These iterations progress in a similar manner as the case with constant insolation in that the lateral growth is slow though the height of the ice sheets does grow. Although, the high elevation plateaus in this set of simulations remain about 200 m less than those found in the previous set of simulations with constant 115 kya. We also note that at 112 kya these insolation simulations continue to maintain the average spatial mass balance while the

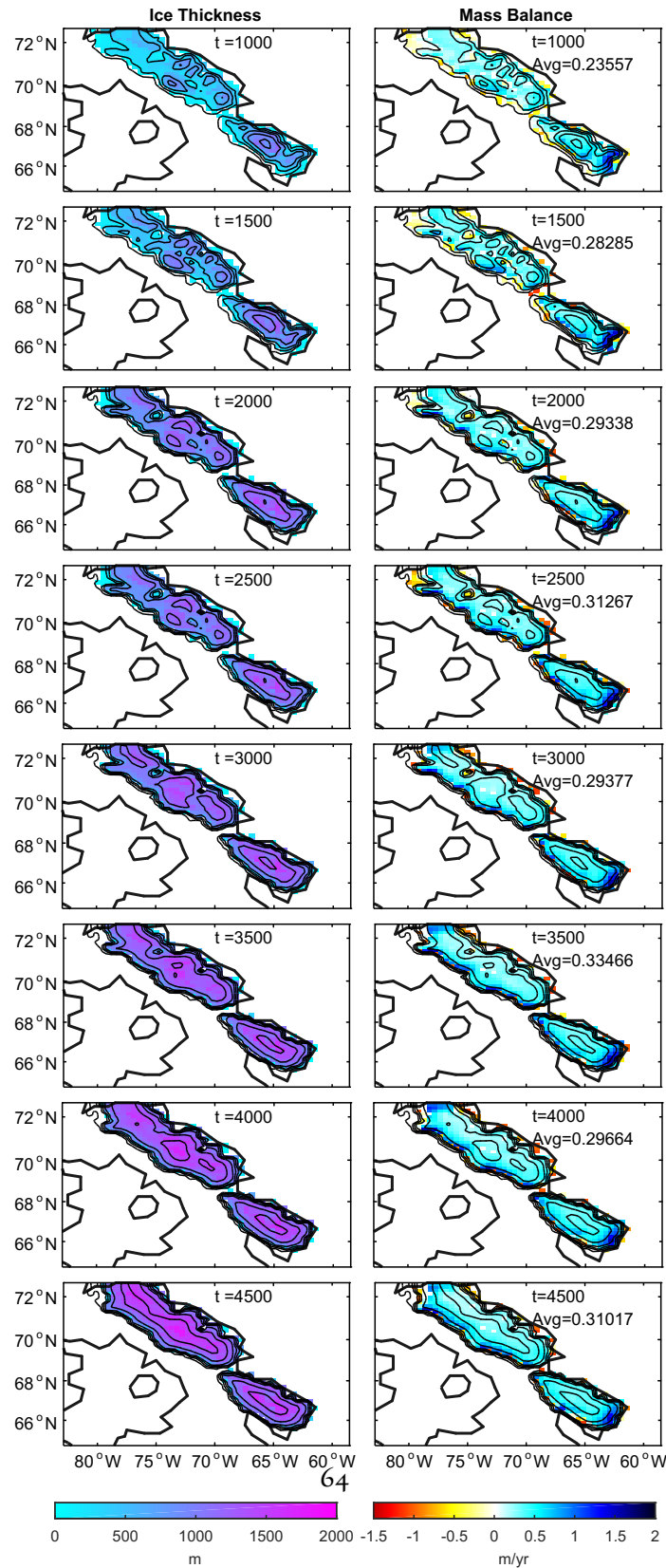


Figure 3.12: Ice growth due to WRF-ice model iterations with insolation changing every 1000 years. Left column is the ice thickness from the ice sheet model. Right column is the mass balance from WRF. The third iteration with insolation from 114 kya is the top row. Time progresses in each row, with the bottom row picturing the ice thickness and mass balance from 111 kya.

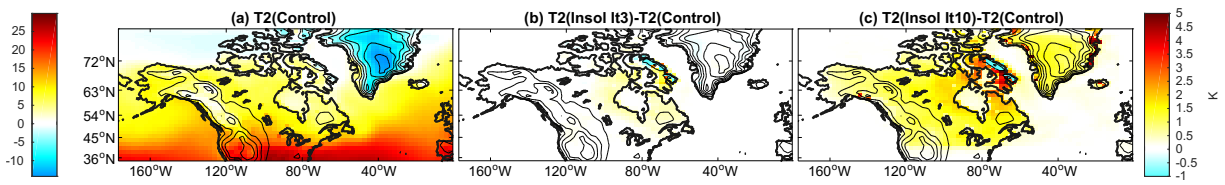


Figure 3.13: Summer 2m Temperature in the Parent Domain in select iterations: (a) The 2m Temperature for the control simulation (b)-(c) 2m Temperature differences during the summer between control and iterations 3 and 10, corresponding to 114 kya and 111 kya respectively.

115 kya insolation actually experiences an increase in the spatial average mass balance. This occurs due to the realistic insolation causing additional warming and a more negative mass balance at low elevations.

During the summer on Baffin Island, there is again cooling from the presence of ice, the unglaciated parts of Baffin Island warm considerably. In fact, as insolation increases slightly so does the temperature. When we look at the precipitation in the domain, we observe a similar pattern to the previous set of simulations. Accumulation increases on the edges of the ice cap due to orography. Then at the top of the ice cap, the elevation desert effect kicks in. The ice cap continually receives less precipitation than in the control simulation. The increases in insolation cause large-scale warming, and when combined with topography changes, the warming on Baffin Island can be over 5 K (Figure 3.13). Thus, the realistic insolation changes cause further warming during inception, which makes the occurrence of inception even more difficult to explain and simulate.

3.3.5 SENSITIVITY ANALYSIS OF THE ICE FLOW MODEL

The combination of WRF and the ice flow model does not indicate ice sheets would grow to cover all of Baffin Island, which was likely necessary to explain the sea-level drop found in the paleorecord. The change in ice surface elevation actually causes a negative feedback on inception by affecting the large scale circulation: temperatures around the ice increase inhibiting ice expansion. This result suggests a closer examination of the ice model and its sensitivities. In particular, we vary the creep

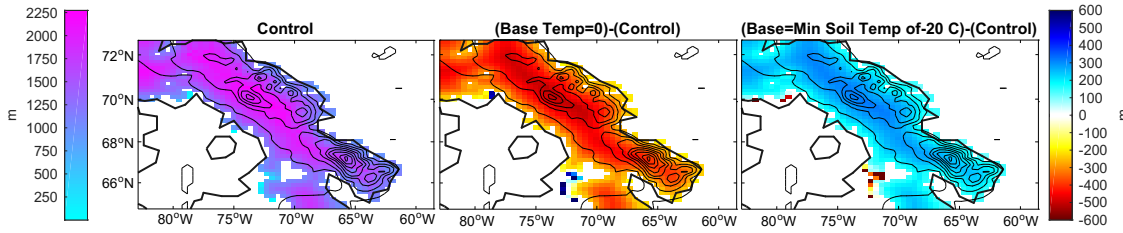


Figure 3.14: Sensitivity of the Ice Model to choice of constant creep parameter $A(T)$, which is related to the temperature profile within the ice as detailed by Cuffey and Paterson (2010). (a) A evaluated at -10°C or $A(-10^{\circ})\text{C}$. (b) $A(0^{\circ}\text{C})$. (c) $A(-15^{\circ})\text{C}$

parameter A to see if rapid ice expansion is possible with different constant choices. We also explore how a successful inception - matching paleorecord estimates - could occur from our simulated mass balance function $G(H)$ and prescribed calving. The ice model in these sensitivity experiments is now run for 10kyrs, unlike our previous asynchronous simulations that ran for 5 kyrs total.

As noted above, we choose the creep parameter ($A(T) = A(-10^{\circ}\text{C})$) based on Cuffey and Paterson (2010). We now test the sensitivity of the ice flow when $A(T)$ is evaluated at warmer and colder temperatures, or $A(0^{\circ}\text{C})$ and $A(-15^{\circ}\text{C})$ respectively. For these simulations, the mass balance used is from the control WRF simulations, and when run for 500 years, the ice model changes are minor. We confirm with WRF simulations that these slight changes in ice elevation from $A(0^{\circ}\text{C})$ and $A(-15^{\circ})$ do not significantly affect the mass balance in the second iteration; the ice still retreats. Thus, we want to examine how the creep parameter affects ice flow on longer timescales, which is why we integrate the ice model in time for 10 kyrs. We expect that a warmer temperature and thus larger A would cause the ice might spread more easily, leading to larger ice cover. Over 10kyrs, we can easily see that instead of accumulating on the peaks of Baffin Island, a somewhat wider, lower elevation plateau develops (Fig. 3.14b). On average, $A(-10^{\circ}\text{C})$ causes the elevation to be 200 m higher than $A(0^{\circ}\text{C})$. More calving occurs with $A(0^{\circ}\text{C})$ ice reaches the coast sooner. We also examine the impact of temperatures at about -15°C (Fig. 3.14c), which Marshall et al. (2002) determined to be the basal temperature of the initial ice sheet growth. As expected, the ice flows more slowly and

accumulates at high elevations.

Next we identify if we have simulated a mass balance sufficient for a successful inception. From Figure 3.3, we note that the mass balance in the control simulation is the most positive at high elevations and least negative at low elevations. All subsequent WRF runs simulate a negative mass balance at low surface elevations. Using the control mass balance, the ice model is integrated over 10 kyr. Of course, this situation is not realistic as our set of simulations show changes in mass balance due to circulation changes as the ice sheet grows. We prescribe the mass balance only on Baffin Island, which was the location of our inner domain, and this causes Baffin Island to be mostly covered by ice, corresponding to 1.75 m sea level drop over 10 kyr. Ice model simulations forced with the mass balances from iterations 2-10 do not reach the western coast of Baffin Island in 10 kyr.

We also question what would happen if we apply the control mass balance to a larger area, outside of Baffin Island and our chosen domain. Based on the temperature and precipitation patterns of the outer domain, the other side of Hudson Bay could have a similar mass balance, especially if the domain is resolved at a higher resolution. We do not include the Torngat Mountains on mainland Canada just to the south of Baffin Island, as this area appears to receive less precipitation and experience higher temperatures. This agrees with observations and other ice modeling simulations showing later glaciation in this area. The ice accumulation with the positive mass balance from our control WRF simulation on mainland Canada causes an equivalent sea level drop of 2.8 m in 10 kyr.

The previous simulations assume immediate calving over any ocean points, which makes sense for the kilometer deep Baffin Bay. The topography on the eastern side of Baffin Island is steep, possibly preventing ice shelves from forming. However, Hudson Bay is only about 200 m deep. We turn off immediate calving for this shallow water body, and instead prescribe a negative mass balance. We extend the regression $G(H)$ linearly to the bathymetry of Hudson Bay. This allows for ice to flow slowly into a larger area surrounding Baffin Island, creating a fairly large ice sheet and corresponding to a sea level drop of 6.6 m. This last scenario gets close to more recent constraints of sea

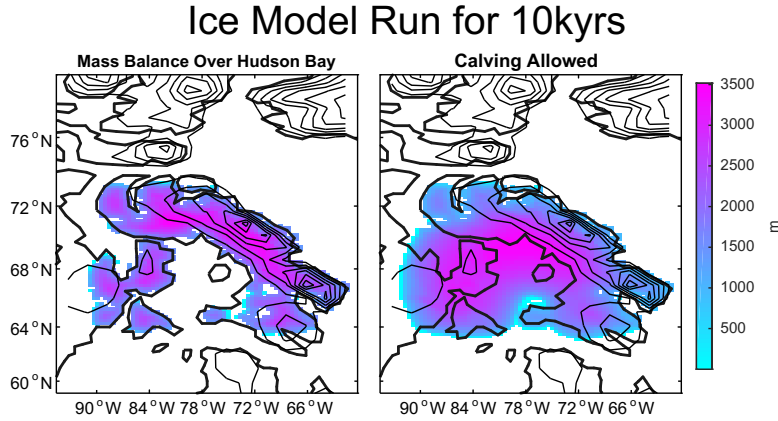


Figure 3.15: The result of running the ice model for 10kyrs with the maximum mass balance, found in the first WRF iteration. (a) The mass balance prescribed only over Baffin Island. (b) Allowing a positive mass balance on mainland Canada (c) Allowing ice flow into Hudson Bay, instead of calving.

level drop from modeling simulations at around 100 kya of 10m (Creveling et al., 2017). This growth of the ice sheet corresponds a successful inception, though slow, suggesting that the negative feedback due to the response of the large scale circulation to expanding ice cover is what prevents our asynchronously coupled model from achieving a successful inception.

3.4 DISCUSSION

In this chapter, we simulated the last glacial inception beginning at 115kya with a regional high resolution configuration of WRF, coupled asynchronously with a simple ice flow model. A net positive mass balance occurs in the 1st WRF iteration simulation driven by 115 kya insolation, cold present day (1986) boundary conditions, and present day ice cover, which could cover all of Baffin Island in ice sheets if that mass balance was maintained. However, updating the ice cover and surface elevation in WRF creates a negative feedback from changes in the large scale circulation: summer temperatures over Baffin Island increase due to a large-scale anticyclonic circulation. This causes

the mass balance over Baffin Island to become more negative at low elevations, stagnating further ice expansion in the ice flow model.

The flow over Baffin Island driven by the 1986 boundary conditions is consistently from the northwest during the summer, with weak speeds of about 6 m s^{-1} . The increase in the ice surface elevation after the 1st iterations of the coupled model causes an anticyclonic response of about 1 m s^{-1} to develop to the south of Baffin Island while a cyclone develops in the north, diverting the winds and bringing in warmer air from the south. This leads to net ablation and arrests the glaciation process. This circulation response is likely dependent on the initial flow (direction and speed) and topographic height (Chen and Lin, 2005; Cook and Held, 1988). Thus, a different meteorology choice could lead to sustained accumulation. Bromwich et al. (2002) identified exceptionally cold conditions on Baffin Island and their varying large scale circulation patterns. Generally, the cold summer circulation flows from the northwest or north-north-west. We examine the role of meteorology directly using boundary conditions that differ from those used in our cold meteorology simulations. For this purpose, WRF is driven using 1979 boundary conditions, which is an average year in regards to temperature and precipitation, and the flow over Baffin Island generally comes from the west. WRF simulates a negative mass balance for this year, and the ice sheet retreats to elevations above 1400 m when the ice model is forced with this mass balance for 500 years. In contrast, the ice cap from our average 1979 meteorology at a 4 km resolution was in approximate equilibrium, leaning towards slight expansion (Birch et al., 2017). Net ablation occurs in the 20 km resolution because of the lower topography. It is possible that during inception that the circulation pattern was different and allowed for both the initial ice expansion and continued cold conditions. WRF did not simulate such a pattern in this study, nor did we have a large sample of “cold” meteorological conditions in the ERA-Interim reanalysis.

We find significant warming on Baffin Island and the surrounding Hudson Bay region in response to the initial ice growth due to anti-cyclonic flow. The formation of anti-cyclones over ex-

panding ice sheets has been noted to allow cold northerly winds to flow over the ice sheet, enhancing inception (Roe and Lindzen, 2001b; Liakka et al., 2012). In particular, small ice sheets cause a wave response that enhances accumulation, while stationary waves induce less cooling as ice sheets increase in size (Liakka and Nilsson, 2010). In contrast, the anticyclone in our simulations forms south of the ice growing on Baffin Island, bringing significant warming even over these relatively small mountain glaciers. Similar to our study, Herrington and Poulsen (2011) found the Hudson Bay warmed in their coupled atmosphere-ice simulations due to anticyclonic flow. Gregory et al. (2012) also found warming on the margins from anticyclones and reduced cloudiness. In both of these studies, though, the warming only slowed growth; the regional cooling due to the growing ice sheet kept the ice advancing. We find the heat fluxes into the region by the developing anticyclone circulation to completely overpower any regional cooling from the snow albedo feedback, stagnating the ice sheet growth. We do observe the development of katabatic winds in response to ice sheet growth in our model, possibly contributing to the drying out of the region and decreasing cloud cover, but these winds do not cool the area outside of the glaciers sufficiently to allow for inception.

We do not find a positive feedback that would lead to a more positive mass balance due to the initial ice growth. However, some of our assumptions and modeling choices could be preventing such a positive feedback. For instance, our WRF simulation prescribes the SSTs and sea ice to those of 1986, but during glacial inception the SST would likely decrease and sea ice extent grow larger (Meissner et al., 2003). Allowing the SST to cool has been shown to induce cooling over land, even if the atmosphere is warm (Yoshimori et al., 2002), and Kageyama and Valdes (2000) noted that storm tracks change with colder SSTs. The amount of sea ice can impact accumulation over the ice sheet (Yin and Battisti, 2001; Kubatzki et al., 2006) due to decreasing atmospheric moisture and evaporation from ocean, according to the temperature precipitation feedback (Källén et al., 1979; Ghil, 1994; Gildor and Tziperman, 2000, 2001; Li et al., 2005). Some studies argued that the cooling during inception did not cause sea ice to extend outside of Baffin Bay until 80ky (Löffverström

et al., 2014), meaning the Atlantic remained ice free, which was conducive to high precipitation. Goñi et al. (2005) also argued that SSTs remained warm until 105 kya, when substantial ice sheets had formed. Both Khodri et al. (2001) and Wang and Mysak (2002) found that including ocean feedbacks caused increased moisture fluxes conducive to inception, while Jackson and Broccoli (2003) and Kaspar et al. (2007) noted that storm frequency increased. Including ocean feedbacks could lead to increased moisture fluxes and thus precipitation, or it could lead to cooler temperatures decreasing melting, which could be more important for inception. In either case, using present day ocean SSTs is likely inaccurate, but more work is needed to understand the impact of the ocean.

We also fix the CO₂ concentration to 290 ppm, although CO₂ decreased during inception. This decrease has been shown to influence the accumulation of northern hemisphere ice sheets (Ganopolski and Calov, 2011; Vettoretti and Peltier, 2004), accounting for up to 50% more ice (Bonelli et al., 2009). We also do not allow for dynamic vegetation, but on Baffin Island this should not matter. Most of these studies dealing with dynamic vegetation refer to the treeline moving southward during glacial inception (Harvey, 1989; Goñi et al., 2005), enhancing ice growth, but on Baffin Island the vegetation is mostly tundra and shrubs, which are not an impediment to snow accumulation as trees would be (Otterman et al., 1984).

We choose to drive our model with 1986 boundary conditions based on the anomalous increase in precipitation and cool temperatures over Baffin Island. The large-scale flow induced by these boundary conditions in conjunction with ice growth causes the negative feedback discussed above, but it is possible that a complete reorganization of the atmosphere could occur as part of the inception process, which we cannot capture with our regional model and prescribed boundary conditions. The Laurentide ice sheet and Fennoscandian ice sheet in Europe are thought to have begun growing at the same time though the Laurentide increased in volume more rapidly (Kleman et al., 2013; Ganopolski et al., 2010). The European ice sheet may have influenced the formation of the Laurentide (Beghin et al., 2014), but such a feedback is missing in our regional model. Roe and

Lindzen (2001b) noted that changes in the stationary wave pattern would affect the downstream climate, “the possibility of a complicated set of interactions in which each of the two major Northern Hemisphere ice sheet influences evolution of the other”.

We limit our study to Baffin Island based on geological records of it being the nucleation of the Laurentide, but using regional modeling on other Canadian mountain ranges could help identify how easily other areas glaciated, contributing to sea level drop. Growth of ice at high elevations simultaneously throughout Eastern Canada is not impossible, which could all coalesce into the Laurentide. For instance, various ice sheet models have shown ice beginning to accumulate all over Canada (Andrews and Mahaffy, 1976; Marshall and Clarke, 1999). While many of these ice modes do not match geological records because they simulate the growth of ice in the West, ice growth on many of the more eastern mountain ranges could perhaps impact the large-scale circulation in a manner conducive to inception.

Though we do find net accumulation in our control simulation using a 20 km resolution, there are inherent issues that come from this resolution. The topography of Baffin Island is very rough, but is greatly smoothed by the 20 km resolution. More importantly, the highest elevation peaks with 20 km resolution is ~ 400 m below the peaks found in realistic Baffin Island topography. Our simulation with average meteorology demonstrates that resolution changes can switch the mass balance from net accumulation to net ablation. For cold meteorology, the resulting mass balance is similar to what we found with a 4 km resolution (Birch et al., 2017).

To get an average mass balance to drive the ice sheet model, WRF is run for 5 years; we use the average mass balance from the last 4 years. We find that the land model required a 1 year spin up time, although often a spin up time of more than a year is recommended (Yang et al., 1995; Cosgrove et al., 2003). We tested running the model for longer than 10 years, but the simulations did not deviate much from the first 4 years. Deser et al. (2014) showed that multiple ensemble members of an atmospheric model simulation can give different results on a time scale up to 50 years. This may not be an

issue here because we are driving our regional model with boundary conditions that constrain the synoptic and planetary-scale flows (Denis et al., 2002).

Using the shallow ice approximation neglects important ice dynamics, in particular it neglects lateral stresses and does not allow for the development of ice streams, which cause rapid ice flow. It has been suggested that a finer grid size in the ice model could make for more accurate flow, as coarser resolutions have been shown to halt ice flow (Van den Berg et al., 2006). However, the shallow ice approximation was also suggested to cause too much ice flow (Le Meur et al., 2004). We are neglecting basal sliding that may develop as the ice base melts due to geothermal heat flux (Marshall and Clark, 2002). The initiating ice sheet was likely frozen at the bed (Marshall et al., 2000), which makes our choice appropriate. Finally, we are not accounting for glacial isostatic adjustment. Herrington and Poulsen (2011) found that the isostatic adjustment was not critical for glacial inception.

Though we find negative feedbacks on the growth of ice sheets due to clouds and large-scale circulation preventing inception in our model, high resolution regional modeling still has potential to make progress on the inception problem. We have captured the mass balance on critical mountainous regions in our control run, which GCMs cannot simulate accurately. When combined with ice flow, the control mass balance causes a significant portion of Baffin Island and the surrounding area to become ice covered. Though the equivalent sea level drop is only 6 m in 10 kyrs, the geological constraints on sea level drop from the Laurentide are not tight. We do find this equivalent sea level drop is closing in on more recent sea level estimates (Creveling et al., 2017). The temperature and precipitation in this scenario is not very different from present day, meaning maintaining this meteorology could be all that is necessary for inception. The warming feedback of the large-scale circulation could be mitigated by a reorganization of the atmosphere, which is very likely. Using a circumpolar regional model to capture the mass balance of the nucleation points of both the Laurentide and Fennoscandian ice sheets is a logical next step.

4

Conclusions

In Earth's recent history, the climate has transitioned into an ice age around every 100 kyrs. GCMs have difficulty simulating glacial inception, in part due to their coarse resolution. I addressed this issue using a high-resolution regional atmospheric model and an ice flow model. I focused on three stages of glacial inception: positive mass balance on current ice caps, ice flow downslope from successive years of accumulation, and the ice sheet-atmosphere interaction that affects further changes in elevation and extent.

In Chapter 2, I examined the sensitivity of mountain glaciers on the Penny Ice Cap on Baffin Island to insolation, meteorology, and topography. This study is unique due to the use of a 4 km resolution regional atmospheric model (WRF). It captures two aspects neglected by GCMs: realistic topography and resolving clouds without the need for convective parameterizations. I found that net accumulation occurs with 115 kya insolation, cold and wet meteorology obtained by driving the regional model with 1986 boundary conditions, and realistic topography. The cooling imposed by these boundary conditions that cause net snow accumulation is only 0.5 degrees cooler than average temperatures over the past 30 years, and the precipitation is only increased by 7%. This moderate change in climate causes the ELA to lower by 200 m and suggests that the temperature and precipitation necessary for inception may not need to be extreme.

Most notably, I found that the realistic topography is critical for inducing ice cap growth in models. The 4km resolution topography peaks are close to 2000 m (as is the observed elevation of the Penny Ice Cap), while the maximum elevation in typical GCMs with a 1 degree resolution is 400 m on Baffin Island. With realistic topography, the model simulated mountain glaciers with accumulation and ablation zones, including a realistic equilibrium line altitude (ELA). In my simulation with low GCM-like topography net melting occurred on the ice cap, and there was no net accumulation zone. This result implies that excessive melting is simulated in GCMs for glacial inception scenarios and also that the melting on mountain glaciers could be overestimated in future climate scenarios. Clouds were confirmed to be a negative feedback during glacial inception, in agreement with previous GCM studies. Compared to clouds today, less shortwave radiation was blocked during the summer when 115 kya insolation was prescribed. This was due to changes in the cloud integrated liquid water content. I also found, though, that cloud cover increased over low GCM-like topography, which could increase the cooling effect of clouds. Thus, the lowered topography can cause two competing effects in GCMs: increased melting from artificially low elevations and cooling from increased cloud cover. This suggests compensating errors that are known to lead to deteriorated pre-

diction skill, because these errors may stop compensating in a future climate change scenario and increase uncertainty. This leads one to wonder if these compensating errors allow coarse resolution GCMs to produce inception where they should not.

Analyzing the mass balance by elevation was used to implicitly capture the second stage of inception, as WRF is not coupled with an ice flow model. I integrated the mass balance of the 4km Penny Ice Cap simulation downslope, and I found that the mountain glaciers would expand significantly when forced by cold meteorology. Also, by prescribing varying amounts of ice cover over the domain, I found, as expected, that the snow albedo feedback induced regional cooling by lowering the ELA.

I then explored in Chapter 3 the second stage of glaciation more fully using an ice flow model based on the shallow ice approximation, which was coupled with the atmospheric model WRF. I again used a regional configuration of WRF to simulate a higher resolution than has been previously attempted for the inception problem. At 20 km resolution the topography was fairly well resolved and only about 400 m below the observed surface elevation. Also the outer domain encompassed all of North America at 100 km resolution, which allowed for an examination of the large-scale response. Asynchronously coupling an ice model with WRF lead to initial expansion of the Barnes and Penny ice caps, but that growth then stagnated. Though the ice caps grew in height, the lateral ice flow was too slow for the 115 kya inception of Baffin Island.

The third stage of glaciation, i.e., the interaction of growing ice sheets and the atmosphere, was also explored through asynchronous coupling. I found that the growing elevation of ice sheet induced a negative feedback on its continued horizontal expansion. Anticyclone development to south of Baffin Island caused warm southerly winds to flow over Baffin Island, which increased ablation and stagnates the expansion of the ice sheets. The snow albedo feedback did provide regional cooling, but was dominated by the warming induced by the large-scale circulation change. Cloud cover decreased and contributed to warming during the summer. The elevation-desert effect was

observed in decreased precipitation on the peaks of the ice caps. There was orographically enhanced precipitation on the margins of the ice sheet, particularly over the Barnes Ice Cap. This is a positive sign of how increasing precipitation can amplify ice growth, but it was not enough to overcome the ablation rates induced by the large-scale circulation.

The highest net positive mass balance occurred in the 1st iteration of WRF during these asynchronous atmosphere-ice modeling experiments. When we used this mass balance in the ice model – albeit with arguably low calving in the Hudson Bay – a sizable ice sheet grew, which was equivalent to 6.6 m sea level drop in 10 kyrs. While on the low end of some observational estimates, this ice growth agrees with more recent geological evidence, especially as the contribution of the Laurentide to the sea level drop is uncertain. Thus, consecutive years of moderately cold, wet meteorology – not so different from today – can lead to inception, but the mechanism by which the atmospheric circulation remains in this state is not clear. Ocean feedbacks or changing atmospheric stationary waves due to the growing Fennoscandian ice sheet may provide the positive feedback necessary for cold temperatures, increased precipitation, and the growth of the Laurentide Ice Sheet.

I conclude that there is potential in regional climate modeling to capture the dynamics of glacial inception, given the importance of realistic topography. I find that an accurate mass balance, leading to net accumulation, only occurs with resolved topography and requires moderately cold meteorology in conjunction with 115 kya insolation. The response of the large-scale circulation to surface elevation changes reveals that Baffin Island may play a crucial role in the large-scale circulation pattern during glacial inception. The role of negative feedbacks on inception due to changes in the large-scale circulation requires further study.



Appendix to Chapter 2

The incoming insolation varies over thousands of years according to changes Earth's orbital parameters (Berger and Loutre, 1991): obliquity (β), eccentricity (ϵ), and precession (ω). Within the WRF radiation module, we are able to change the obliquity explicitly, but to account for eccentricity and precession, WRF uses an eccentricity factor (Paltridge and Platt, 1976), which is multiplied by the solar constant. This factor is a Fourier expansion of the square of the ratio of present day mean distance from the sun ($\bar{R}^*$) to the actual sun-Earth distance (R^*).

$$\left(\frac{\bar{R}^*}{R^*}\right)^2 = 1.000110 + 0.34221 \cos \theta_0 + 0.001280 \sin \theta_0 + 0.000719 \cos 2\theta_0 + 0.000077 \sin 2\theta_0 \quad (\text{A.1})$$

where $\theta_0 = 2 * \pi * t / 365$ approximates the position in the orbit and t is the Julian date ($t = [1, 365]$).

To modify Equation A.1, we use the ratio of the semi major axis ($\bar{R}^*$, same as above) to the distance from the sun (R^*).

$$\left(\frac{\bar{R}^*}{R^*}\right)^2 = \frac{(1 + \epsilon \cos \theta)^2}{(1 - \epsilon^2)^2}, \quad (\text{A.2})$$

where again ϵ is the eccentricity and θ is the position in the elliptic orbit. With θ and ϵ , we fit a Fourier expansion to Equation A.2 like (Paltridge and Platt, 1976) did. To apply precession, we simply shift the curve according to the difference in present day precession angle and past precession angle. We validate this method with present day orbital parameters from Table A.1, and then apply this method to yield 115 kya and 128 kya insolation.

Table A.1: Orbital parameters, as calculated by Berger and Loutre (1991): obliquity (β) in degrees, eccentricity (ϵ), and longitude of perihelion from the moving vernal equinox (precession angle, ω) in degrees.

Insolation	Time	β	ϵ	ω
Minimum (J)	115 kya	22.438	0.043983	109.54
Mean ($+\delta J$)	0 kya	23.446	0.017236	101.37
Maximum ($+ + \delta J$)	128 kya	24.142	0.041094	257.12

B

Appendix to Chapter 3

B.1 SHALLOW ICE MODEL SET-UP

The simple ice sheet model is based off of Oerlemans (1981), where ice flow is represented by a non-linear diffusion equation and the diffusivity contains a constant term called the creep factor. This parameter determines the ice viscosity and depends on the temperature of the ice (Cuffey and Paterson, 2010), meaning it can be denoted $A(T)$, where

$$A(T) = A_0 e^{-Q/RT}$$

If $T < -10^\circ\text{C}$, then $A_0 = 3.61 \times 10^{-13} \text{ Pa}^{-3}/\text{s}$ and $Q = 60 \times 10^3 \text{ J/mol}$, but if $T \geq -10^\circ\text{C}$, then $A_0 = 1.734 \times 10^3 \text{ Pa}^{-3}/\text{s}$ and $Q = 139 \times 10^3 \text{ J/mol}$.

The surface of the ice is colder than the base, meaning that throughout a real ice sheet temperature, and thus $A(T)$, vary. In fact, the basal temperature often reaches 0°C for thick enough ice sheets. For an ice model using the shallow ice approximation the standard procedure is to take the vertical average of $A(T)$ (Weertman, 1957; Goodman and Pierrehumbert, 2003), where

$$A(T) = \frac{1}{h} A(T(z)) dz$$

Since $A(T(z))$ varies by height, we have to calculate a vertical temperature profile of our ice sheet. We use the average temperature of the ice from WRF and the geothermal heat flux of 55 W m^{-1} and assume a linear vertical temperature profile. Then we evaluate $A(T)$ for this temperature profile as found in

and take the vertical average. Typically, this vertical average captures $A(T)$ at warmer temperatures at the base of ice sheet, since A does not vary linearly but exponentially. Thus, it can be an order of magnitude larger at the base of the temperature compared to the surface of ice sheet. For our simulations, we calculate $A(T) =$, which is the same as $A(-10^\circ\text{C})$ from Cuffey and Paterson (2010). -10°C also turns out to be the basal temperature of ice sheets at the beginning of inception Marshall et al. (2002). Since the basal temperatures are warmer than the surface and flow is more likely at the base with warmer temperatures, we feel confident in our choice of A .

FINITE DIFFERENCE APPROXIMATION

We can use the flux form for the finite difference approximation. We use forward Euler and second order centered difference, where n is the time step, i is the node in the x direction, and j is the node in the y direction.

$$\frac{H_{i,j}^{n+1} - H_{i,j}^n}{\Delta t} = \frac{D_{i+\frac{1}{2},j}^n \left(\frac{H_{i+1,j}^{*n} - H_{i,j}^{*n}}{\Delta x_i} \right) + D_{i-\frac{1}{2},j}^n \left(\frac{H_{i,j}^{*n} - H_{i-1,j}^{*n}}{\Delta x_{i-1}} \right)}{\frac{1}{2}(\Delta x_i + \Delta x_{i-1})} \quad (\text{B.1})$$

$$+ \frac{D_{i,j+\frac{1}{2}}^n \left(\frac{H_{i,j+1}^{*n} - H_{i,j}^{*n}}{\Delta y_j} \right) + D_{i,j-\frac{1}{2}}^n \left(\frac{H_{i,j}^{*n} - H_{i,j-1}^{*n}}{\Delta y_{j-1}} \right)}{\frac{1}{2}(\Delta y_j + \Delta y_{j-1})} + G \quad (\text{B.2})$$

where

$$D_{i+\frac{1}{2},j}^n = AH_{i+\frac{1}{2},j}^{m+2} \left[\left(\frac{H_{i+1,j}^* - H_{i,j}^*}{\Delta x_i} \right)^2 + \left(\frac{1}{2} \left(\frac{H_{i+1,j+1}^* - H_{i+1,j-1}^*}{\Delta y_{j-1} + \Delta y_j} + \frac{H_{i,j+1}^* - H_{i,j-1}^*}{\Delta y_{j-1} + \Delta y_j} \right) \right)^2 \right]^{(m-1)/2} \quad (\text{B.3})$$

$$D_{i-\frac{1}{2},j}^n = AH_{i-\frac{1}{2},j}^{m+2} \left[\left(\frac{H_{i,j}^* - H_{i-1,j}^*}{\Delta x_{i-1}} \right)^2 + \left(\frac{1}{2} \left(\frac{H_{i,j+1}^* - H_{i,j-1}^*}{\Delta y_{j-1} + \Delta y_j} + \frac{H_{i-1,j+1}^* - H_{i-1,j-1}^*}{\Delta y_{j-1} + \Delta y_j} \right) \right)^2 \right]^{(m-1)/2} \quad (\text{B.4})$$

$$D_{i,j+\frac{1}{2}}^n = AH_{i,j+\frac{1}{2}}^{m+2} \left[\left(\frac{1}{2} \left(\frac{H_{i+1,j+1}^* - H_{i-1,j+1}^*}{\Delta x_{i-1} + \Delta x_i} + \frac{H_{i+1,j}^* - H_{i-1,j}^*}{\Delta x_{i-1} + \Delta x_i} \right) \right)^2 + \left(\frac{H_{i,j+1}^* - H_{i,j}^*}{\Delta y_j} \right)^2 \right]^{(m-1)/2} \quad (\text{B.5})$$

$$D_{i,j-\frac{1}{2}}^n = AH_{i,j-\frac{1}{2}}^{m+2} \left[\left(\frac{1}{2} \left(\frac{H_{i+1,j+1}^* - H_{i-1,j+1}^*}{\Delta x_{i-1} + \Delta x_i} + \frac{H_{i+1,j}^* - H_{i-1,j}^*}{\Delta x_{i-1} + \Delta x_i} \right) \right)^2 + \left(\frac{H_{i,j}^* - H_{i,j-1}^*}{\Delta y_{j-1}} \right)^2 \right]^{(m-1)/2} \quad (\text{B.6})$$

and

$$H_{i+\frac{1}{2},j}^{m+2} = \left(\frac{H_{i+1,j} + H_{i,j}}{2} \right)^{m+2}$$

$$H_{i-\frac{1}{2},j}^{m+2} = \left(\frac{H_{i,j} + H_{i-1,j}}{2} \right)^{m+2}$$

$$H_{i,j+\frac{1}{2}}^{m+2} = \left(\frac{H_{i,j+1} + H_{i,j}}{2} \right)^{m+2}$$

$$H_{i,j-\frac{1}{2}}^{m+2} = \left(\frac{H_{i,j} + H_{i,j-1}}{2} \right)^{m+2}$$

This model uses a time step of one year. At each time step, a new ice thickness is calculated based on the equations above. Then with the new ice thickness and basal topography known, a new surface elevation is calculated ($H^* = B + H$).

The edge and center of the ice sheet can cause $D = 0$, which is unrealistic behavior. Thus, we set a minimum D , but as noted by Oerlemans (1981), the ice model does not depend strongly on this choice.

We also characterize the mass balance according to elevation, where G is actually $G(H^*)$. Thus, the mass balance is also updated after each time step, since it depends on H^* . For our main set of simulations, we characterized the mass balance using elevation bins.

B.2 ADDITIONAL PLOTS OF ALL ITERATIONS

The precipitation over Baffin Island illustrates how the margins of the ice cap gradually receive more precipitation as the elevation increases (Fig. B.1). We also present the geopotential of the outer domain showing the anticyclone through all simulations in Fig. B.2.

The precipitation in the outer domain is shown in Figure B.3 with the control in the upper left and the remaining plots depict the anomalies from the iterations. The precipitation in the outer domain is much less over Baffin Island than the inner domain, but the outer domain gives us a picture of the moisture going into the inner domain. The tundra regions of Baffin Island and the Hudson Bay area receive noticeably less precipitation as the iterations go on. The atmospheric circulation has shifted such that precipitation in critical areas decreases. Note that the outer domain actually has precipitation increases on the ice cap, the opposite of the inner domain, due to orographic precipitation. GCMs with domain sizes even coarser than this one may have issues diagnosing orographic precipitation.

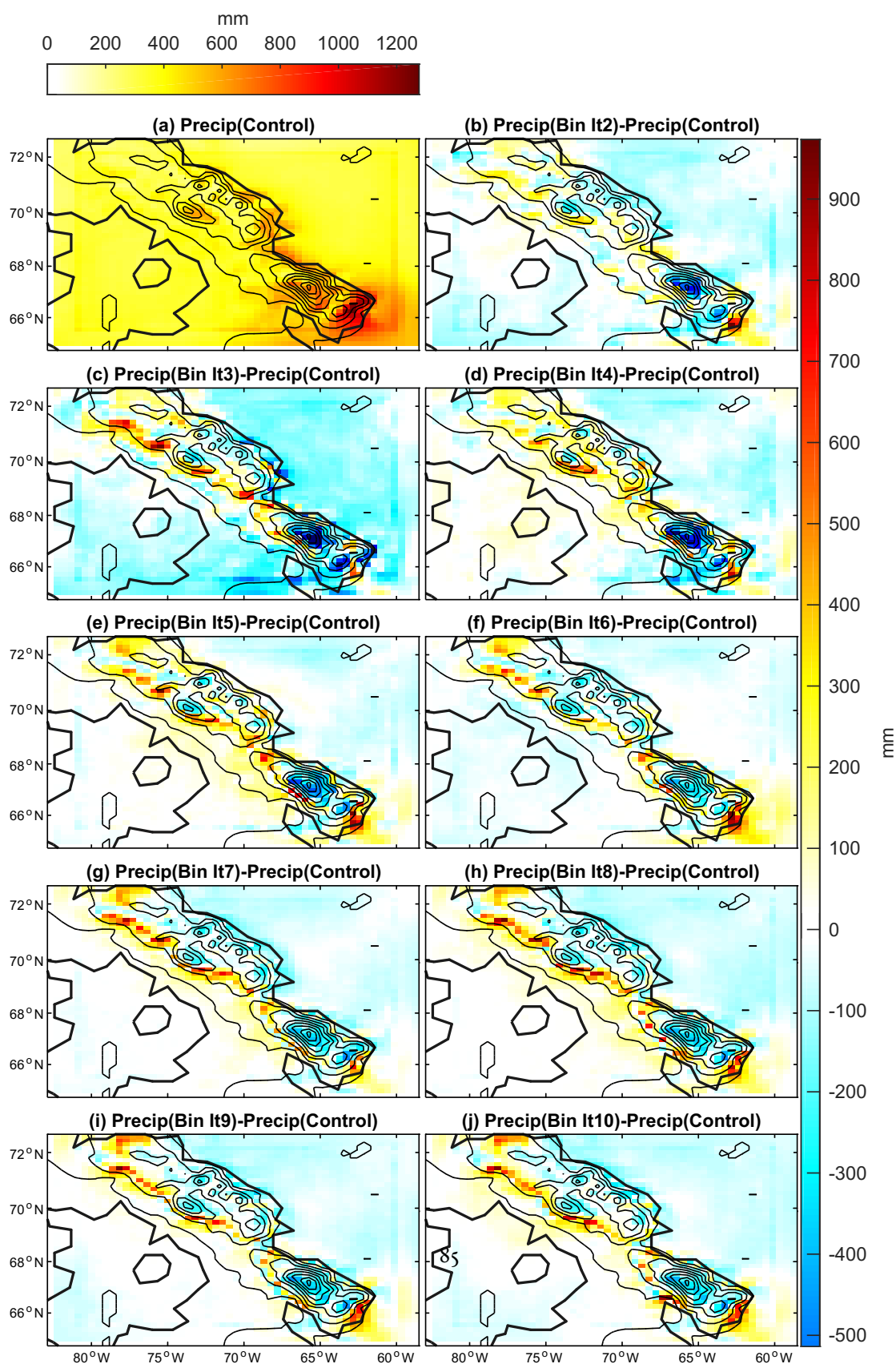


Figure B.1: Annual Precipitation on Baffin Island: (a) The Precipitation for the control simulation. Note the influence of orography. (b)-(j) Precipitation differences between control and all iterations.

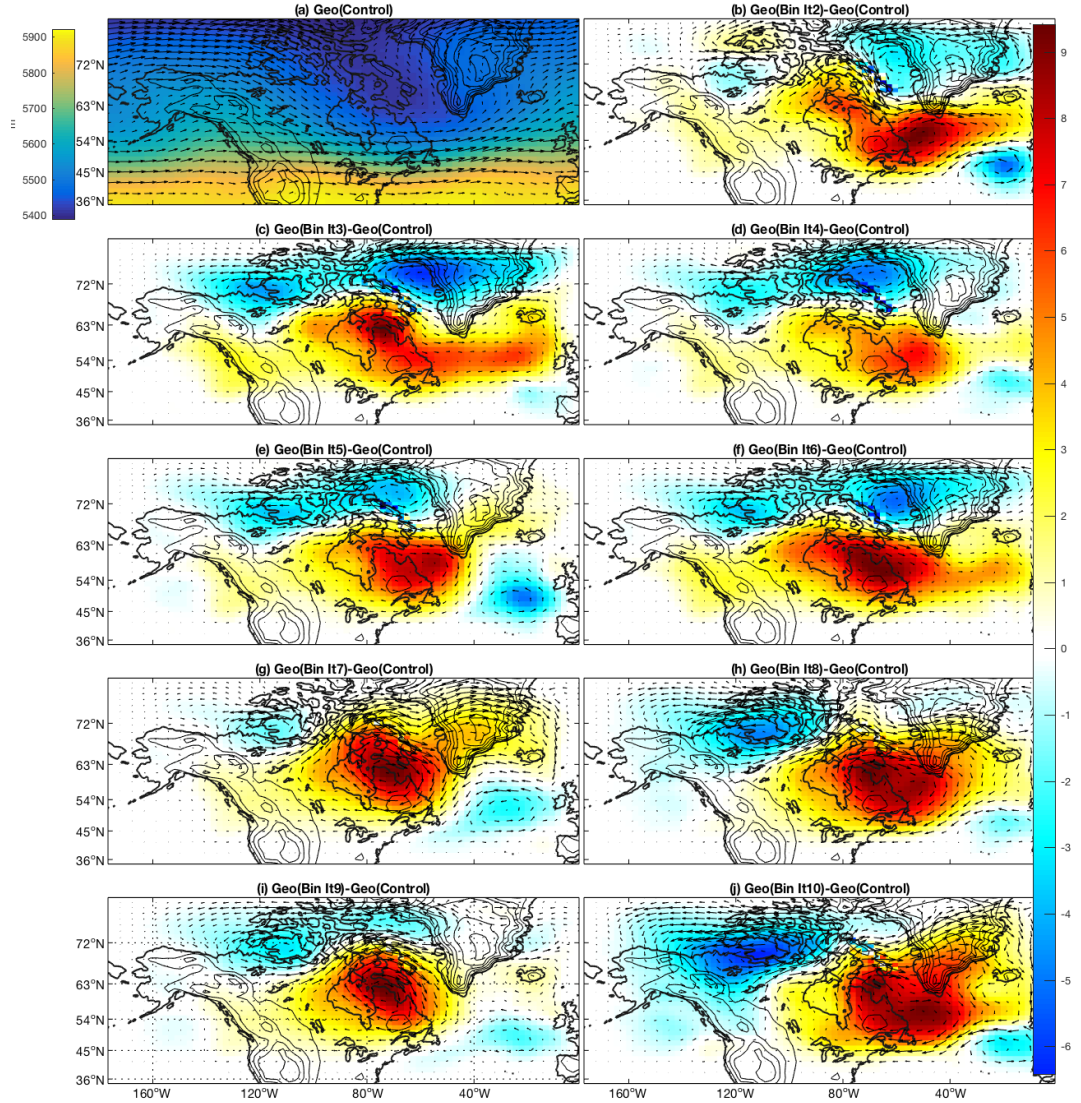


Figure B.2: (a) Summer 500 mbar Geopotential Height in the Parent Domain with wind vectors. Wind speeds are 6 m/s over Baffin Island. (b)-(j) Geopotential differences during the summer between the control and subsequent iterations. Wind anomalies of 1 m/s are overlaid.

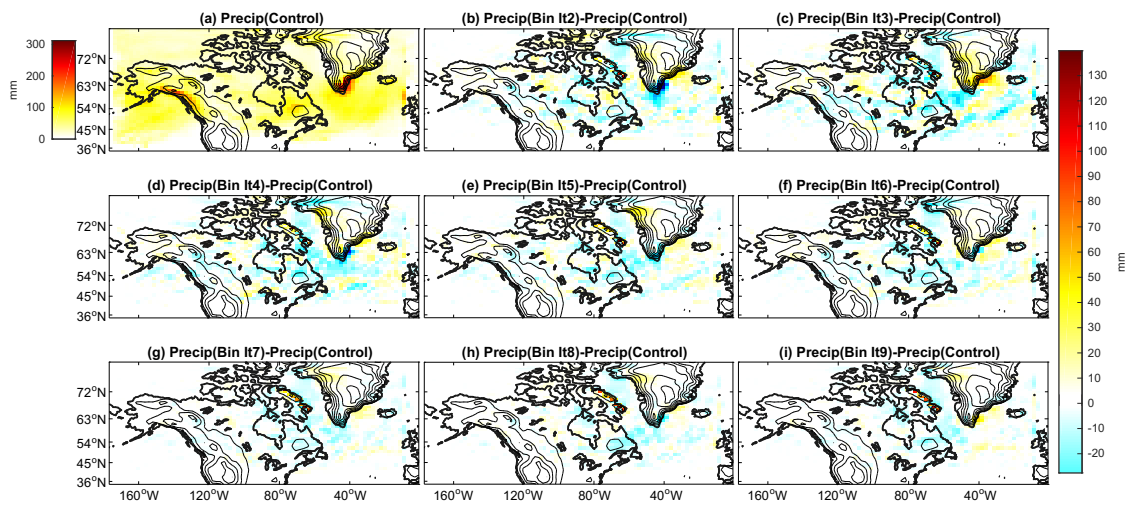


Figure B.3: Annual Precipitation in the Parent Domain: (a) The Annual Precipitation for the control simulation (b)-(i) Precipitation differences during the summer between control and iterations.

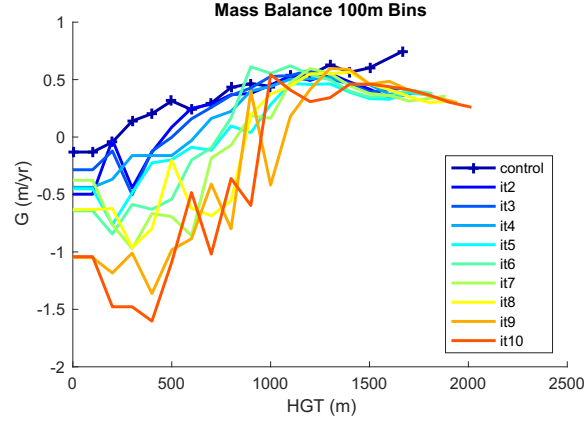


Figure B.4: The Mass Balance Binned by Elevation used in simulations with realistic insolation from 115 kya to 111 kya, where iteration 2 uses 115 kya insolation and iteration 10 uses 111 kya.

B.2.1 INSOLATION VARIED FROM 115KYA TO 111 KYA

During the last glacial inception, insolation did not remain at the 115 kya minimum as used in our main set of simulations. According to the integrated insolation in Fig., the insolation increased as did the summer temperatures. We test this hypothesized negative feedback by updating the insolation every 1,000 years. The simulations proceed as we expected with warming enhanced and the ice sheet being further stagnated as seen in the main chapter. However, Fig. B.4 illustrates that the margins of the ice cap and also the tundra are much warmer here than the temperature in our main set of simulations Fig 3.3. The mass balance becomes even more negative at low elevations when insolation is realistically changed.

We also include the temperature changes in the outer domain for all iterations with varying insolation. We can see that the domain continually gets warmer in Figure B.5, up to more than 5 C by 110 kya, which strongly prevents ice growth.

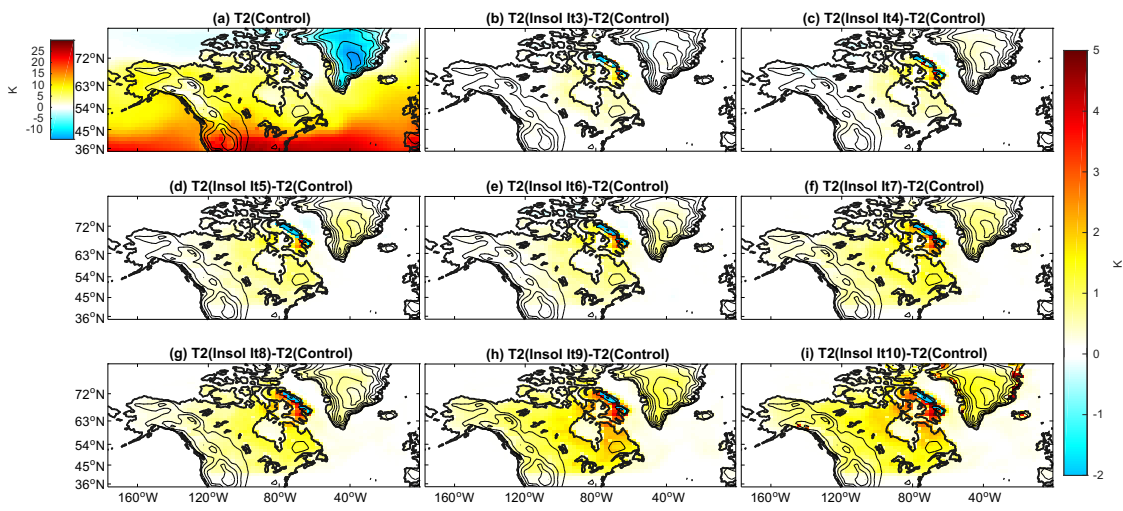


Figure B.5: 2m Temperature During the Summer (a) The control simulation summer temperature with present day ice cover and 115 kya insolation (b)-(i) Temperature differences during the summer between control and subsequent iterations. With an asynchronous coupling time step of 500 years. Insolation varies from 114 kya in (b) to 111 kya in (i).

B.3 ADDITIONAL ASYNCHRONOUS COUPLED ATMOSPHERE-ICE SIMULATIONS

In light of the negative feedback on glacial inception arising from the ice sheet-atmosphere interaction, we attempted the asynchronous iterations with different mass balances and coupling schemes. We thought that slightly different glacial flow and resulting ice sheet shape might induce different atmospheric responses, as noted by Chen and Lin (2005). Our results are present below, but do not differ significantly from our main set of simulations, the ice still stagnates on Baffin Island.

B.3.1 ICE FLOW CHARACTERIZED BY THE REGRESSION OF MASS BALANCE ON ELEVATION

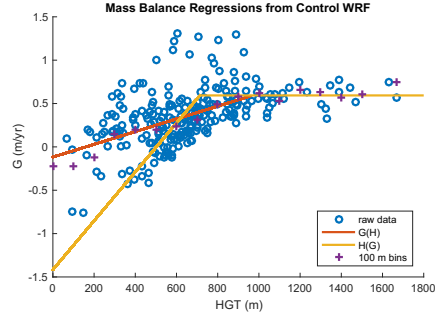


Figure B.6: The scattered mass balance of the control WRF simulation with the binned mass balances denoted by purple "+", which were used in the main set of simulations. Also pictured are the regressed mass balances, $G(H)$ and $H(G)$, used in the below iterations. Note the plateau at high elevations for the inclusion of the elevation desert effect.

Though we did a asynchronous coupling with binned mass balances for the ice model, we also attempted to fit regression lines to the mass balance from WRF. We regressed the mass balance on elevation ($G(H)$) and also the elevation on mass balance ($H(G)$), leading to very different slopes in Fig. B.6. The slope for $H(G)$ is actually very similar to our slope from Birch et al. (2017), but neither slope seems to fit the data well, which is why we found binning the data for effective at capturing the mass balance.

Also fitting a line, requires us to prescribe the elevation desert effect by averaging the mass balances above 1500 m and using that average value for all elevations above that height. In this way,

we create the pictured plateau. We perform asynchronously coupled simulations with $G(H)$ in Figure B.7 and $H(G)$ in Figure B.8, but get the same negative feedback found in our main set of simulations.

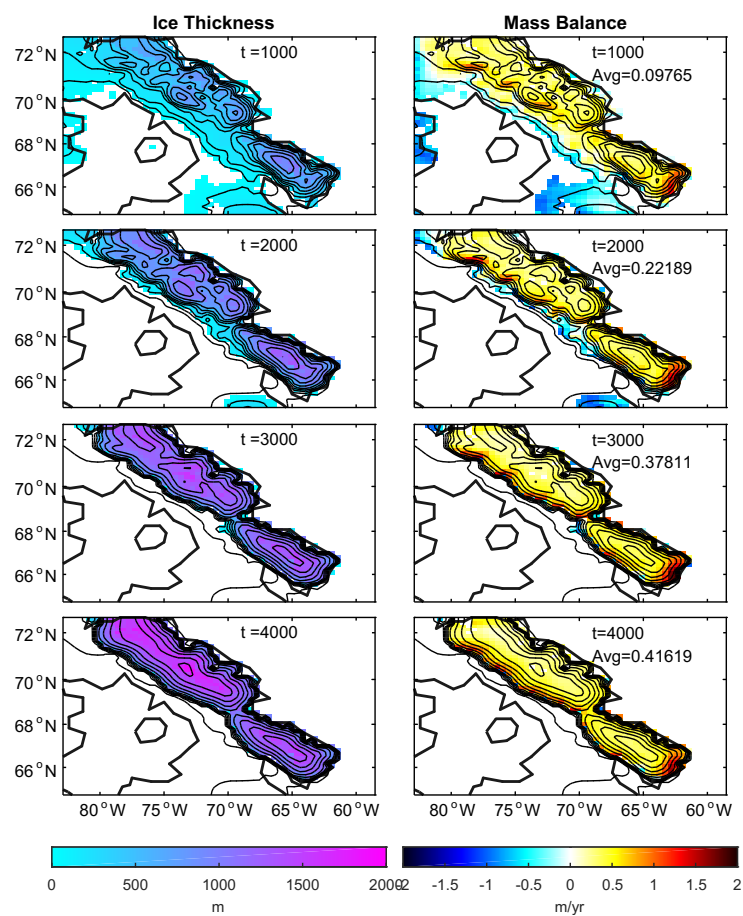


Figure B.7: Ice growth due to WRF-ice model iterations with the mass balance based on $G(H)$. Left column is the ice thickness from the ice sheet model. Right column is the mass balance from WRF.

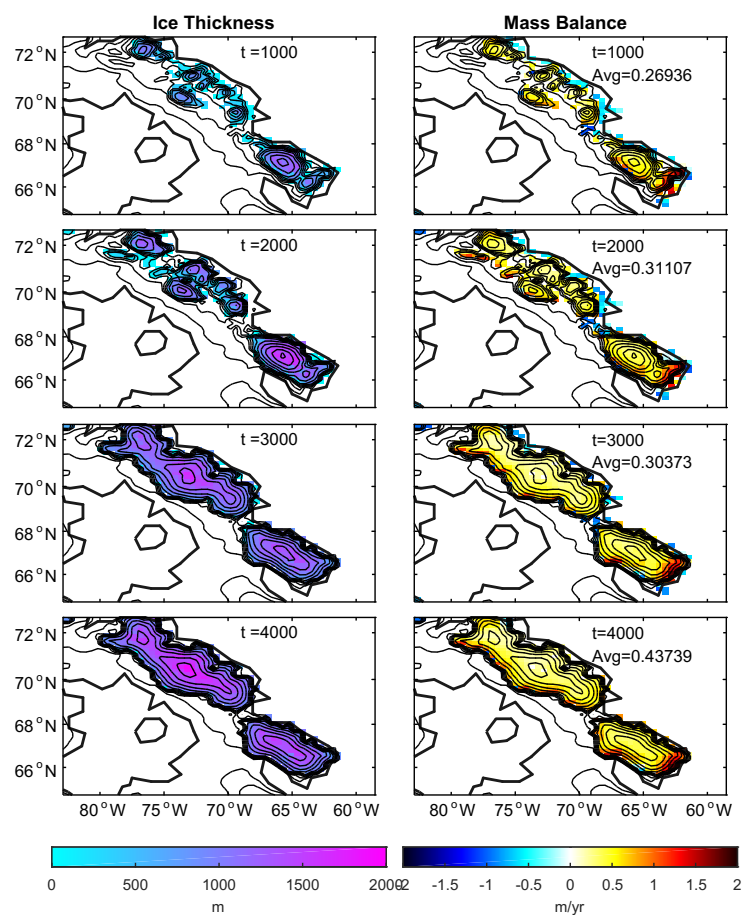


Figure B.8: Ice growth due to WRF-ice model iterations with the mass balance based on $H(G)$. Left column is the ice thickness from the ice sheet model. Right column is the mass balance from WRF.

B.3.2 ICE MODEL ITERATIONS WITH RUNGE KUTTA 2 TIME STEP

Assuming the mass balance from WRF is maintained during the ice model integration is likely in error. This assumes an almost linear response, where the same mass balance characterizes the growth for 500 years. In other words, we used an Euler approach in our main set of simulations, but a more nuanced time step, such as Runge Kutta 2 (RK2), may capture mass balances more representative of the climate. We consider the mass balance from the second iteration of WRF to be a half step in time. We then use this mass balance to run the ice model for a full time step using the ice initialized based on the data from the NASA Operation IceBridge Mission (Kurtz and Farrell, 2011). A full time step of RK2 uses the mass balance from iteration 2 (considered a half step here) and the ice from iteration 1. As we go forward in time, subsequent full time steps use mass balances from half time steps and ice cover from the previous full step. At the end of 5 kyrs, we still find very little growth. We consistently find a strong negative feedback on the ice sheet from anticyclonic response in the atmosphere. We note that it is possible that an asynchronous modeling scheme may not capture the mass balance and ice evolution. However, a synchronous approach by Gregory et al. (2012) simulated similar ice growth as in their asynchronous approach .

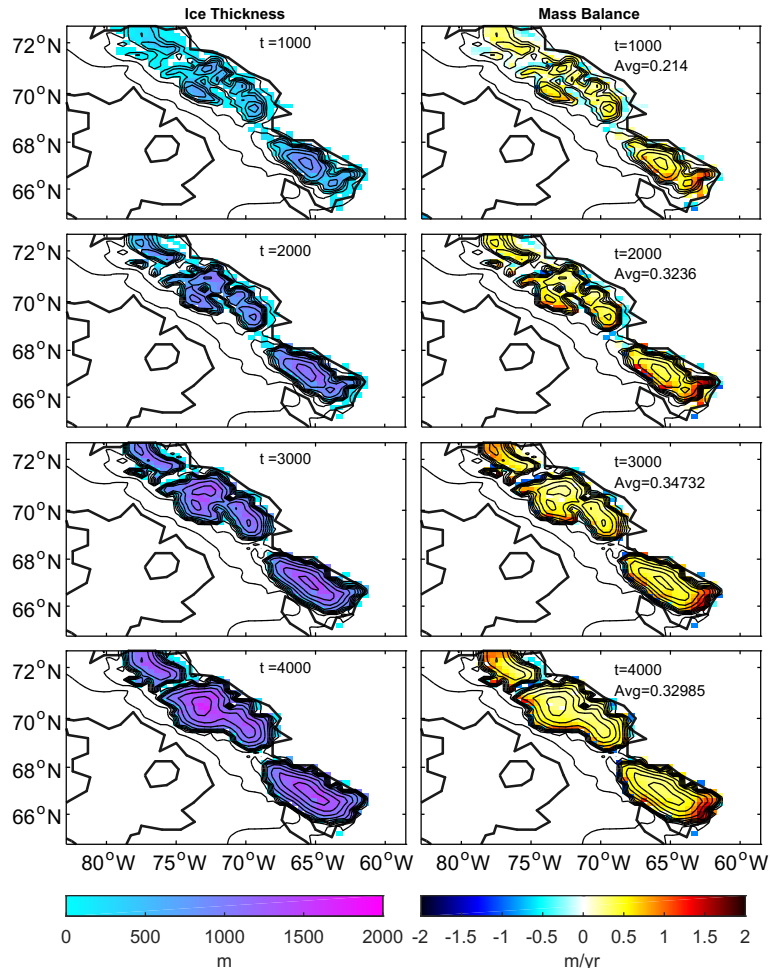


Figure B.9: Ice growth due to WRF-ice model iterations with RK2 time step. Left column is the ice thickness from the ice sheet model. Right column is the mass balance from WRF.

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