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The Greenland Ice Sheet in a Changing Climate Snow, Water, Ice and Permafrost in the Arctic (SWIPA) 2009

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D. Dahl-Jensen, J. Bamber, C.E. Bøggild, E. Buch, J.H. Christensen, K. Dethloff, M. Fahnestock, S. Marshall, M. Rosing, K. Steffen, R. Thomas, M. Truffer, M. van den Broeke and C.J. van der Veen

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Preface

The *Greenland Ice Sheet in a Changing Climate* report details the results of work conducted to date on the Greenland Ice Sheet (GRIS) component of the Arctic Council's Cryosphere project – *Snow, Water, Ice and Permafrost in the Arctic (SWIPA)*. As such, this report is the first in a series of reports presenting the results of the SWIPA project.

The SWIPA project was established by the Arctic Council in April 2008 as a follow-up to the 2005 Arctic Climate Impact Assessment (ACIA). Its goal is to assess current scientific information on changes in the Arctic cryosphere, including the impacts due to changes in the cryosphere that have potentially far reaching implications for both the Arctic and the Earth as a whole.

The SWIPA project is coordinated by the Arctic Council's Arctic Monitoring and Assessment Programme (AMAP) in cooperation with the International Arctic Science Committee (IASC), International Arctic Social Sciences Association (IASSA), International Polar Year, and WCRP-CliC Program (for further information see www.amap.no/swipa). The Greenland Ice Sheet component of the SWIPA project is led by Denmark.

This report presents the scientific and technical information compiled under the first phase of the Greenland Ice Sheet component. Future SWIPA reports will include an update of the information concerning the Greeenland Ice Sheet, in particular the sections dealing with potential impacts on biological systems and human populations. These potential impacts can only be fully addressed in relation to the combined effects of changes in all components of the cryosphere – and thus form part of the SWIPA integration assessment.

The *Greenland Ice Sheet in a Changing Climate* report has been subjected to an extensive international peer review process to ensure the scientific validity of the information presented. The results presented in this report are also summarized in a popular-style report: *Summary of the Greenland Ice Sheet in a Changing Climate* that presents the main findings of the Greenland Ice Sheet assessment in a format more comprehensible to a broader audience.

The full scientific report is prepared in English only. The summary report is also available in Chinese, Danish, French, Greenlandic and Russian translations. The SWIPA-GRIS reports, together with other SWIPA products, will be presented at the UNFCCC COP15 meeting in Copenhagen in December 2009. The results of the SWIPA project will also be provided to the UNFCCC Intergovernmental Panel on Climate Change (IPCC) for use in future IPCC assessments.

AMAP would like to thank all of the scientific experts that have contributed to this assessment. A list of the main authors and contributors is included in the colophon to this report. Thanks are also due to Denmark for leading this component of the SWIPA work, and in particular to Dorthe Dahl-Jensen for leading the GRIS component and Henning Thing for his work on the GRIS component coordination.

The support of the Arctic countries is vital to the success of SWIPA and AMAP work in general. Furthermore, the SWIPA-GRIS project could not have been conducted without the additional financial support received from Canada, Denmark, Norway, and the Nordic Council of Ministers.

1. Introduction

1.1. Evolution of the Greenland Ice Sheet

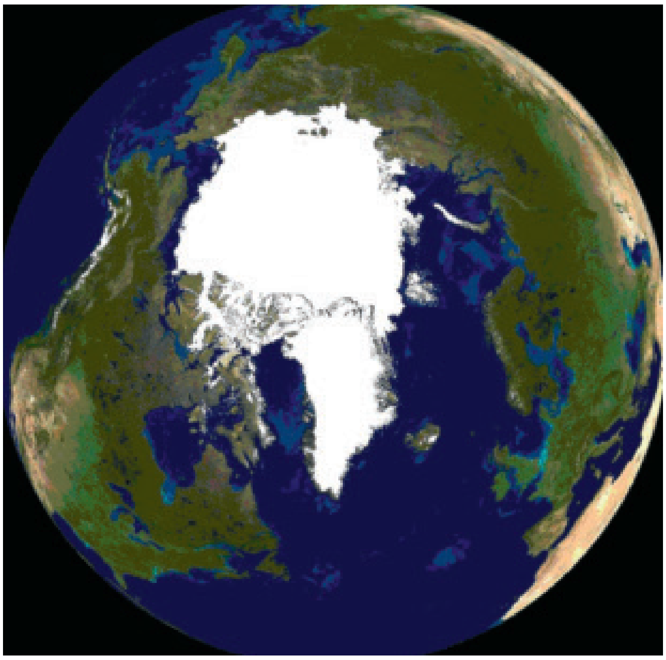


Figure 1.1 *The high latitudes of the Northern Hemisphere are covered with ice and snow. The high reflectance (albedo) of the white areas reduces the absorbance of incoming solar energy. As the white areas diminish in a warming climate more solar energy will be absorbed by the ice-free ocean and land surfaces. Source: Earth View image.*

The Greenland Ice Sheet (Figure 1.1) is the largest body of ice in the Northern Hemisphere and, globally, is dwarfed only by the Antarctic Ice Sheet. If the entire Greenland Ice Sheet were to melt, the volume of water released from the 2.93 million km^3 of ice that it contains would cause global sea level to rise by about 7 m (Table 1.1).

The aim of this assessment was to compile recently gained scientific results, focusing specifically on the impacts of recent changes in the Greenland Ice Sheet. Although many reports on climate change have addressed impacts resulting from the evolution of the Greenland Ice Sheet, none provide a thorough and up-to-date assessment of the importance of the Greenland Ice Sheet in a changing climate. The importance of the ice sheet as a component of the Arctic cryosphere and the enormous amount of new observations and data provide the background for a better understanding of the evolution of the Greenland Ice Sheet and its potential impacts.

The surface topography of the Greenland Ice Sheet determines atmospheric weather patterns over the ice sheet itself and across a large surrounding region. Plus, the white surface of the snow and ice reflects high energy radiation from the sun. As a result, changes in the snow- and ice-covered area and in the properties of the ice sheet surface will alter the albedo of the area. Regionally, the sum of these changes will have direct and indirect socio-economic impacts on the people of Greenland whose traditional way of life has been closely linked to nature. Fishing, travel, access to natural resources such as oil and minerals, living conditions, and freshwater supply are all influenced by the changing Greenland Ice Sheet.

The high latitudes of the Northern Hemisphere are the regions that have experienced the strongest changes in climate since the 1951-1980 reference period (Figure 1.2). While the global annual temperature anomaly for the period 2005-2007 in relation to the 1951-1980 reference period is 0.7 °C the anomaly over Greenland is 1.5 °C,

Table 1.1 *Total volume of the ice bodies on Earth and the equivalent global sea level change, if the ice bodies were to melt completely.*

	Volume, 10^6 km^3	Equivalent sea level rise, m	Source
Antarctic Ice Sheet	25.4	57	Lythe et al. (2001)
Greenland Ice Sheet	2.93	7.5	Bamber et al. (2001)
All remaining glaciers and ice caps	0.051 – 0.133	0.15 – 0.37	Lemke et al. (2007)

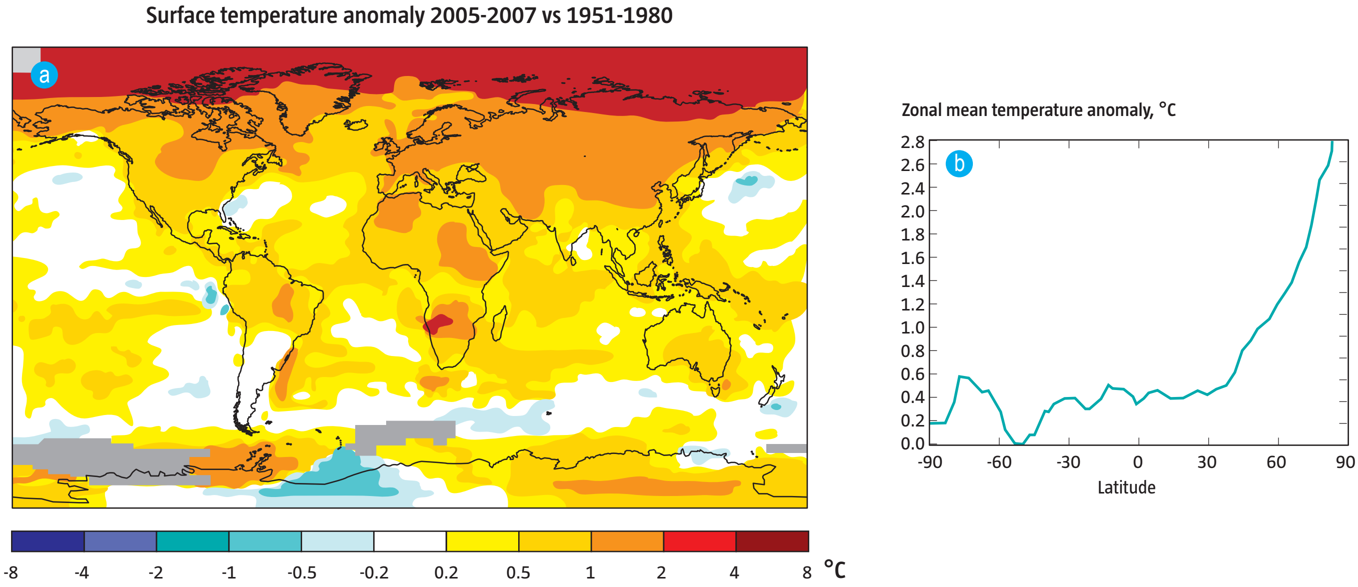


Figure 1.2 *Mean surface temperature changes between the period 2005-2007 and the reference period 1951-1980: (a) map averaged over 1200 km, (b) zonal means of the data shown on the main map. Source: data.giss.nasa.gov/gistemp.*

which is twice the global value. Average surface temperature changes observed in the Arctic are larger than those observed in the Antarctic.

The trend of strongly enhanced changes in surface temperature in the high northern latitudes is not only an issue in terms of the changes observed to date but is also a cause for concern in terms of the predicted increase in future temperatures. In the SRESA1B scenario reported by the UN Intergovernmental Panel on Climate Change (IPCC) in its Fourth Assessment Report, annual temperatures over the Arctic (i.e., all areas north of 60° N) were projected to be 2.8 to 7.8 °C higher in 2100 AD than they were during the 1951-1980 reference period, with winters warming at a greater rate than summers (Meehl et al., 2007). The predicted temperature increase is serious and would be the largest that the Greenland Ice Sheet will have experienced since the last interglacial period, the Eemian, 130 000 to 115 000 BP (North Greenland Ice Core Project members, 2004).

1.2. Past changes of the Greenland Ice Sheet

The global climate has changed enormously over time; there have been periods with glaciations and periods with no ice on Earth. While the Northern Hemisphere experienced ephemeral glaciations from 38 000 000 to 4 000 000 BP, the onset of extensive glaciations did not occur until 3 000 000 BP (Lunt et al., 2008).

Over the last three million years, the Greenland Ice Sheet is believed to have waxed and waned in accordance with glacial and interglacial periods (Weidick, 1993; Weidick and Bennike, 2007) (Figure 1.3). Glacial deposits (i.e., moraines) 20 to 50 km from the present shoreline of Greenland can be used to reconstruct the Greenland Ice Sheet in earlier climatic periods. While the maximum extent of the ice sheet during glacial periods can be observed through studies of moraines and marine sediment cores, the degree to which the Greenland Ice Sheet retreated in earlier warmer periods in not clear. During the last glacial maximum, 25 000 to 14 000 BP, annual temperatures over Greenland are believed to have been 25 °C colder than at present (Cuffey et al., 1994; Cuffey and Clow, 1997; Dahl-Jensen et al., 1998), sea level to have been 120 m lower than at present (Lambeck et al., 2002), and the volume of the Greenland Ice Sheet to have been 140% higher than at present (Huybrechts, 2002; Lambeck et al., 2002).

Throughout the last glacial period, the climate was very unstable with 25 rapid climate changes, termed 'interstadials' or 'Dansgaard-Oeschger events' (DO; Figure 1.3). The area and volume of the Greenland Ice Sheet increased to 40% more than the present area and volume. The north Atlantic region and the Greenland Ice Sheet experienced rapid temperature increases of 10 to 15 °C occurring over a few decades (North Greenland Ice Core Project members, 2004; Landais et al., 2005; Steffensen et al., 2008) together with sea level increases of 5 to 20 m (Siddall et al., 2003). The rapid changes are believed to have been connected to enormous surges of ice into the ocean, especially from the Laurentide Ice Sheet over North America. After the abrupt warmings during the D-O events the surface temperatures gradually cooled over 1000 to 5000 years, before the next rapid warming occurred. The contribution of

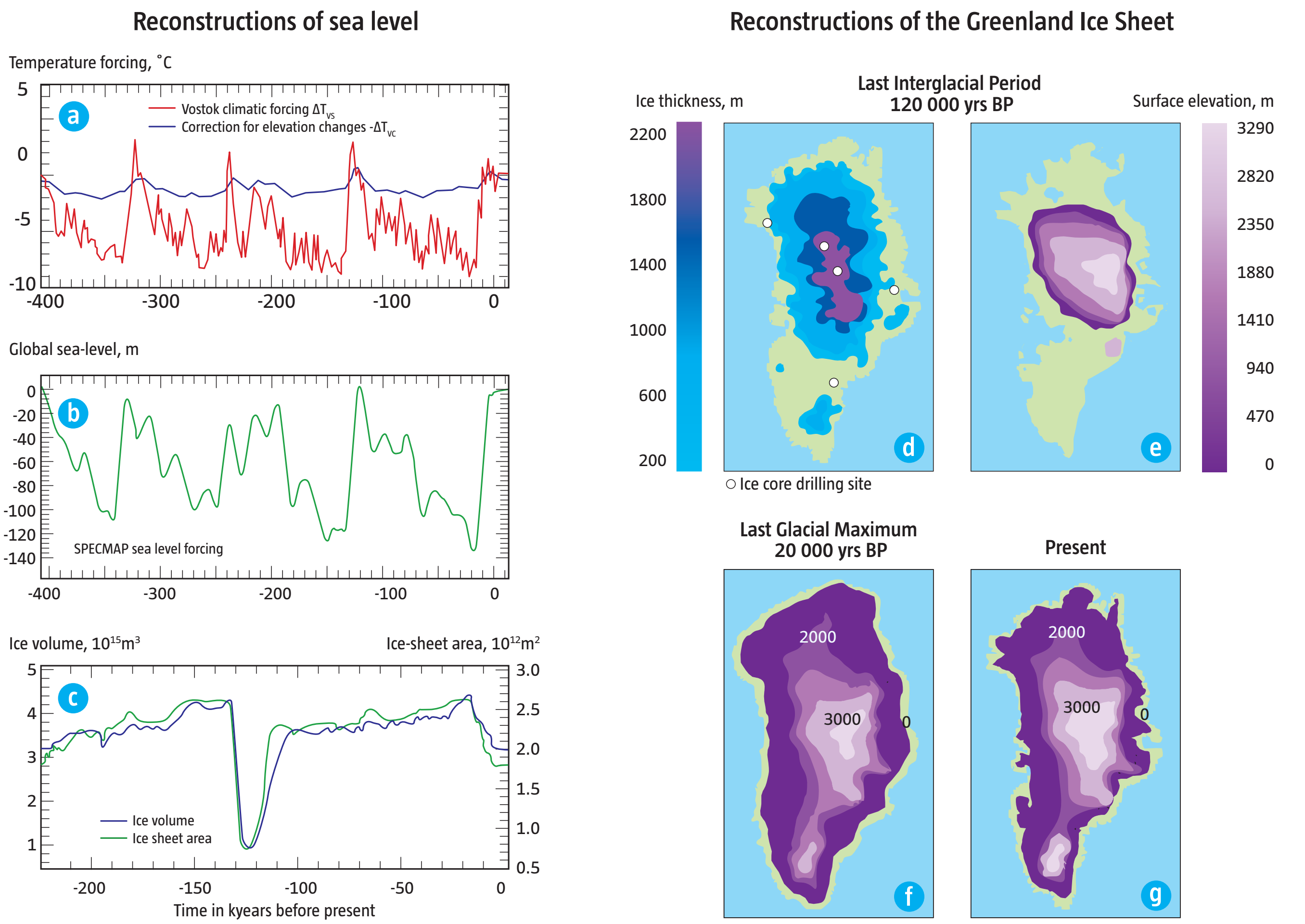


Figure 1.3 *The volume and area of the Greenland Ice Sheet have changed as the climate has evolved through glacial and interglacial periods. (a–c) Reconstructions of past temperatures, global sea level, and volume and area of the Greenland Ice Sheet (Huybrechts, 2002). The reconstructed temperatures, based on the Vostok ice core data (a), are compared with reconstructed sea level changes (b) over the last 400 000 years. The volume and area of the Greenland Ice Sheet for the last 200 000 years are shown in (c). Snapshots of the modeled Greenland Ice Sheet from the Last Interglacial Period, 130 000 to 115 000 BP, from (d) Otto-Bliesner et al. (2006) and (e) Cuffey and Marshall (2000) show the very reduced volume during the previous warm period. During the cold Last Glacial Maximum, 20 000 BP, the Greenland Ice Sheet volume increased and model simulations from (f) Huybrechts (2002) estimate that the added volume of the Greenland Ice Sheet was equivalent to a reduced average global sea level lowering of 2.9 m. For comparison, the model simulation of the present Greenland Ice Sheet is shown in (g) Huybrechts (2002).*

the Greenland Ice Sheet to the sea level rises that occurred during these rapid warming events is not known.

After the glacial period, the climate warmed into the present interglacial period, the Holocene, and the coastal regions around Greenland experienced isostatic uplift of up to 100 m in response to the retreating ice (Letréguilly et al., 1991; Weidick, 1993; Huybrechts, 2002).

In the present Holocene climate (the last 12 000 years) temperature change has been moderate compared to changes experienced during the last glacial period. After the last glacial period, during the period named the 'climatic optimum' 8000 to 5000 years ago, Greenland temperatures were 2 °C warmer than in the 1951-1980 reference period (Dahl-Jensen et al., 1998). Modeled volume changes in the Greenland Ice Sheet differed significantly during the climatic optimum depending on the extent of the ice sheet during the last glacial maximum, the mass balance applied, and the evolution of the temperature in the ice (Letréguilly et al., 1991; Huybrechts, 1996; Marshall and Cuffey, 2000; Clarke and Marshall, 2002). Investigations of the ice sheet in the region around the Jakobshavn Isbræ glacier on the west coast of Greenland show that 8000 to 5000 years ago the margin had retreated farther east than at present (Weidick et al., 1990; Weidick and Bennike, 2007).

The previous interglacial period, the Eemian 130 000 to 115 000 BP, is a very interesting period. The average air temperature is believed to have been 5 °C warmer than in the 1951-1980 reference period and to have remained so for several thousand years (North Greenland Ice Core Project members, 2004). The reaction of the Greenland Ice Sheet during the Eemian can be seen as an analogue to the evolution of the ice sheet in future warming scenarios. During this warm interglacial period, sea level was 4 to 6 m higher than at present (Stirling et al., 1998; Lambeck et al., 2002).

Model reconstructions of the ice sheet during the Eemian show very reduced ice volumes, especially in southern Greenland (Letréguilly et al., 1991; Cuffey and Marshall, 2000; Otto-Bliesner et al., 2006), predicting the sea level change to be 1 to 5 m based on the change in volume of the Greenland Ice Sheet. Findings of Eemian ice in several Greenland ice cores and the presence of older ice at the base of the southern Greenland ice core drill site, DYE 3 (North Greenland Ice Core Project members, 2004; Willerslev et al., 2007) support volume changes equivalent to a 1 to 3 m rise in sea level.

1.3. Evolution of the Greenland Ice Sheet in the future

Results from modeling the past evolution of the Greenland Ice Sheet indicate the need for improvements in order to model the evolution of the ice sheet in a changing climate. Not enough is known about many fundamental parameters, such as ice sheet mass balance, due to the vast extent of the ice sheet. In addition, dynamic processes, such as basal sliding and ice stream dynamics, are poorly understood.

This inadequate knowledge of processes is affecting the ability to predict the evolution of ice sheets and led the IPCC to conclude that:

> *Taken together, the ice sheets of Greenland and Antarctica are very likely shrinking, with Greenland contributing about 0.2 ± 0.1 mm yr^{-1} and Antarctica contributing 0.2 ± 0.35 mm yr^{-1} to sea level rise over the period 1993 to 2003. There is evidence of accelerated loss through 2005. Thickening of high-altitude, cold regions of Greenland and East Antarctica, perhaps from coastal regions of Greenland and West Antarctica in response to increased ice outflow and increased Greenland surface melting.* (Lemke et al., 2007:376).

Since the latest IPCC report (2007) was published, several other compilations have raised concern about the evolution of the big ice masses in response to global warming (Rignot et al., 2008; Steffen et al., 2008).

There is consensus from observations that the mass loss from the Greenland Ice Sheet has accelerated since 2000. While the contribution from the Greenland Ice Sheet is currently 10% to 20% of the observed global sea level rise of 3 mm/y, the Greenland Ice Sheet is believed to be capable of reacting more strongly to the warming over the next 100 years with a mass loss that could increase sea level by a significant fraction of 1 m (Rignot et al., 2008; Steffen et al., 2008).

2. The Greenland Ice Sheet today

2.1. Overview of the Greenland Ice Sheet

2.1.1. Arctic climate and large-scale synoptic observations

Greenland is the largest island in the world with a surface area of 2.2 million km², stretching 2600 km from 59.8° N to 83.6° N, where – for several months of the year – it is either polar night or continuous daylight. Over 80% of Greenland is covered by a dome of inland ice (Figure 2.1) rising along an average gradient of about 1% from sea level in various regions to over 3300 m along the central spine.

Greenland plays a pivotal role in determining the climate of the Northern Hemisphere because of its size, location, elevation gradient and mass of freshwater stored in the ice sheet. During summer, Greenland is ideal for monitoring changes in meridional energy transport into the Arctic because it is located in the predominant direction of cyclone flow. Greenland's extreme North Atlantic location and the ice-filled ocean that surrounds it are the principal factors influencing the climate. The northern branch of the Gulf Stream, known as the North Atlantic Current, runs northward along the Norwegian coast into the Arctic Ocean, mixes with cold polar water and returns southward along the east coast of Greenland. Nearly all the water that leaves the Arctic Ocean passes through Fram Strait (Figure 2.1) via the East Greenland Current, eventually wrapping around Cape Farewell and continuing northward along the west coast of Greenland for several hundred kilometers. A similar, counterclockwise gyre is active along the coast of West Greenland with warm water moving northward until it meets colder polar water flowing through Kennedy Channel. The resultant cold current flows southward along northeastern Canada.

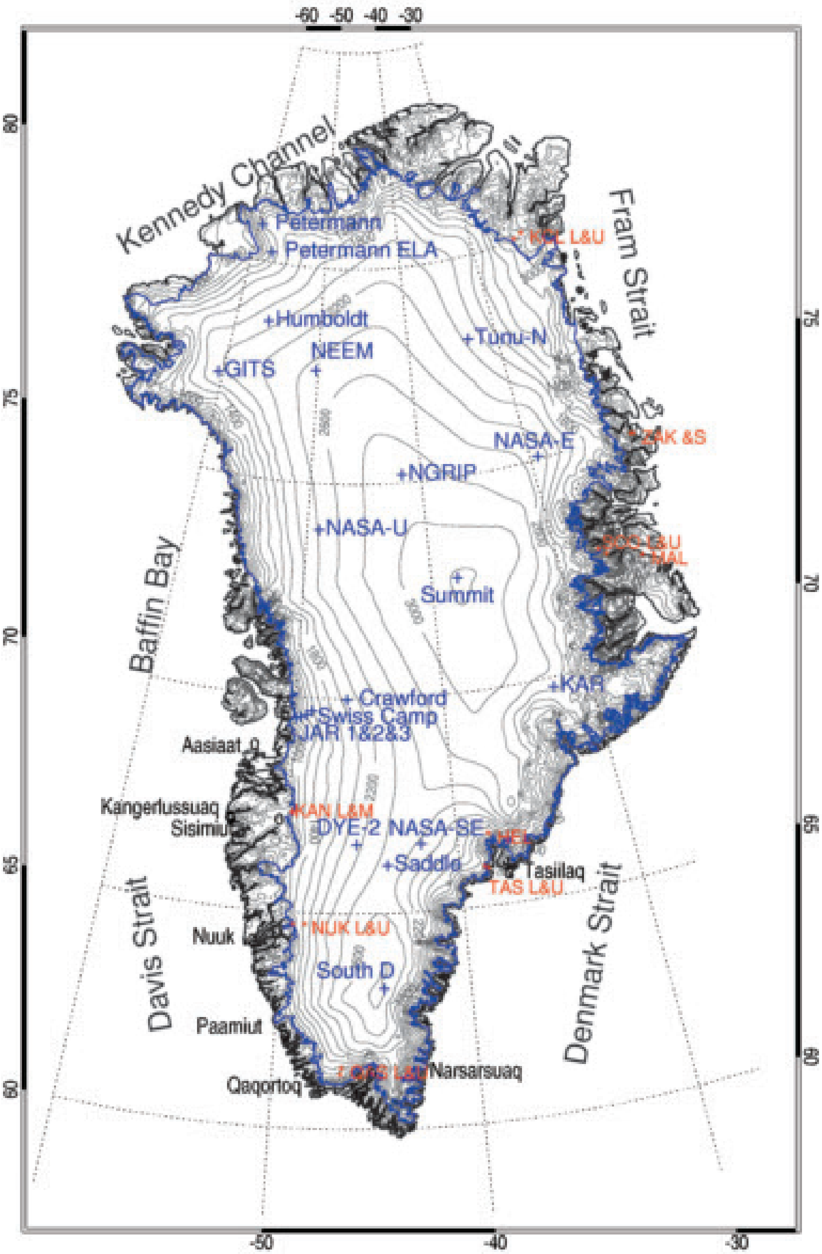


Figure 2.1 *Greenland Ice Sheet elevation map and location of WMO climate stations along the coast and Greenland Climate Network (GC-Net) automatic weather stations [blue] in the ice sheet interior and the climate stations from the Geological Survey of Denmark and Greenland [red]. Source: Konrad Steffen, CIRES, University of Colorado.*

Wind in the lower troposphere is forced to flow along the coasts of Greenland because of the height and size of the ice sheet. As a result, Greenland is an important participant in the exchange of air mass between the southern and northern latitudes of the Northern Hemisphere. Northerly and southerly air flow is about evenly distributed during summer. However, the predominant flow during winter is northerly due to high pressure in the colder regions to the west or northwest. Airflow in the free atmosphere at the 500 hPa level plays a key role in the Greenland climate because it governs the North Atlantic storm track. Generally, airflow at 500 hPa is from the southwest during winter and from the west during summer.

A typical North Atlantic cyclone develops as a wave in the polar front and propagates along the front. Therefore, winter cyclones generally travel along the east coast of the United States tracking along the edge of the Gulf Stream heading northeast. The cyclones often pass south of Greenland continuing toward Iceland and into the Norwegian Sea. However, they can also track more to the north through Davis Strait and into Baffin Bay. Occasionally, a cyclone will split near the southern tip of Greenland producing separate lows that track along both coasts. During summer, lows tend to be less intense but are frequently more northerly and, therefore, often influence summer conditions in western Greenland.

Passing cyclones are generally accompanied by strong winds. However, during periods with no cyclone activity, the wind regimes are determined by local conditions, which usually relate to katabatic flow on the Greenland Ice Sheet, sea breezes during summer or land breezes during winter near the coasts. Whether the high velocity katabatic winds on the ice sheet propagate down the fjords to the coastal areas largely depends on the temperature of the air mass as it reaches the head of the fjords. If the katabatic air mass is warmer than the air in the fjord due to adiabatic warming it will not be able to replace the air near the top of the fjord where it will be experienced as a warm 'foehn wind'. As a result, it will ride above the colder and denser surface air mass. If the katabatic flow is colder than the air in the fjord, it will replace the local air mass and probably travel all the way to the coast where it will be experienced as an icy downhill wind.

Climatic conditions in the fjords in the absence of cyclonic or katabatic influences are usually characterized by sea breezes during summer and land breezes during winter due to ocean-land temperature differences. The winds associated with passing cyclones are, generally, very strong and heavily influenced by local topography and direction of the wind relative to the coast. Winds blowing directly toward the coast will often lead to precipitation due to orographic lifting. This scenario can often lead to very strong barrier winds, which blow clockwise relative to the land mass. Another special feature of the Greenland wind regime is that of very rapid change from calm to gale force conditions. These situations often develop as a result of cold Canadian air reaching the eastern coast of Greenland via the ice sheet behind a cyclone tracking to the northeast. The ice sheet topography determines the direction of the cold outflow from the ice sheet, focusing the wind into the low-lying coastal regions. The Greenlandic term for such an event is *pitoraq* (*piteraq* in East Greenland) which is closely related to the word for 'being attacked'. Piteraqs are most common during autumn and winter. Wind speeds typically reach 50 to 80 m/s. On 6 February 1970, the community Tasiilaq, also known as Ammassalik and the most populous community in East Greenland, was hit by a very strong piteraq causing severe damage. Since the beginning of the 1970s, special piteraq warnings have been issued by the Danish Meteorological Institute.

Surface-air temperature in Greenland during summer is dominated by radiation effects as the sunlight returns to the Arctic. As a result, the mean temperature in July in the northernmost part of the country is only about 2 °C colder than in the southernmost part. However, temperatures at the coast are strongly influenced by the ocean and sea-ice variability; inland temperatures in ice-free regions can, therefore, be 5 °C warmer than at the coast. The proximity of the ice sheet does not seem to cool neighboring areas because the air flow off the ice during summer is usually warmer than the low lying local air mass due to adiabatic warming.

The latitudinal temperature gradient is much greater during winter than during summer with average sea level February air temperatures in the north of -36 °C and -4 °C in the south. Variability is also much greater without the moderating influences of the open ocean in the north and the melting ice sheet. Katabatic storms often raise winter temperatures above freezing via the combined effects of adiabatic warming and mixing of the inversion layer, particularly in the southern regions, where winter temperatures in the fjords can reach 10 °C (Cappelen, 2003).

Analysis of the long-term temperature record (1840 to 2001) from coastal stations in Greenland (Vinther et al., 2006) puts the more recent observations from satellites and weather stations on the Greenland Ice Sheet into a broader context (Box, 2002). The warmest decades in the long-term Greenland temperature record (1840 to 2000) are the 1930s and the 1940s. Generally, the west coast experienced warming from 1873 to 1930 followed by cooling until 1988. There has been a warming trend since 1985 of 2-4 °C in West Greenland, primarily driven by winter temperature anomalies. The current warm period, beginning in 1988, is not unprecedented because the record warm decades occurred during the 1930s and 1940s.

The recent warmth occurred in spite of a persistently positive North Atlantic Oscillation (NAO) since around 1970. The NAO has a significant influence on the coastal temperature record particularly in the western and southern regions of Greenland during winter. The spectra of the temperature records display peak power at 3.7 years which Box (2002) attributed to the NAO, which has a spectral peak at ~4.0 years. The positive trend in coastal temperatures in southwest Greenland appears to be responsible for the ice-sheet thinning observed in the region (Ohmura et al., 1999).

Volcanism has a significant cooling effect on temperatures in southern and western Greenland with a peak lag of 5 to 10 months following major eruptions (Box, 2002). The warming trend in Greenland since the early 1990s, in spite of major volcanic eruptions in 1982 and 1991, stands in contrast to the warmth of the 1930s and 1940s during a period of anomalously low volcanic activity (Overland et al., 2004).

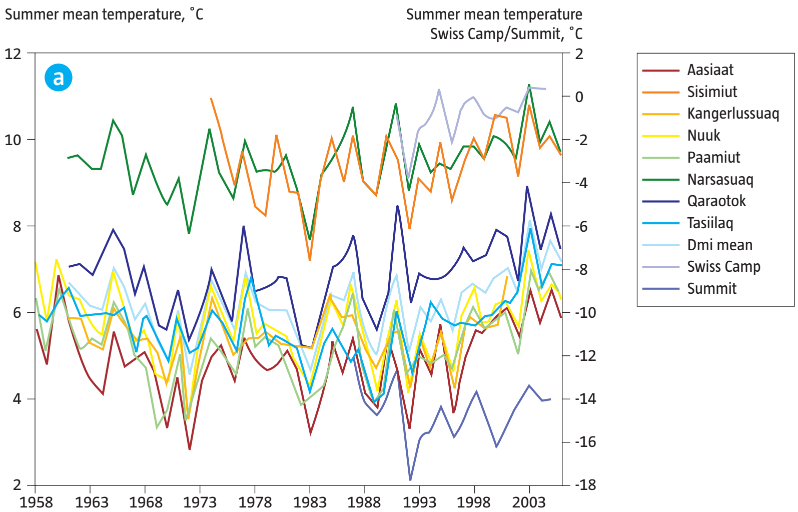
Meteorological station records and regional climate model output were combined to develop a continuous 168-year (1840 to 2007) spatial reconstruction of monthly, seasonal, and annual mean Greenland Ice Sheet near-surface air temperatures (Box et al., 2009). Box and co-workers found that volcanic cooling episodes were concentrated in winter and along the western ice sheet slope, that inter-decadal warming trends coincided with an absence of major volcanic eruptions, and that 2003 was the only year in the 168-year series with a warm anomaly that exceeded three standard deviations from the 1951-1980 base period. The magnitude of the annual whole ice sheet 1919-1932 warming trend is 33% greater than the 1994-2007 warming. Box et al. (2009) also found that the recent warming was stronger along western Greenland in autumn and southern Greenland in winter. Spring trends marked the 1920s warming onset while autumn leads the 1994-2007 warming.

The climate on the Greenland Ice Sheet has been studied intensively in an effort to quantify the surface mass balance and to estimate its contribution to global sea level. However, *in situ* observations are limited due to the large area and remoteness of the ice sheet. The notable exception is the climate record from the Greenland Climate Network (GC-Net) of automatic weather stations, operating since 1995 and distributed across the ice sheet (Figure 2.1). Scientists are, therefore, forced to rely on remotely sensed observations and estimations from downscaled global circulation models or high-resolution regional climate models. However, systematic biases in the models relative to the *in situ* record illustrate that the climate of the ice sheet is more or less decoupled from the coastal regions where most of the direct observations that feed the models are made. Nevertheless, a consensus picture of change due to a warming climate since 1995 is emerging.

2.1.2. Climate observations

The Danish Meteorological Institute (DMI) data are the most comprehensive meteorological records available for Greenland's coastal region and have been homogenized, most notably the original observations have been checked and the data compared with time series of related climatic elements for the same stations, with the specific purpose of studying long-term climatological trends (Cappelen et al., 2001).

Monthly air temperature records from eight DMI synoptic stations at the coast of Greenland (located mainly in the southwest but including Tasiilaq in the southeast) show a pronounced warming since the early 1990s. The warming follows an overall regional cooling trend between 1958 and 1992, concentrated in the winters of the 1960s to 1980s (Hanna and Cappelen, 2003). Summer trends of the 7-station DMI average (i.e., excluding Kangerlussuaq because of its relatively short record) indicate a significant warming for 1961 to 2006 of 0.9 °C.



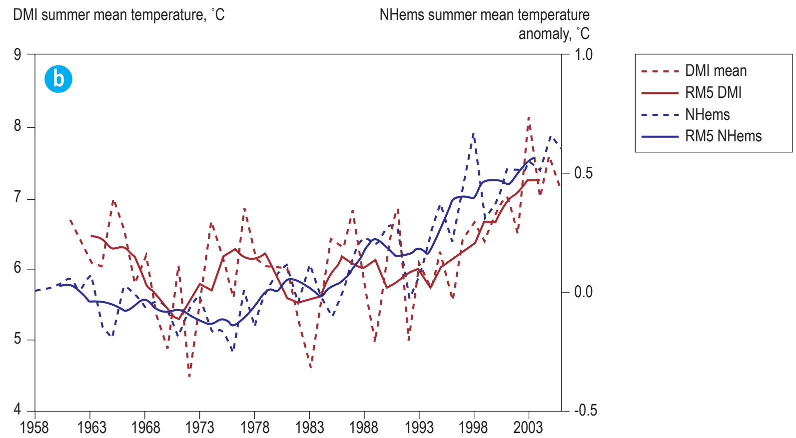


Figure 2.2 *(a) Summer (June-July-August) mean temperature 1958 – 2006 at Greenland climate stations (for locations see Figure 2.1). (b) Greenland summer mean temperature from the mean of seven DMI stations (see Figure 2.1, except Kangerlussuaq) compared with Northern Hemisphere summer mean temperature, with five-year running means of both series. Note some early years in Greenland / Northern Hemisphere antiphase (e.g., 1965, 1971, 1983), as highlighted by the disparate running means, but the post-Pinatubo (1992 onward) period shows better agreement in terms of interannual variability, correlation, and strong upward trends. Source: Hanna et al. (2008).*

For the DMI station average, 2003 had the warmest summer (i.e., June to August) on record, with a mean temperature of 8.1 °C at 3.3 standard deviations above the most recent climatological 'normal' period (1971-2000) mean. The second warmest summer occurred in 2005 (7.6 °C) at 2.5 standard deviations above the 1971-2000 mean. 2005 was more than 0.5 °C warmer than the third warmest summer, 2006, (7.07 °C) which was closely followed by 2001 (7.02 °C), 1965 (7.00 °C), and 2004 (6.97 °C) (Hanna et al., 2008).

The GC-Net automatic weather station 'Swiss Camp' (1170 m above sea level in 1991 and since then decreasing 0.32 m annually (Stober and Hepperle, 2006; Figure 2.1) record (Steffen and Box, 2001) was used to gauge temperature changes on the western flank of the Greenland Ice Sheet, where extensive seasonal melt and relatively high run-off from this relatively low elevation zone contribute a large proportion of the total run-off (Hanna et al., 2005; Box et al., 2006). This record, by far the longest GC-Net series, spans 19 years (1991-2009), and its interannual variability is significantly correlated with that of the mean of the seven DMI coastal stations (de-trended series r = 0.65, $p < 0.01$; Figure 2.2).

Similar to the positive (7-station mean) DMI temperature trend, Swiss Camp summer mean temperatures increased significantly by 2.4 °C from 1991-2008, or 2.2 °C for the period 1993-2008, excluding the global cooling effect of the Mount Pinatubo eruption in 1992. The three summers of 2003, 2004, and 2005 were almost equally record warm (mean temperatures 0.3, 0.3, and 0.2 °C, respectively) at Swiss Camp, alongside 1995 (0.5 °C) and the record year 2007 (1.2 °C). The latter season has been previously noted for its relatively high modeled run-off compared with most other years during 1958-2003 (Hanna et al., 2005). The latest data from the Swiss Camp GC-Net record show a continued warming with another record mean summer temperature of 1.2 °C in 2007; 0.7 °C above the previous maximum in 1995.

A reprocessed and updated 1987-2005 near-surface air temperature series for Summit (3205 m above sea level) (Shuman et al., 2001) shows a slight overall -0.3 °C cooling in summer, in contrast to all the other Greenland temperature records (all from much lower elevations and generally around the ice sheet margin; Figure 2.2). This new Summit series is a re-analyzed composite, primarily of University of Wisconsin and ongoing GC-Net automatic weather station data, supported by Special Sensor Microwave / Imager (SSM/I) brightness temperature data (Shuman et al., 1995, 1996, 2001). There is a highly significant correlation between individual years' fluctuations in the de-trended DMI and the Summit series (Figure 2.2).

Three hypothesized possible causes for the cooling trend observed at Summit compared to the warming trend along the coast are: (1) continued relative suppression of more regional climatic change by thermal inertia of the huge central Greenland ice mass as noted for Arctic Ocean sea ice (Serreze and Francis, 2006); (2) a differential response of the high elevation zones of the Greenland Ice Sheet in accordance with the well-known lower tropospheric warming / higher atmospheric cooling response to increased greenhouse gases (Stott et al., 2006), which has been demonstrated specifically for Greenland in a recent analysis of radiosonde data spanning 1964-2006 (Box and Cohen, 2006); and / or (3) regional changes in wind, cloud cover, or radiation patterns over the Greenland Ice Sheet.

This observed pattern (coastal, i.e., marginal, warming combined with little change or slight cooling in the high interior) is opposite to the output of simulations from Atmosphere-Ocean General Circulation Models (AOGCMs). In all high-resolution temperature changes studied in the present century, Huybrechts et al., (2004) and Gregory and Huybrechts (2006) found that modeled summer warming is actually greatest over the interior of Greenland and least along the coast.

High-resolution (5 km × 5 km) Surface Air Temperature (SAT) data were bilinearly interpolated from the 0.5 ° resolution 40-year European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) and corrected for ECMWF terrain errors using empirically derived ice sheet surface lapse rates, as explained by Hanna et al. (2005). These SAT data provide corroborating evidence for anomalously (3 °C) high summer temperatures around the Greenland margins during recent warm summers with a noted high melt, which occurred most widely within the southern and western marginal ablation zones of the ice sheet.

The accumulation zone of the Greenland Ice Sheet (above 2000 m elevation) apparently had cold summer anomalies in 2003 and 2006. This is in contrast to warm anomalies in more outlying areas but in line with the Summit temperature results discussed above. Both Greenland and Northern Hemisphere summer temperatures exhibit strongly rising trends since the early 1990s, although the earliest part of this period (1992-1993) characterizes the general temperature recovery following the cooling after the 1991 Mount Pinatubo volcanic eruption (Robock and Mao, 1995).

Greenland Ice Sheet precipitation – downscaled from ECMWF operational analyses and re-analyses (Hanna et al., 2005) – follows a significantly increasing trend of 90.9 km^3/y, compared with a standard deviation of 69.7 km^3/y for 1958-2006. Additional precipitation, due to warmer air temperatures, mainly in the form of snow accumulation, therefore largely (i.e., ~80%) offsets rising Greenland run-off over the same period (Hanna et al., 2008).

Observations and models both indicate the occurrence of recent high snow accumulation events in the winter of 2004/2005, concentrated in western Greenland (Nghiem et al., 2007), and winter–spring 2002/2003 in southeastern Greenland (Krabill et al., 2004; Hanna et al., 2006). Huybrechts et al. (2004) hypothesized that such events may become more frequent in Greenland as storm tracks intensify or shift position with climate change.

2.1.3. Surface and basal properties and processes

Surface melting in the ablation zone (Figure 2.3) is highly sensitive to changes in surface albedo. Melting of ice in the ablation zone exposes a surface layer of dust that was originally deposited with snowfall high on the ice sheet. Dust in snow and on ice reduces the solar reflection (albedo) and, thus, increases surface melt further. Surface albedo

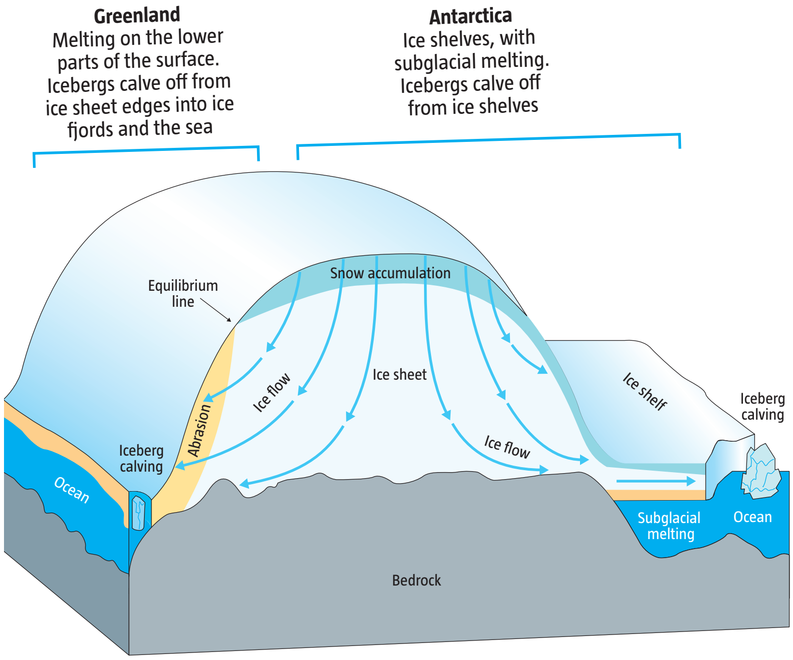


Figure 2.3 *The ice cover in Greenland and Antarctica has two components – thick, grounded, inland ice that rests on a more or less solid bed, and thinner floating ice shelves and glacier tongues. An ice sheet is actually a giant glacier, and like most glaciers it is nourished by the continual accumulation of snow on its surface. As successive layers of snow build up, the layers beneath are gradually compressed into solid ice. Snow input is balanced by glacial outflow, so the height of the ice sheet stays approximately constant through time. The ice is driven by gravity to slide and to flow downhill from the highest points of the interior to the coast. There it either melts or is carried away as icebergs that eventually melt, thus returning the water to the ocean from where it came. Outflow from the inland ice is organized into a series of drainage basins separated by ice divides that concentrate the flow of ice into either narrow mountain-bounded outlet glaciers or fast-moving ice streams surrounded by slow-moving ice rather than rock walls. In Antarctica, much of this flowing ice has reached the coast and has spread over the surface of the ocean to form ice shelves that are floating on the sea but are attached to ice on land. There are ice shelves along more than half of Antarctica's coast, but very few in Greenland. Sources: UNEP Maps and Graphs; Konrad Steffen, CIRES, University of Colorado.*

decreased dramatically at Swiss Camp during 1993 to 2008 and this led to increased surface melt and ablation at Swiss Camp (Figure 2.4).

Basal melt and meltwater exert a strong influence on ice flow (Fahnestock et al., 2001). The limited knowledge of basal melt is derived from models and sparse observations of bed properties. A basal melt rate of 0.1 m/y over a substantial area requires a regional geothermal heat flux of 970 MW/m^2, which is 17 times higher than the continental background of 56.5 MW/m^2 (Sclater et al., 1980). This does not account for heat conducted through the ice. For most of the ice sheet, the heat conducted through the ice is close to the background geothermal heat flux. It can be lower in low-accumulation regions because the thermal gradient in the basal ice is lower. The volume of meltwater being produced in the areas of high basal melt is of the order of 1 km^3/y. This water and warm basal ice are responsible for the onset of rapid iceflow in the large ice stream that drains the north side of the summit dome in Greenland (Fahnestock et al., 2001).

Basal flow can transport ice at velocities exceeding rates of internal deformation (i.e., from hundreds to more than 10 000 m annually) and glacier surges, tidewater glacier flow, and ice stream motion are governed by basal flow dynamics (Clarke, 1987). Glaciers and ice sheets that are susceptible to basal flow can move quickly and erratically, making them intrinsically less predictable than those governed by internal deformation. They are more sensitive to climate change because of their high rates of ice turnover which gives them a shorter response time to climate (or ice margin) perturbations. In addition, they may be directly responsive to increased amounts of surface meltwater production associated with climate warming. This latter process is crucial for predicting dynamic feedbacks to the expanding ablation area, longer melt season, and higher rates of surface meltwater production that are predicted for most ice masses.

Although basal meltwater has traditionally been thought to be the primary source of subglacial water, models have shown that supraglacial streams with discharges of over 0.15 m^3/s can penetrate down through 300 m of ice to reach bedrock, via self-propagation of water-filled crevasses (Arnold and Sharp, 2002). There are several possible subglacial hydrological configurations: ice-walled conduits, bedrock conduits, water film, linked cavities, soft sediment channels, porous sediment sheets, and ordinary aquifers (Mair et al., 2001; Flowers and Clarke, 2002).

Modern interest in water flow through glaciers originated in a pair of theoretical papers published in 1972. In one, Shreve (1972) discussed the influence of ice pressure on the direction of water flow through and under glaciers, while Röthlisberger (1972) presented a theoretical model for calculating water pressures in subglacial conduits in the other. Through a combination of these theoretical considerations and field observations, Röthlisberger concluded that the englacial drainage system probably comprises an arborescent network of passages. The millimeter-sized fingertip tributaries of this network join downwards into ever larger conduits.

Locally, moulins provide large direct connections between the glacier surface and the bed. Beneath a valley glacier, the subglacial drainage is likely to occur as a tortuous system of linked cavities transected by a few relatively large and comparatively straight conduits. The average flow direction in the combined system is controlled by a combination of ice overburden pressure and bed topography, and, generally, is not normal to contours of equal elevation on the bed.

Luthje et al. (2006) studied the summer evolution of supraglacial lakes on the Greenland ice margin using a one-dimensional (1-D) model to calculate the surface ablation for a bare ice surface and beneath supraglacial lakes for 30 days in the summers of 1999 and 2001. The surface ablation beneath the lake was enhanced by 110% to 170% for the two years 1999 and 2001 compared with the ablation for bare ice. Within the region of the ice sheet where supraglacial lakes presently occur, the area covered by supraglacial lakes was found to be more or less independent of the summer melt rate but controlled by topography. Luthje and co-workers predicted that, inside the ablation region, the area covered by supraglacial lakes will remain constant even in a warmer climate.

Although theoretical studies usually assume that subglacial conduits are semicircular in cross section, there are reasons to believe that this ideal is rarely realized in nature. Much of the progress in subglacial hydrology has been theoretical, as experimental techniques for studying the englacial hydraulic system are few and, as yet, not fully exploited and observational evidence is difficult to obtain. How directly and permanently do these effects influence ice dynamics? It is not clear at this time. The process is well known in valley glaciers where surface meltwater that reaches the bed in the summer melt season induces seasonal or episodic speed-ups (Iken and Bindschadler, 1986). Speed-ups have also been observed in response to large rainfall events (e.g., O'Neel et al., 2005).

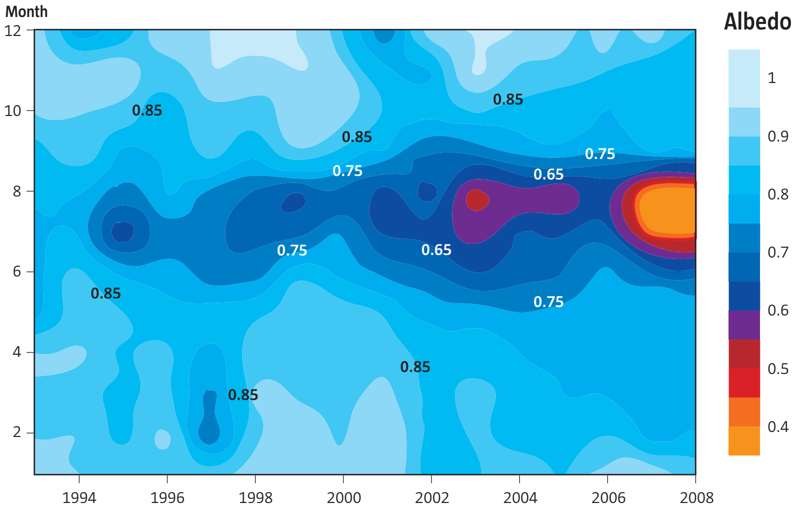


Figure 2.4 *Interannual variability of monthly mean albedo at Swiss Camp (1993-2008). Note the decrease in surface reflectivity (albedo) for the summer months (June–August) with values as low as 0.4 in 2007 and 2008. Increase in summer air temperatures at this elevation enhances snow and ice melt as well as reducing albedo. Hence, less solar radiation is reflected and this further increases the surface melt (positive albedo feedback). Source: Konrad Steffen, CIRES, University of Colorado.*

2.1.4. Summary

Is the climate in Greenland changing? Coastal records since 1840 show the warmest decades to be the 1930s and 1940s and a warming trend since 1988 of 2 to 4 °C in West Greenland, mainly during the winter months. Recent, short-term temperature records on the ice sheet in West Greenland (Swiss Camp) show a 2.2 °C warming in summer since 1991 with the highest air temperatures in 2007. Air temperatures at the top of the Greenland Ice Sheet (Summit) revealed a slight cooling of -0.3 °C between 1987 and 2005.

2.2. Surface mass balance

2.2.1. Introduction

The total mass balance of the Greenland Ice Sheet (section 2.4) determines its overall state of health and is the sum of two terms: the surface mass balance (SMB), discussed here, and solid ice discharge, discussed in section 2.3. The SMB is the net mass added or removed from the surface of an ice sheet by a range of processes. The aim of this section is to introduce the factors that control SMB, and how they respond to climate change. This is achieved by reviewing recent estimates of the components that make up the SMB, discussing the degree of consistency in these components, both from observations and modeling, and considering areas of uncertainty requiring further development.

The annual SMB is defined as the sum of mass fluxes towards and away from the ice sheet surface, integrated over a year and the area of the ice sheet. Mathematically, this is expressed as:

$$\mathrm{SMB} = \int \mathrm{SSMB}\ \delta A = \int (P + \mathrm{TMT} + R + \nabla \bullet Q)\ \delta t\ \delta A \qquad 2.1$$

where SSMB is the specific surface mass balance (the local SMB); A is area; P is precipitation, comprising the solid fraction (snow, hail, freezing rain) as well as the liquid fraction (rain); TMT is turbulent moisture transport, comprising surface evaporation, sublimation / deposition and snowdrift sublimation; R is run-off, comprising the liquid water fraction from (sub)surface melting and / or rain that leaves the firn / ice column; and Q is the snowdrift mass flux which, if it varies spatially, erodes / deposits snow on the ice sheet. The only positive term in Eqn 2.1 is P; all the other terms are responsible for mass loss.

Snowdrift plumes being blown into the ocean have been occasionally observed, but $\nabla \bullet Q$ is believed to significantly impact SMB only locally and is not further considered here (Box et al., 2006). The accumulation zone is the region where SSMB > 0 and the ablation zone is where SSMB < 0. The equilibrium line altitude is the elevation where SSMB = 0. For an ice sheet as a whole, to be in a state of balance, the SMB must match the solid ice flux flowing into the ocean. For the Greenland Ice Sheet, this was roughly the case prior to the 1990s but there has since been a marked increase in both solid ice flux (Rignot and Kanagaratnam, 2006) and run-off, R (Hanna et al., 2008).

The average SMB for the three reconstructions (MAR, PMM5, Hanna; see Table 2.1) covering the last 50 years is 285 Gt/y with a range of 62 Gt. This range is 22% of the mean value and provides some measure of the uncertainty in the SMB (although, using a similar approach, the uncertainty in the individual components is much greater). This compares with an estimate of precipitation of 582 Gt/y. Thus, around half the precipitation is lost as run-off during this period although estimates vary substantially (Table 2.1). It should be noted that run-off is not equal to the total volume of meltwater, M, because a proportion of this percolates into the underlying snowpack and refreezes. Considerable uncertainty exists in estimating this refreezing term, RF (see section 2.2.4).

There are a number of approaches for estimating SMB. The first, and the only tractable approach until quite recently, involves the interpolation of *in situ* observations of SSMB estimates from snowpits, ice cores, stake measurements and automatic weather stations to estimate net accumulation (Figure 2.5). The data were combined with simplified models of run-off, R, to determine the SMB (Ohmura and Reeh, 1991; Reeh, 1991; Ohmura et al., 1999; Bales et al., 2001; Steffen and Box, 2001). The more recent interpolations to determine accumulation rates also included a formal assessment of the error associated

Table 2.1 *Estimates of the SMB components in Gt/y for Greenland.*

Model	Period	Area × 10^7 km²	P	TMT	Rain	M	R	RF	SMB
MAR [a]	1958 – 2007	1.701	550	-5	20	532	-282	250	264
PMM5 [b]	1958 – 2006	1.691	638	-63	16	228	-213	48	326
Hanna [c]	1958 – 2007	1.678	559	-35	26	313	-261	77	264
Mote [d]	1988 – 1999	1.648	591	-69	23		-255		239
Reeh [e]	1990s	1.707	552 [f]				-279		273

P: Precipitation, comprising the solid fraction (snow, hail, freezing rain) as well as the liquid fraction (rain); TMT: Turbulent Moisture Transport, comprising surface evaporation, sublimation/deposition and snowdrift sublimation; Rain: liquid fraction of precipitation; M: total volume of Meltwater; R: Run-off, comprising the liquid water fraction from melt and rain that leaves the firn/ice column; RF: Re-Freezing; SMB: Surface Mass Balance.

[a] Modèle Atmosphérique Régional (Fettweis, 2007); [b] Polar Mesoscale Model version 5 (Box et al., 2004); [c] Hanna et al. (2002, 2005); [d] Mote (2003); [e] Reeh et al. (1999); [f] Reeh and co-workers (1999) did not estimate the individual terms in accumulation and the tabulated value, therefore, represents P – TMT – Rain.

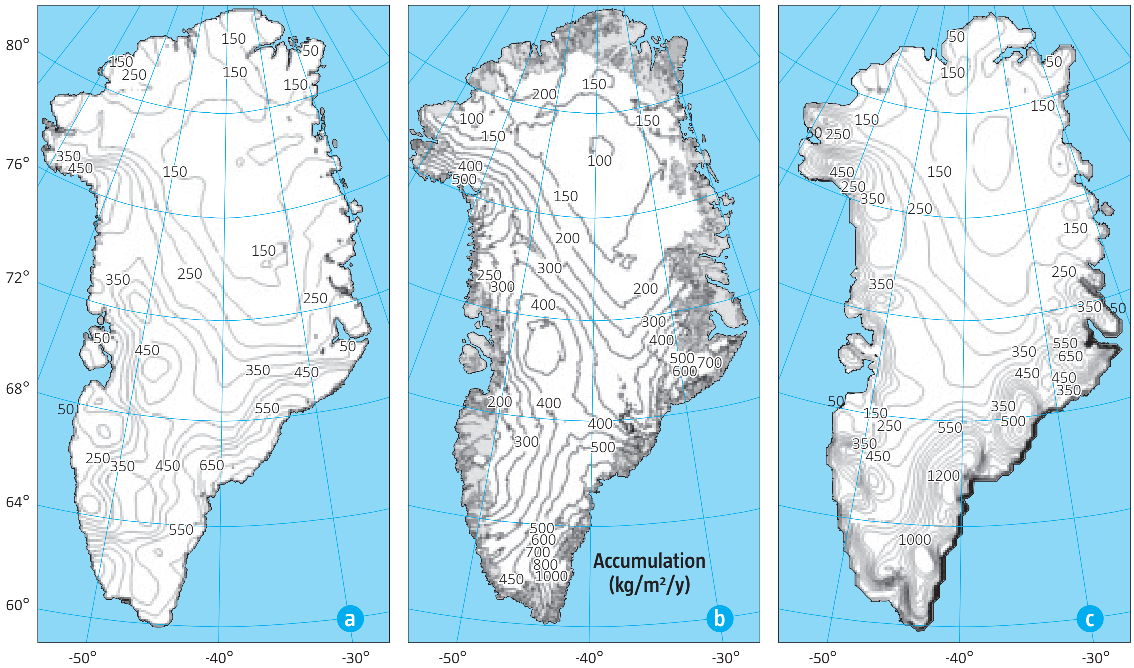


Figure 2.5 *Net accumulation (snowfall – TMT) from three recent estimates. Figures a and b are from observations, while c is from a regional climate model. Sources: a) Bales et al. (2001, 2009); b) Cogley (2004); c) MAR (Fettweis, 2007).*

with field data and its interpolation (Cogley, 2004; Bales et al., 2009), as discussed in section 2.2.2.

Commonly, run-off has been estimated using an empirical model of melt, M (see section 2.2.3), combined with simplified assumptions about refreezing, RF, calibrated using stake measurements (Reeh, 1991; Braithwaite et al., 1992; Braithwaite, 1995). The errors in run-off from this approach are less well defined, and it is apparent from Table 2.1 that estimates for R vary substantially (up to 25%) between different models and that the individual components contributing to run-off (M and RF) vary by far more than 25%. Thus, the agreement in R is largely based on compensating differences. It is also important to note that different run-off models have very different sensitivities to forcing and, as a consequence, the ability to predict the true response of the SMB to a future change in climate contains significant uncertainties (Bougamont et al., 2007).

Recently, alternative approaches for estimating SMB have been developed, especially following the release of a high-quality, consistent ~50 year global climatology produced by ECMWF. The data set is known as ERA-40 (i.e., ECMWF Re-Analysis covering the ~40 year period 1957-2002). ERA-40 is a hindcast generated using the ECMWF atmospheric forecast model driven by surface and satellite meteorological observations (Uppala et al., 2005). ERA-40 produces 6-hourly and daily fields of key atmospheric parameters at 1.125° resolution, which equates to a grid cell size of ~125 km in latitude. This resolution is too coarse to be used unadjusted for SMB calculation as the ablation zone, for example, is mostly less than a single grid cell in width. As a result, various approaches, commonly termed downscaling, have been used to interpolate ERA-40 to higher resolution. The simplest approach has been to downscale surface air temperature using seasonally adjusted lapse rates and high-resolution surface topography as well as a simple interpolation of precipitation (Hanna et al., 2002, 2005). This is the 'Hanna' data set referred to in Table 2.1 and Figure 2.6.

Another approach is to use an atmospheric regional climate model (RCM) as a physically-based 'interpolator' of the reanalysis data. Generally, an RCM is run at high resolution for a limited area and forced at its boundaries by climate data from a coarser-resolution global climate model. To date, two RCMs have been used to produce a high-resolution time series of the Greenland Ice Sheet SMB (Box et al., 2004, 2006; Fettweis, 2007) although others are under development.

It is interesting to consider how the SMB might respond to changes in climate. Increased near-surface air temperatures bring about increased moisture and, therefore, precipitation. Simulations with coupled atmosphere-ocean GCMs suggest a 5% /K sensitivity of P to surface air temperature (Gregory and Huybrechts, 2006). This value can be combined with SMB 'climate sensitivities' (especially to changes in temperature and precipitation) estimated from an energy balance model that suggests that the area-averaged SMB sensitivity, CT, to temperature is -49 mm/K and to precipitation, CP, 3.8 mm/% (Oerlemans et al., 2005). Using the sensitivity of precipitation to temperature, given above, makes it possible to express CP with respect to temperature, which gives CP = 19 mm/K. Thus the increase in run-off greatly outweighs the increase in accumulation (49 *vs* 19) for a fixed temperature change. Assuming the area of the ice sheet to be 1.7 million km^2 then the increased mass loss from the ice sheet is 51 km^3/K, excluding any dynamic response. Thus, simulations of the future SMB for various greenhouse gas emissions scenarios generally predict a decreasing SMB (Gregory and Huybrechts, 2006). Importantly, this is what has been suggested to have taken place over the last ~15 years (Figure 2.6) (Hanna et al., 2008).

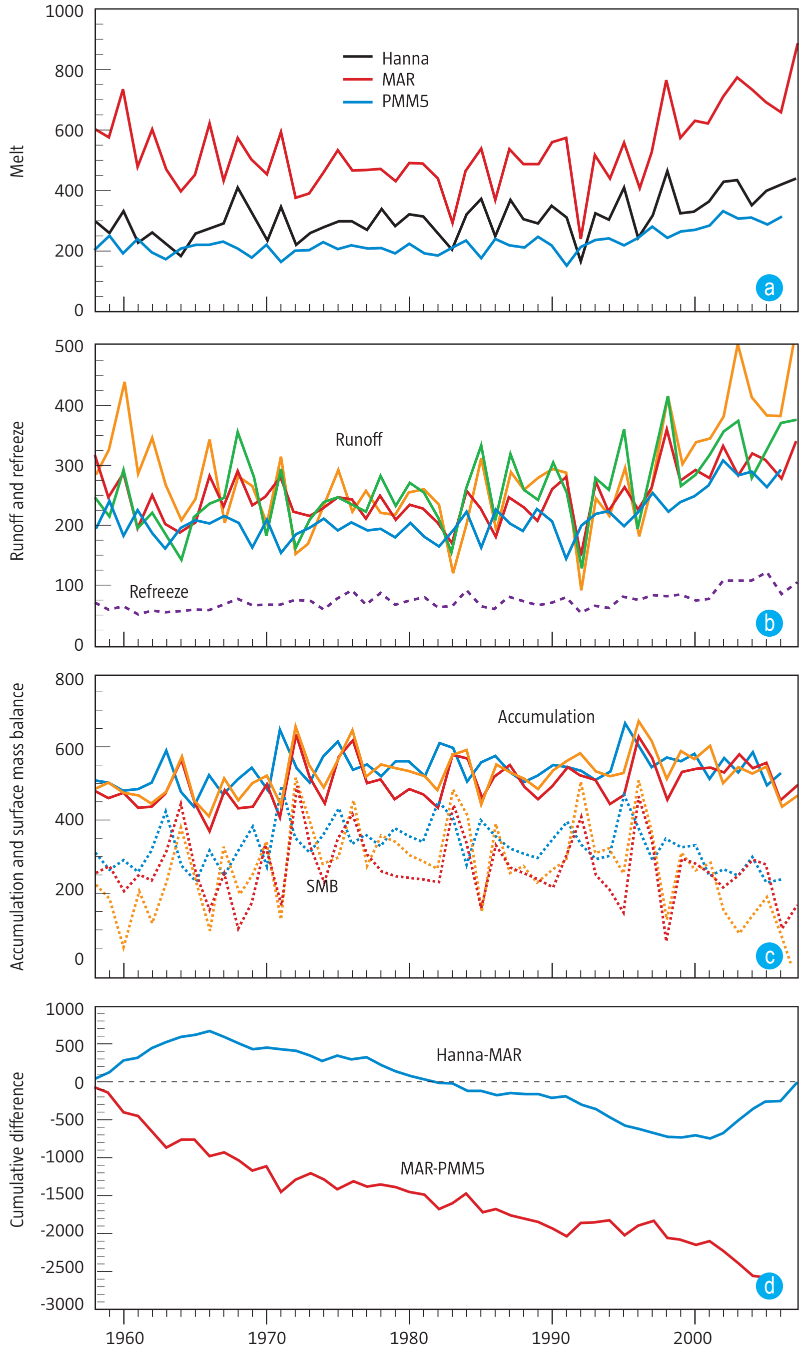


Figure 2.6 *Time series of surface mass balance components from three different models, discussed in the text. Units are in Gt/y. (a) Melt, (b) run-off (solid) and refreezing (dashed), (c) accumulation (solid), SMB (dashed), (d) cumulative difference in SMB for Hanna-MAR and MAR-PMM5. Note that the y-axis scales are not the same for each plot. Source: Jonathan Bamber, University of Bristol.*

2.2.2. Accumulation

2.2.2.1. Observations

The Program for Arctic Regional Climate Assessment (PARCA) is a NASA-funded, multi-decadal, coordinated effort to improve estimates of a wide range of key parameters for the Greenland Ice Sheet (Abdalati, 2001). Among the measurements taken were over 100 shallow ice core measurements of accumulation. These were combined with a subset of an additional 250 historical observations to produce the most comprehensive observation-based map of accumulation across the ice sheet to date. The estimated mean error for this compilation was 24% with regional variations of 15% to 30% (Bales et al., 2001). This compilation has recently been updated with a number of new cores and coastal station data (Figure 2.7) (Bales et al., 2009). The recent update resulted in regional changes, compared with the earlier study, of as much as ±50%, suggesting that uncertainties in accumulation rates were significantly underestimated in previous studies.

The average accumulation rate of 300 kg/m^2 over the ice sheet agrees well with an earlier estimate obtained from different sources and covering a different time period (Ohmura and Reeh, 1991), and with another estimate using a different interpolation approach (Cogley, 2004). The apparent agreement in the mean value could give rise to a false confidence in the reliability of estimated accumulation rates, but examination of the spatial distribution demonstrates marked differences (Figure 2.5).

Accumulation rates have also been derived from downscaling of climate re-analysis data (e.g., Fettweis, 2007).

An error analysis of *in situ* snow pit and ice core data suggested an uncertainty of 7.7% in the averaged 30-year mean. This is similar to the error estimate from the most recent study (Bales et al., 2009) and may be a reasonable value for the area- and time-averaged uncertainty. It does not reflect, however, the interannual and spatial uncertainties which are very much greater (Bales et al., 2009). Uncertainties in the spatial pattern can have a significant impact on modeled estimates of ablation due to the influence of snowfall on surface albedo and refreezing (see discussion in section 2.2.3).

In addition, rain is a small but important term. Although its volume is low (~4% of total precipitation), its impact can be important because upon refreezing it raises the heat content of the snowpack, and in doing so also increases total melt and run-off volume. The increased percentage of precipitation falling as rain in the future will have a stronger influence on run-off and, hence, the SMB than its percentage change might initially suggest.

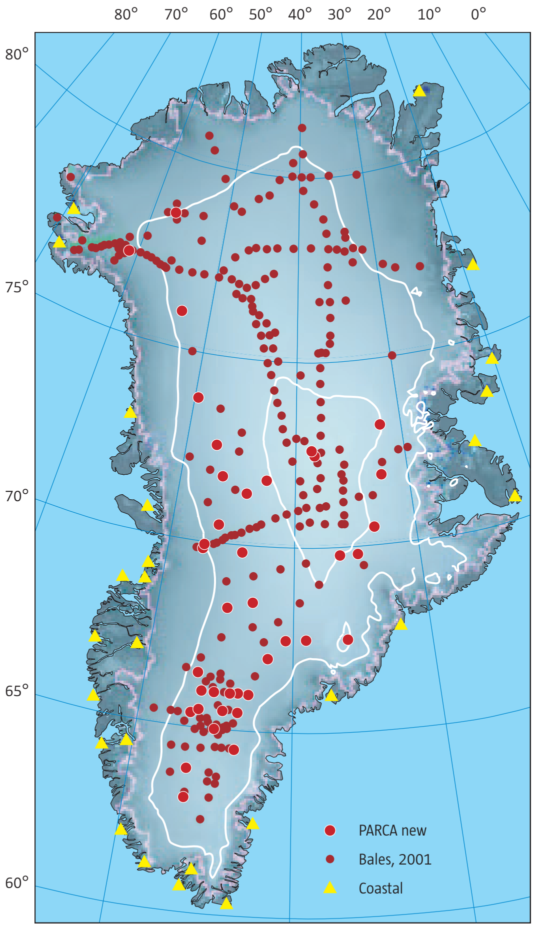


Figure 2.7 *Locations of ice cores and coastal meteorological station data used to reconstruct accumulation rates over the ice sheet. Source: Bales et al. (2009).*

2.2.2.2. Regional climate models

One of the most promising advances in assessing the SMB of the Greenland Ice Sheet, its temporal evolution and the relative importance of different components and their sensitivity to climate has been the development of RCMs. The first high-resolution RCM for Greenland was based on a modified version of the fifth generation mesoscale model known as PMM5 (Polar MM5; Box et al., 2004). This model ran at 24 km resolution and was calibrated using *in situ* data to correct for biases in melt energy and water vapor fluxes (Box et al., 2004). It was run, initially, for the period 1957-2004 forced by ERA-40 data (Box et al., 2006), see Table 2.1 and Figure 2.6.

More recently, a 25 km resolution RCM has been developed by Gallee and Schayes (1994) and applied to the Greenland Ice Sheet by Fettweis (2007) for the period 1958-2006. Unlike PMM5, the MAR (Modèle Atmosphérique Régional) has not been calibrated or corrected using *in situ* data. Comparison of modeled water vapor fluxes from MAR with automatic weather station data suggests that TMT is underestimated in MAR (Fettweis, 2007). Precipitation and run-off, however, are broadly comparable with other estimates from modeling and observations. This is surprising, because the individual terms that comprise the run-off (M and RF) are much greater than, for example, those of Hanna et al. (2008) (Table 2.1, Figure 2.6). Melt is ~80% and RF a factor ~9 times greater in MAR.

It should be noted that RCMs are the only approach where all the components of the SMB can be uniquely separated. *In situ* data provide the net SMB: i.e., precipitation minus TMT terms and run-off. Also, measuring melt volume and refreezing, in the field, is challenging (Bøggild et al., 2005). The models suggest that TMT is generally negative, indicating net sublimation / evaporation, with a magnitude of about 10% of P, except in MAR, where it is an order of magnitude smaller.

2.2.3. Ablation

2.2.3.1. Observations

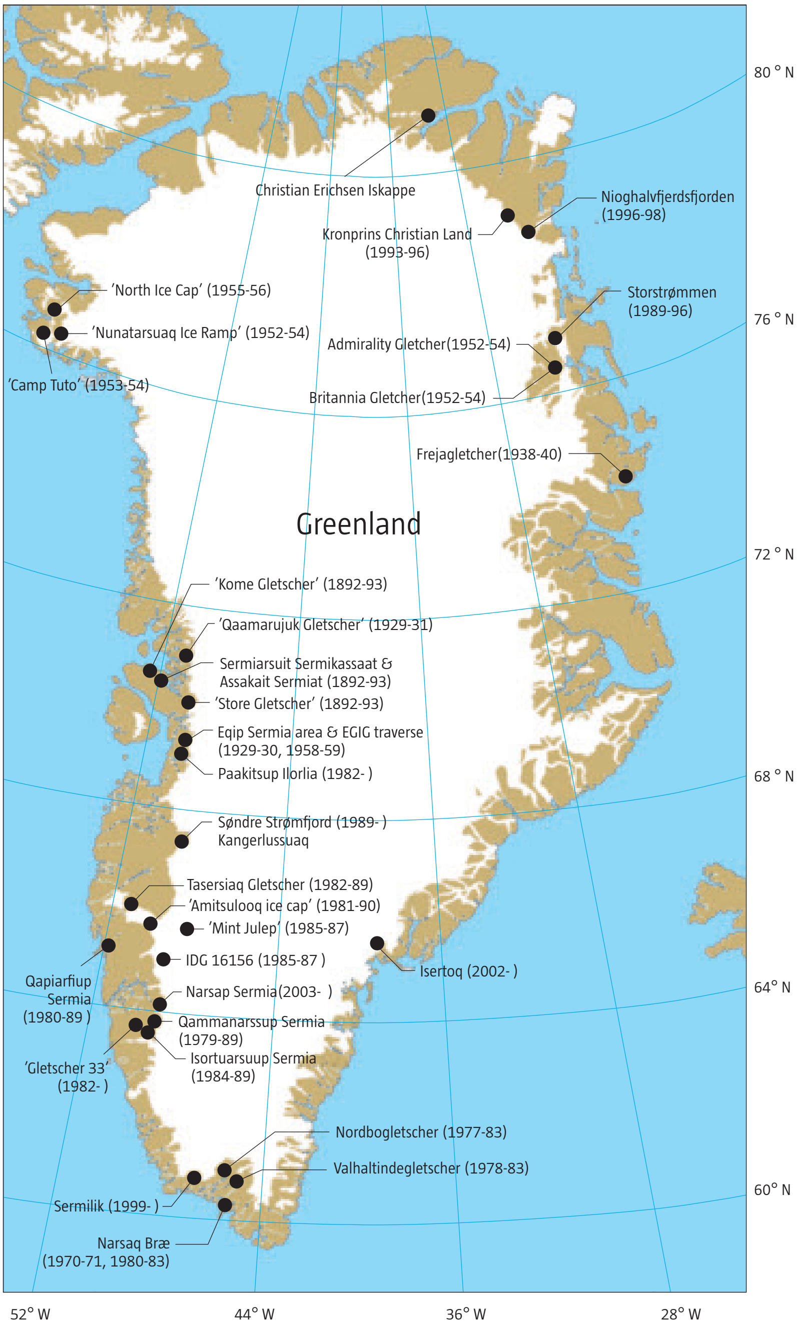
Direct observations of ablation rates around the margins of the Greenland Ice Sheet are sparse both in space and time (Figure 2.8). Ablation has been measured using stakes at a number of sites in Greenland (Braithwaite et al., 1992) and, more recently, a technique for making automated measurements has been proposed (Bøggild et al., 2004a). However, the spatial coverage is far from representative. The most comprehensive and uninterrupted *in situ* data set has been collected since 1990 by Utrecht University along a 150-km transect near Kangerlussuaq, central west Greenland known as the K-transect (Greuell et al., 2001; van de Wal et al., 2005). Stake measurements of mass balance have been collected at eight sites along the K-transect since 1990 and automatic weather station data are available since 1997 at one site and since 2003 at two further sites.

This represents the longest continuous record of *in situ* mass balance and ablation data for the Greenland Ice Sheet, but how representative of other parts of the ablation zone the data are is uncertain. For example, a modeled time series used to assess the spatial autocorrelation of SMB showed little autocorrelation in the ablation zone in the southeast, a negative correlation in the northeast, and a strong positive correlation along the western margins suggesting that ablation depends on local conditions (as would be expected) as well as larger-scale climate forcing (Box et al., 2006). Although the *in situ* data cannot be used to provide direct estimates of ice-sheet integrated ablation or the individual ablation components, they can be used to validate and calibrate satellite and model-based estimates.

Run-off has also been estimated using calibrated satellite observations of passive microwave melt extent and duration (Mote, 2003) and satellite-derived albedo data (Greuell and Oerlemans, 2005). Both approaches rely on an empirically derived relationship between melt duration or albedo and run-off. Estimating run-off from routinely acquired satellite observations is a promising technique but is empirically based and it remains to be shown whether the relationships derived are robust under current and future (changing) climate conditions. Recent results have shown, for example, that for areas in Antarctica where melt occurs, passive microwave-derived values underestimate the SMB (Magand et al., 2008).

In addition to measuring run-off, passive microwave data have also been used to measure the area of the Greenland Ice Sheet subject to surface melt (Abdalati and Steffen, 2001; Mote, 2007). Results indicate that the melt area has been steadily increasing since satellite observations began (Figure 2.9). Although this is not a direct measure of run-off (because it does not estimate the volume of

Figure 2.8 *Location of ice sheet ablation measurements up to 2004. Source: Bøggild et al. (2004b).*



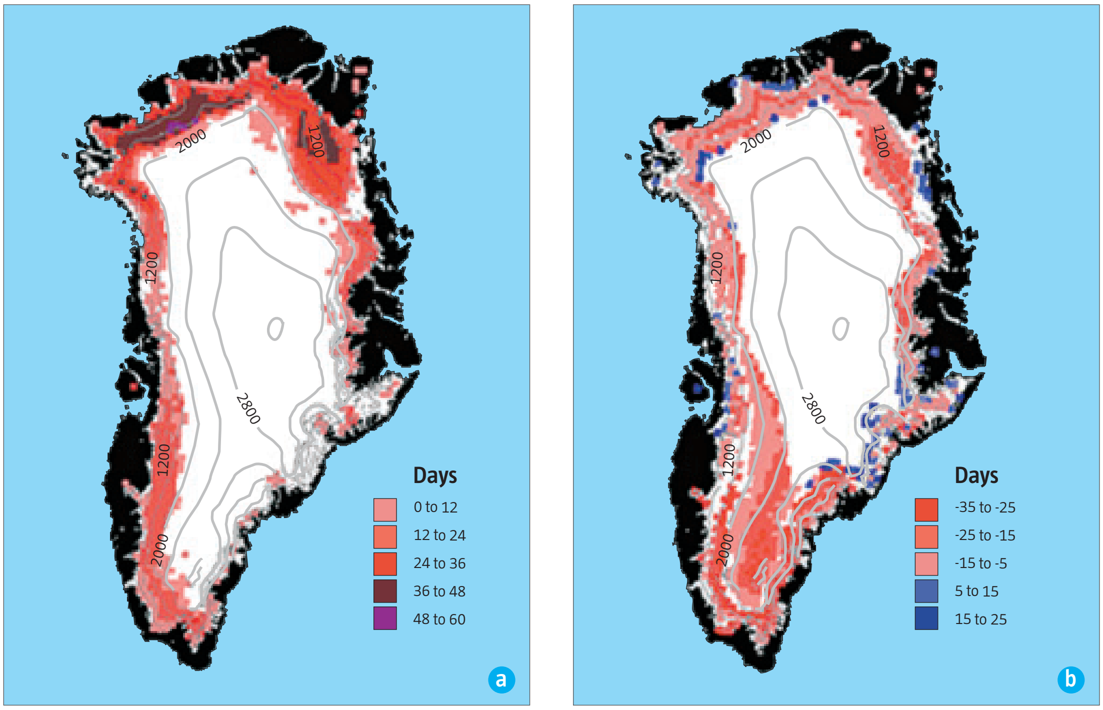


Figure 2.9 *a) Departure of the 2008 melt frequency from the mean melt frequency over the period 1973-2008. b) Departure of the date of melt onset in 2008. Elevation contours are shown in grey, and black indicates land area. Source: Mote et al. (2007); TL Mote, pers. comm.*

melt nor refreezing) there is a good correlation between melt area, number of melt days, and run-off (Mote, 2003).

In this approach, the satellite-based melt duration data are used as a proxy for surface air temperature. Although this method has some merit, the development of a consistent, homogeneous temperature data set derived from ERA-40 and extending back to 1958, provides a longer time series with roughly the same correlation with run-off (Hanna et al., 2002, 2005). Its advantage, however, is that it represents a direct observation of changing surface properties as opposed to a model reconstruction and can, therefore, be used to corroborate the reconstructions derived from re-analysis data. Plus, it provides observational evidence for changes in the length of the melt season and its onset date (Figure 2.9).

2.2.3.2. Run-off modeling

The first reasonable quantitative attempts to model run-off from the Greenland Ice Sheet used a positive degree day (PDD) approach (Reeh, 1991). Here, the melt volume is related to the number of days where the near surface air temperature (usually taken at 2 m above the surface) exceeds 0 °C, and by how much. Melt volume is calculated by multiplying the positive temperature by a degree day factor (typically ~ 8 mm/°C per day for ice). This approach has been, and still is, used extensively for modeling the present-day run-off and SMB of the Greenland Ice Sheet (Braithwaite, 1995; Abdalati et al., 2001; Mote, 2003; Hanna et al., 2006; Rignot and Kanagaratnam, 2006) as well as its response in the future (Huybrechts et al., 2002, 2004).

The compelling advantage of the PDD approach is that it relies on a single, smoothly varying and easily interpolated parameter, i.e., near surface air temperature. Climatological maps of this parameter have been produced from a combination of coastal station and inland automatic weather station data (Ohmura, 1987) and more recently downscaled from ERA-40 (Hanna et al., 2002, 2005). The weakness of the approach is that it relies on an empirically based relationship, calibrated using a limited range of data for present-day conditions only. As a result, it is unclear how well a PDD model can predict melt in the future (Bougamont et al., 2007). For example, a PDD model has no sensitivity to changes in cloud cover or type unless they impact near surface air temperature. Modifications to the approach have been suggested to incorporate more of the physical processes involved (Pellicciotti et al., 2005) but remain, fundamentally, empirical.

A more physically rigorous approach is

to calculate the energy balance at the surface and, from this, the quantity of energy available for melting. Most recently, energy balance models have been used to estimate melt volume over Greenland by downscaling meteorological hindcast data such as ERA-40 (Box et al., 2004, 2006; Fettweis et al., 2005). Generally, these types of model incorporate all the key processes that affect melt but, as a result, require a large number of meteorological inputs that may not be constrained particularly well. For example, to estimate the turbulent, sensible and latent heat fluxes requires prescribing a surface roughness as well as atmospheric stability corrections. It also requires near-surface humidity and the vertical wind profile in the boundary layer, all of which may not be particularly well known.

The single most important parameter for controlling run-off in the energy balance, however, is surface albedo, α. This can be estimated by the model based on snow age and / or density (Greuell and Konzelmann, 1994; Konzelmann et al., 1994) or prescribed from satellite observations (Stroeve, 2001). The advantage of an energy balance model compared to a PDD approach is that it incorporates the physical processes responsible for melt explicitly and should, therefore, respond appropriately to changing climate forcing. To do this, however, requires the 'correct' forcing. For long integrations (e.g., multi-centennial) such input data typically do not exist or may not be reliable enough to justify the use of a more complex and more 'data hungry' energy balance model. Thus, although an energy balance model may provide a more physical representation of the processes, it will not always be the best tool to use.

Preconditioning of the snowpack and the firn is very important for subsequent melt and run-off. In the Greenland Ice Sheet 1958-2003 annual SMB series of Hanna et al. (2005), high run-off years (except 2003) are generally synchronous with low precipitation / accumulation and *vice versa*. More accumulation results in a higher albedo for a longer time, which, in turn, reduces absorbed energy available for melt; the available surface energy needs to melt any snow first before ice can melt in the ablation region. Also, higher volumes of meltwater are retained in the thicker snowpack, which tends to reduce net run-off. Thus, accurate reconstruction of the timing, pattern and volume of accumulation is important for reliable estimation of run-off.

Another important consideration in modeling run-off is refreezing of meltwater within the snowpack. A variety of approaches exist from the simplest, using a fixed fraction of the annual accumulation (Reeh, 1991) to full snow metamorphism models (Bougamont et al., 2005; Mernild et al., 2006; Fettweis, 2007). The latter contain explicit terms for the diagenesis of the snowpack as meltwater percolates into it but can be driven only by an energy balance model because they require the energy balance of the snowpack to be calculated, which is not possible using a simpler approach. A study of three relatively simple refreezing schemes for the Greenland Ice Sheet found that they accounted for ~20% of the total melt production, produced similar estimates, and had reasonably similar sensitivities to changing climate (Janssens and Huybrechts, 2000).

More recent analyses incorporating these simple approaches as well as snow metamorphism models reached a different conclusion, however, and found that the different models produced both different volumes of refreezing and had different sensitivities (Bougamont et al., 2007; Wright et al., 2007). The MAR RCM, for instance, estimates that refreezing is ~50% of the total melt production (Table 2.1). These studies suggest that further work on understanding and modeling refreezing is required. They also indicate that modeling refreezing remains a significant uncertainty in present-day and future estimates of SMB. A major hurdle for improving refreezing estimates is the difficulty in obtaining reliable *in situ* validation data (Bøggild, 2000).

2.2.4. Differences and trends in measured and modeled SMB

Table 2.1 summarizes some recent estimates of Greenland SMB, derived from a variety of methods. It is important to note that the values presented in Table 2.1 are not entirely independent. The models Mote and PMM5 derived accumulation rates from almost identical downscaling approaches. Hanna, MAR and PMM5 all use the same source: ECMWF climate re-analysis (known as ERA-40) and operational analysis data as the input for their downscaling approach. Thus, biases in ERA-40 may affect all of these estimates in similar ways.

Estimates of the net SMB from downscaling of ERA-40 data agree reasonably well when averaged over several years but there are substantial differences in spatial patterns, individual components of the SMB, and interannual variability. Figure 2.6 compares the components of the SMB from several different approaches: A simple downscaling of ERA-40 derived by Hanna et al. (2005) and the RCMs, MAR (Fettweis, 2007) and PMM5 (Box et al., 2006). It is not surprising that the trends are in agreement and that there is reasonable correlation between minima and maxima as this is largely constrained by the common forcing used in the simulations (i.e., ERA-40). There are, however, substantial departures in absolute values. For 2003, for example, the Fettweis (2007) SMB anomaly is ~183 Gt less than the other two estimates, which is about the value of the SMB for that year in MAR.

Figure 2.6d shows the cumulative difference in SMB for Hanna-MAR, and MAR-PMM5. In the former, the cumulative difference reaches ±642 Gt. In other words, depending on the time period chosen, the magnitude of model-dependent cumulative difference in SMB can be greater than the total mass loss (the sum of solid ice flux and run-off) from the Greenland Ice Sheet each year. The standard deviation of the difference between the Hanna and the MAR annual SMB estimates is ±78 Gt compared with an interannual variability of 95 Gt and 124 Gt for Hanna and MAR, respectively.

In addition, the refreezing component, RF, is highly uncertain. The range for RF between the three time series that use ERA-40 data as input is from 49 to 250 Gt/y: a difference that is 70% of the mean SMB. Refreezing is, therefore, an important, yet poorly known, component of the SMB, which has received relatively little attention compared with other terms, partly due to the difficulty in making direct measurements of its magnitude. Although the fundamental physical processes that influence SMB are, in the main, relatively well understood, Table 2.1 and Figure 2.6 clearly show that much uncertainty still remains in constraining it.

Figure 2.10 summarizes the main trend in SMB over the past 50 years. During the 1960s and the early 1970s, SMB was 10% to 20% below the 50-year mean. Since the mid-1990s, however, SMB has decreased steadily to its lowest value over the entire reconstructed period, such that the most recent years are more than 100 Gt (or 35%) below the 50-year mean and 200 Gt below the value in the mid-1990s. This is the most striking trend in the time series and has been linked to a global, rather than regional, warming trend that exceeds the background natural variability (Hanna et al., 2008). This is an important conclusion, which suggests that the trend is likely to continue into the future.

2.2.5. Summary and recommendations

The mean SMB, based on the three ~50-year records, is 284 Gt/y with a range between estimates that is 22% of the mean, providing a qualitative indication of the uncertainty in the SMB. This is about half the standard deviation of the interannual variability, which ranges from 62 Gt/y for PMM5 to 124 Gt/y for MAR. Since the mid-1990s, run-off has increased significantly with only a modest change in accumulation, resulting in a reduction in the SMB of around ~200 Gt over the last 13 years (Figure 2.10). This trend in SMB appears to be a response to global climate change and above the level expected due to natural variability (Hanna et al., 2008). Also of note, is that 2007 had the lowest SMB of any year in the 50-year time series.

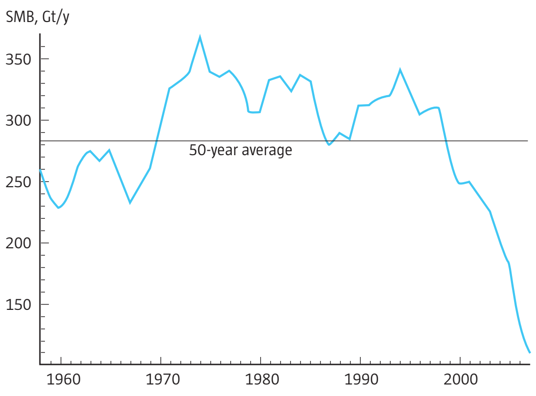


Figure 2.10 *Mean SMB of the three reconstructions (MAR, PMM5, Hanna) shown in Figure 2.6. The data have been smoothed over a running 5-year period to reduce interannual variations. Source: Jonathan Bamber, University of Bristol.*

Significant differences exist between the various estimates of the components of the SMB derived from numerical modeling and downscaling of climate re-analysis data. These differences can be as large as the net SMB for a given year and the standard deviation of the differences is of the order of 92 Gt. This is of the same order of magnitude as the inferred increase in mass loss due to changes in ice dynamics over the past decade (Rignot and Kanagaratnam, 2006).

The differences suggest that mass budget calculations (where the solid ice flux is subtracted from the SMB to estimate the overall net mass balance of the Greenland Ice Sheet) are seriously hindered by uncertainties in the SMB. These uncertainties originate from a range of sources. To reduce them to a level useful for mass budget calculations and reliable SMB predictions, will require a concerted effort to collect more targeted *in situ* SMB data, especially from the percolation and ablation zone, with which to calibrate and validate the models. Improvements in process understanding and modeling of key processes such as refreezing, blowing snow and subgrid-scale effects will also be needed.

2.3. Ice discharge

The Greenland Ice Sheet loses mass by melting as well as by discharge of ice into the ocean. The discharge consists of calving of icebergs as well as submarine melting. This section addresses the mechanisms of fast ice flow leading to transport of ice from the inland to the margin and the mechanisms of iceberg calving and submarine melting.

The margin of the Greenland Ice Sheet is characterized by spatially variable flow; areas of slow flow are separated by fast-flowing outlet glaciers and ice streams (Figure 2.11; Weidick, 1995; Rignot and Kanagaratnam, 2006). Parts of the ice are in direct contact with the surrounding ocean. In southern Greenland, ice-ocean contact is restricted to fast-flowing outlet glaciers that are topographically constrained by fjords, which often extend subglacially into the present-day ice sheet. The slower flowing ice usually terminates on dry land inland from the coast. However, in southwest Greenland ice marginal lakes are quite common, and ice can be lost there by freshwater calving (Warren, 1991). The role of calving into ice marginal lakes has not been properly investigated. It is likely to become increasingly important as the Greenland Ice Sheet retreats into the bedrock depression created by its own weight. In the north, along a significant section of the ice margin, both slowly and rapidly flowing ice is in direct contact with the ocean.

Around the Greenland Ice Sheet, the highest rates of ice discharge and the highest variability of discharge are observed at ocean-terminating outlet glaciers (Figure 2.11). The major mass loss mechanism for these glaciers is calving into the ocean (Reeh, 1968, 1994) and melting from ocean water (Rignot, 1996; Rignot et al., 1997). Rignot and Steffen (2008) showed that at Petermann Glacier 80% of the ice discharged into the ocean can be due to melting under the floating glacier tongue. Commonly, estimates of ice discharge into the ocean implicitly include mechanical calving as well as melting by the ocean, since the ice contributes to sea level as soon as it crosses the grounding line and begins to float. Many of the outlet glaciers with high discharge rates flow through deep and narrow fjords. The largest and best studied of these is Jakobshavn Isbræ on Greenland's west coast, for which the mechanisms of fast flow have been examined in detail (Iken et al., 1993; Funk et al., 1994; Lüthi et al., 2002, 2003). The findings are summarized here to illustrate the mechanism of fast outlet glacier flow.

Jakobshavn Isbræ is characterized by a deep subglacial trough that extends to 1500 m below sea level about 50 km inland from the current ice front (Clarke and Echelmeyer, 1996). This fjord (i.e., the trough) is relatively narrow (~ 5 km) but channels the ice from a drainage area that comprises about 6.5% of the Greenland Ice Sheet (Bindschadler, 1984; Echelmeyer et al., 1991). Ice flow in such a channel is inherently 3-dimensional. The various commonly-used ice flow models that are simplified by approximations based on shallow ice (i.e., where ice thickness is much smaller than horizontal extent) are not adequate to fully describe the flow in this type of channel. Such models include the Shallow

Ice Approximation (Hutter, 1983), and the Shallow Shelf Approximation (MacAyeal and Thomas, 1982), which can also be applied to shallow ice streams with low basal drag (MacAyeal, 1989). Instead, a fully 3-D thermomechanical treatment of ice flow, without approximation of the stress terms in the momentum balance, is necessary (Hutter, 1983) to describe outlet glaciers whose depths are comparable to their widths. No such model has yet been applied to either the Greenland or Antarctic ice sheets or even to an entire drainage basin of one of these ice sheets. (The hierarchy of ice flow models is discussed in section 3.2.)

The relatively steep and very thick ice of Jakobshavn Isbræ creates high driving stresses that are balanced by high basal drag, particularly at the flanks of the subglacial trough (Lüthi et al., 2003; Truffer and Echelmeyer, 2003). The high driving stress influences the stress field far into the surrounding ice sheet. Using a numerical model, Lüthi et al. (2003) found differences to sheet flow at lateral distances from the ice stream of seven times the ice sheet thickness. The high stresses lead to high deformation rates, because they are combined with soft ice at depth, due to a temperate layer and lower viscosity Wisconsian ice (Funk et al., 1994; Lüthi et al., 2002; Truffer and Echelmeyer, 2003).

The presence of this temperate ice adds an additional modeling challenge, because it introduces another free boundary: the cold-temperate surface (Hutter, 1982) and it requires the treatment of waterflow in the temperate layer. Also, the existence of a temperate basal layer implies water at the bed and a certain amount of basal motion. This is a boundary layer process that needs to be parameterized through a 'sliding law'. The most commonly used sliding laws (see also section 3.2) relate the basal motion to the basal shear stress and the subglacial water pressure; the latter is generally unknown but thought to be relatively high, i.e., comparable to the ice overburden pressure.

This type of sliding law has well-known limitations, which are due to spatially and temporally variable parameters and to the limited knowledge of spatial and temporal variability of basal water pressure, which is an important variable (e.g., Iken and Truffer, 1997). The present inability to predict the nature and the amount of basal motion under large changes in geometry, water supply, and stresses is the major limitation to credibly predictive models of outlet glaciers.

Reeh (1994) used mass balance considerations to estimate a total ice discharge into the ocean of 316 Gt/y. This is higher than a subsequent estimate of 170 to 270 Gt/y, which is based on calving rate estimates from parameters such as ice thickness at the front and water depth (Biggs, 1999). Both estimates applied for some time before the 1990s. Reeh's (1994) estimate is very close to one by Rignot and Kanagaratnam (2006) for the 1995, which was derived from measurements of ice surface velocity and thickness at flux gates, as well as an estimate of surface ablation downstream from these gates. Rignot and Kanagaratnam (2006) also estimated discharge for 2000 and 2005, which showed a progressive increase in

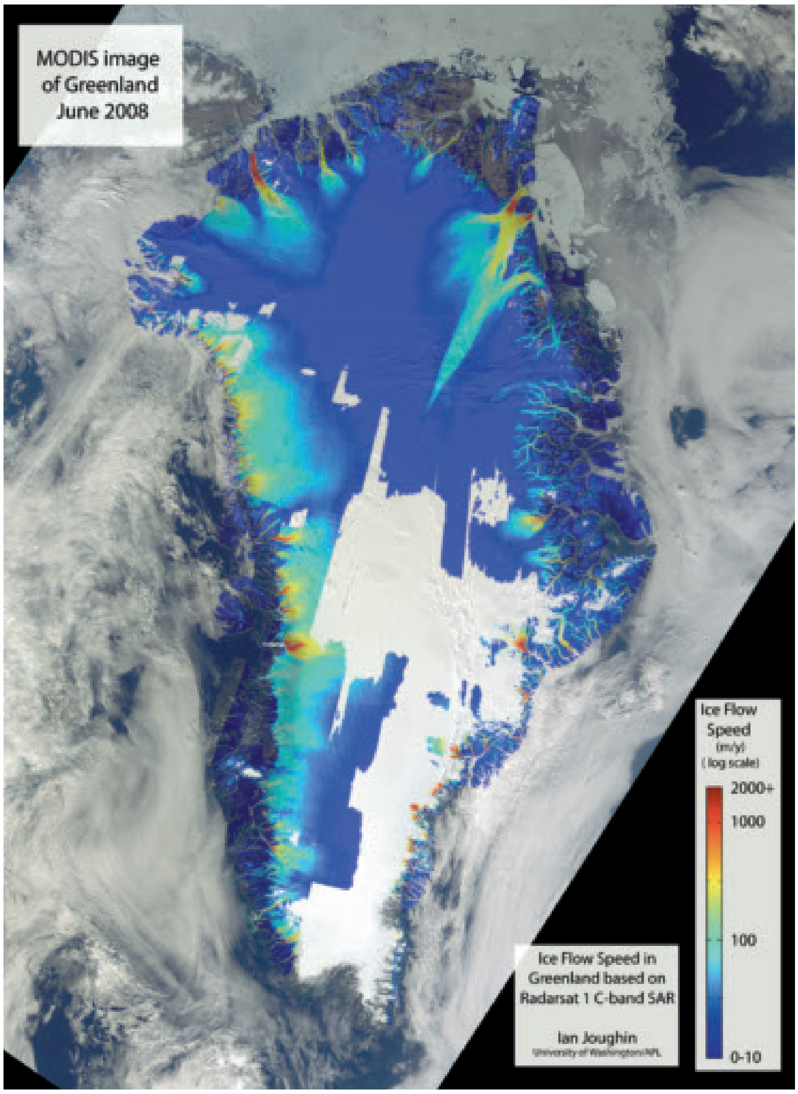


Figure 2.11 *Velocity mosaic of the Greenland Ice Sheet derived from radar satellite data. The mosaic shows how the slow flowing inland ice transitions into the fast moving outlet glaciers, and the large variability of flow along the margin. Source: Joughin I and Fahnestock M, unpubl. (2008).*

Table 2.2 *Calving flux estimates.*

Calving flux estimates, Gt/y	pre-1990	1995	2000	2005
Reeh (1994) [a]	316			
Biggs (1999) [b]	170-270			
Rignot and Kanagaratnam (2006) [c]		321	354	421

[a] Estimate is based on mass balance considerations; [b] estimate based on empirical calving rate relationships derived in Greenland and elsewhere; [c] estimate derived from measurements of surface velocity and ice thickness at flux gates and a correction for surface ablation downstream of these gates. The measurements have been converted to Gt/y assuming a density of 900 kg/m³.

discharge within just a few years (Figure 2.12). The various estimates are summarized in Table 2.2. It is clear that ice calving into the ocean is a significant ice loss mechanism for the Greenland Ice Sheet, and that it can be highly variable (Bamber et al., 2007). Ice calving is a major factor accounting for the recent negative mass balance of the Greenland Ice Sheet.

The variability of ice discharge is confirmed by observations of individual outlet glaciers, which have shown dramatic changes in recent years. In the period 2002-2007, Jakobshavn Isbræ went through a period of thinning, retreat by loss of its floating tongue, and acceleration to twice its former speed (Thomas et al., 2003; Joughin et al., 2004, 2008c; Podlech and Weidick, 2004; Luckman and Murray, 2005). There is no evidence of similar increases in discharge in the 20th century at Jakobshavn Isbræ (Weidick and Bennike, 2007), but old moraines and glacier trim lines do indicate that the glacier has lost its floating tongue at least once since the Little Ice Age (15th to 19th centuries), while it was retreating to a position that remained stable for the next 50 years (Csatho et al., 2008). Large changes have also been observed at two big glaciers on Greenland's east coast: Kangerdlugssuaq Glacier and Helheim Glacier (Howat et al., 2005, 2007; Luckman et al., 2006; Stearns and Hamilton, 2007).

Rignot and Kanagaratnam (2006) showed that the acceleration of ice discharge is widespread. Between 1995 and 2000 many of the outlet glaciers south of 66° N accelerated,

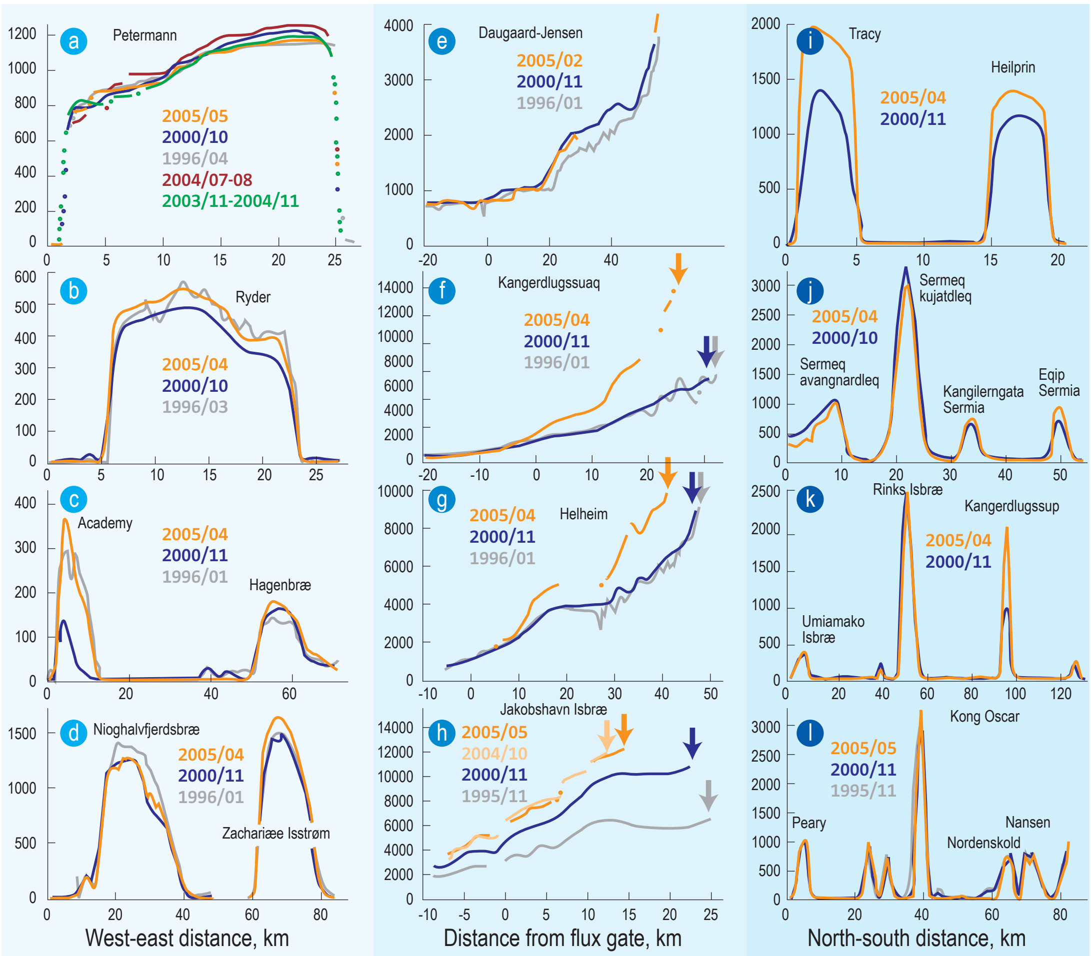


Figure 2.12 *Flow acceleration at a selection of outlet glaciers. Cross profiles of velocity (a-d and i-l) were used for ice discharge calculations. e-h show along-flow velocity profiles. Note that acceleration is accompanied by glacier retreat. Source: Rignot and Kanagaratnam (2006).*

often doubling their discharge. This pattern spread to 70° N by 2005, and appears to have been related to regional warming. The detailed pattern is complicated, however. For example, Howat et al. (2007) reported that the Helheim and Kangerdlugssuaq glaciers decreased their discharge again to near former levels in 2006. This was partially due to a slow-down connected with a re-advance of the terminus, and partly due to thinning, so that at similar ice velocity less volume can be discharged. Other glaciers, such as Jakobshavn Isbræ, have maintained their high rates of ice discharge (Dietrich et al., 2007; Amundson et al., 2008; Joughin et al., 2008c).

The out-of-balance flux of ice into the ocean from many of southern Greenland's outlet glaciers has led to large thinning of near margin ice. This thinning is affecting large parts of the drainage basins (Stearns and Hamilton, 2007; Joughin et al., 2008c). For Jakobshavn Isbræ, the thinning is closely related to increased ice flow that can be accounted for by an inland-propagating change in ice surface slope (Joughin et al., 2008c).

Coincident with ice-flow acceleration is a widespread retreat of outlet glacier terminus positions. Moon and Joughin (2008) recorded terminus positions of 203 outlet glaciers and concluded that retreat accelerated during 2000 to 2006 compared to 1992 to 2000. Again, the detailed picture is more complicated and Moon and Joughin suggested that there is a close relationship between air temperature and ice front advance / retreat, without suggesting that there is a cause and effect relationship between the two.

The observed acceleration of outlet glaciers coincides with an increase in teleseismically discovered glacial earthquakes (Ekström et al., 2003, 2006). The earthquakes occur seasonally and are strongly related to the occurrence of large calving events (Joughin et al., 2008b). Ekström et al. (2003) and Tsai and Ekström (2007) attributed the seismicity to large slip events on outlet glaciers. Recent ice displacement measurements do not confirm this hypothesis (Amundson et al., 2008; Nettles et al., 2008), and the former authors favor a model in which the glacial earthquakes are generated by the scraping of large overturning icebergs. Only a few seismic studies have been carried out on the ice and they point to a rich variety of seismic sources, many of which are presently not well understood (Rial et al., 2009).

Various hypotheses have been proposed for the causes of the speed-up of outlet glaciers. Thomas (2004) and Johnson et al. (2004) presented a model of Jakobshavn Isbræ in which the floating tongue provides resistance (i.e., backstress) to ice flow. The loss of the floating tongue then leads to ice flow acceleration. This model is also invoked by Joughin et al. (2008c) who observed a striking correlation between the speed of Jakobshavn Isbræ and the seasonally varying position of the calving front. Howat et al. (2005) used a force balance method to show how the loss of grounded ice near the terminus of Helheim Glacier leads to flow acceleration. The recently observed increases in ice discharge were always coincident with the loss of a floating tongue. But once the process is initiated, a pattern of fast flow, thinning and retreat can become established. This is a common observation at temperate tidewater glaciers, which only rarely develop floating tongues (Meier and Post, 1987). The idea of rapid tidewater glacier retreat is directly linked to that of a reverse bed slope, i.e., the glacier bed deepens in the up-glacier direction. Nick et al. (2009) suggested that, in the case of Helheim Glacier, the phase of accelerated flow may be relatively short-lived, but it remains to be seen how universally applicable this model is.

While the exact form of a calving law remains unknown (e.g., Benn et al., 2007), it is clear from observations that a retreat into deeper water will lead to higher rates of calving, thinning, and further retreat. Vieli et al. (2001) and Schoof (2007) showed that no stable terminus position can exist on a reverse bedrock slope. This holds in advance as well as in retreat. The glacier acceleration that accompanies retreat can then be due to the effect of thinning, as suggested by Meier and Post (1987) and used by Pfeffer (2007).

Thinning influences glacier velocity in several ways: (1) Spatially-uniform thinning leads to a reduction in driving stress and thus in deformation velocity. (2) Thinner ice leads to a reduction in effective stress (ice overburden minus basal water pressure). This is because the basal water pressure has to exceed a level that is given by sea level for subglacial drainage to occur. As a tidewater glacier thins, basal water pressure cannot decrease due to this set downstream level, causing an overall drop in effective pressure. The reduction of effective pressure, as the lower part of the glacier approaches floatation, is expected to lead to higher rates of basal motion in any reasonable sliding law. (3) Enhanced thinning near the terminus leads to increases in surface slope and hence driving stress and ice velocity.

Pfeffer (2007) derived a stability index for tidewater glaciers by considering the effects of (1) and (2), which have opposite effects on ice flow. Pfeffer showed that an ice thickness above a certain threshold:

$$h_{th} = C \cdot \rho_w / \rho_i \, h_w \qquad 2.2$$

is stable against thinning in that thinning leads to a reduction in ice flow, whereas an ice thickness below that threshold reacts to thinning by ice acceleration. Here, h_w is the depth of the water and ρ_w and ρ_i are the densities of ocean water and glacier ice, respectively. C is a constant that depends on the choice of the ice flow law and the sliding law. Pfeffer (2007) suggested that C = 4/3, given flow parameters in the literature.

The analyses of Vieli et al. (2001), Schoof (2007), and Pfeffer (2007) all apply to flowline models and do not consider the embedding of an outlet glacier in its surrounding ice sheet. This could be an important factor contributing to the stability of outlet glaciers. As an outlet glacier retreats and thins, the surface slope increases in a lateral direction (i.e., towards the ice sheet), and more convergent flow should be expected, which could stabilize the glacier and delay further thinning and retreat (Howat et al., 2007).

Measurements from outside the fast flowing ice in West Greenland have also indicated a seasonality in ice flow (Lüthi et al., 2002; Zwally et al., 2002b; Joughin et al., 2008b; van de Wal et al., 2008). This seasonality is related to the seasonality of surface melt and was attributed to meltwater reaching the base of the ice (Zwally et al., 2002b). Das et al. (2008) observed the draining of a supraglacial lake to the bed and a corresponding local speed-up of the ice sheet. They concluded that the integrated effect of multiple lake drainages could explain the observed acceleration of the ice sheet. However, for several reasons, it is unlikely that increased melt directly leads to the observed speed-up of outlet glaciers.

First, the ice-stream acceleration was first observed near the termini of these glaciers and then propagated inland (Joughin et al., 2004; Howat et al., 2007). Second, at least in the case of the main channel of Jakobshavn Isbræ, there is no observed distinct seasonal reaction to the large variability of meltwater supply. Rather, the seasonality of ice flow is closely related to the position of the glacier terminus, which varies seasonally, but with a distinctly different phase than the meltwater production (Joughin et al., 2008c). Third, van de Wal et al. (2008) did find a large seasonal velocity signal that correlates well with meltwater production, but does not translate into a correlation between annual velocities and melt. In fact, the mean annual velocity of the ice sheet has a negative correlation with melt. This has also been reported for mountain glaciers (e.g., Truffer et al., 2005) and is presumably due to the theoretically expected increase in effectiveness of water drainage as discharge increases.

Reeh et al. (2001) and Joughin et al. (2008c) observed a correlation between sea-ice cover and the stability of glacier fronts. While it is perhaps difficult to imagine a sea-ice cover that provides substantial resistance to glacier flow, it is possible that icebergs held together by sea ice inhibit calving, which then leads to glacier advance because ice is not lost at the terminus (Joughin et al., 2008c). This can then have an influence on ice velocities by the increased resistance due to lateral shear, and perhaps by decreased effective pressures near the grounding line due to ice thickening. In practice, it is often difficult to tell whether the influence of sea ice is direct, or whether the sea ice reacts to the same forcing as the outlet glaciers, e.g., warmer ocean water.

That all large observed changes in ice-stream flow initially occurred near the ocean and then propagated inland, points to the importance of understanding ice-ocean interactions. Holland et al. (2008) reported that the acceleration of Jakobshavn Isbræ is temporally coincident with a pulse of deep warm ocean water that was observed along the west coast of Greenland. Typical fjord circulation patterns that bring in warm saline water at depth and return colder and fresher water at the surface could bring this pulse to the glacier front and contribute to the very high melting rates that were reported on Jakobshavn Isbræ before the disintegration of its floating tongue (Thomas et al., 2003). Howat et al. (2008) also noted a correlation between sea surface temperature and outlet glacier acceleration and retreat in southeastern Greenland.

In a different setting, Motyka et al. (2003) showed that melting of submerged ice at a tidewater glacier front could account for about half of the total ice flux into the ocean. The important factors for melting were the temperature of the deep fjord water and the strength of the fjord circulation. Motyka et al. (2003) hypothesized that the buoyant freshwater emerging at the glacier's base plays an important role in the fjord water circulation. This provides an interesting physical link between atmospheric temperatures, surface melt, and fjord circulation.

The fast flow of Jakobshavn Isbræ and other outlet glaciers can be explained as a consequence of subglacial topography. This is not, however, necessarily the case for all outlet glaciers. One that stands out in particular is the northeast Greenland ice stream (Fahnestock et al., 1993; Figure 2.11), which extends far into the ice sheet (Joughin et al., 2000). The fast flow of this ice stream might be controlled by geothermal heat. Fahnestock et al. (2001) used the spacing of deep radar layers to derive the geothermal heat flux under this ice stream and discovered unusually high melt rates.

Some of Greenland's glaciers also show surge-type behavior: long periods of quiescent flow interrupted by short periods of very rapid flow (Meier and Post, 1969). Glacier surges are a non-equilibrium phenomenon that is believed to be due to a positive feedback between basal water storage and basal motion (Kamb et al., 1985; Harrison and Post, 2003) that can last several months. Slower surges at polythermal glaciers might also be thermally controlled (Clarke et al., 1984). In Greenland, surging has been reported for the northeastern ice stream Storstrømmen (Reeh et al., 1994) and several glaciers in central East Greenland (Murray et al., 2002; Jiskoot et al., 2003). A short surge-like flow acceleration was also seen at Ryder Glacier (Joughin et al., 1996). It is important to remember the temporary nature of such changes in ice discharge. The mechanisms (and perhaps duration) are quite different from outlet glacier changes.

2.4. Total mass balance

2.4.1. Total net balance of the Greenland Ice Sheet

The total mass balance of an ice sheet is the time-varying rate of change in the mass of ice present in the sheet. It is the net result of the mass changes due to the addition and loss processes discussed in the previous sections. Owing to the seasonal nature of these processes, the mass of an ice sheet varies throughout the year, between years, and over many years as longer-term trends in climate become established. There are strong spatial gradients in mass balance that must be taken into account to determine variations in the total mass of the ice sheet.

For the Greenland Ice Sheet, which is subject to substantial surface melt and the resulting run-off of meltwater, mass exchange through the year has a large seasonal component at lower elevations caused by winter accumulation giving way to summer surface melt. These changes around the ice sheet margins are accompanied in outlet glaciers by changes in ice flow and iceberg production that may be linked in part to summer warming. At high latitudes and high elevations in the interior of the Greenland Ice Sheet, the relatively low input of snow and relatively small loss of snow to sublimation and wind transport lead to smaller amplitude annual variations in mass.

The time it takes for significant changes in accumulation or rates of ice loss around the margins of the Greenland Ice Sheet to impact on the ice flow in the bulk of the ice sheet's interior determines the response of the ice sheet to forcings on ice-age time scales. Changes in the flow of the Greenland Ice Sheet over millennia are driven by slowly changing temperature profiles and ice characteristics. The rate of these changes is controlled by thermal diffusion, slow flow of ice downwards and outwards as new snow is added to the ice sheet surface, and the slow evolution of the shape of the ice sheet as changing accumulation and melt patterns add and remove mass.

Prior to recent measurements of changes in the rates of outlet glacier flow and melt in Greenland, the ice sheet appeared to be roughly in balance. The mass turnover in the Greenland Ice Sheet was estimated at 500 Gt/y by Benson (1962) with about half the mass added through snowfall each year lost by surface melt and subsequent run-off, and the other half returned to the ocean through outlet glacier discharge leading to iceberg calving. But, as discussed in section 2.3, recent observations show quite large and rapid changes in surface melting and ice discharge. At the same time, different analyses of the total mass balance show the Greenland Ice Sheet overall to have lost mass since the early 1990s.

2.4.2. Techniques employed to observe or estimate total mass balance

Current techniques for estimating or measuring total mass balance include: (1) the mass budget approach, comparing total net snow accumulation with losses by ice discharge and meltwater run-off; (2) repeated altimetry, to estimate volume changes; and (3) tracking temporal changes in gravity, to infer mass changes.

The first two provide estimates of mass balance for regions included within survey boundaries, whereas the third provides an estimate for large regions with less distinct boundaries. All three techniques can be applied to the Greenland Ice Sheet as a whole, within limitations specific to each technique.

2.4.2.1. Mass budget approach

Changes in the Greenland Ice Sheet mass can be inferred from the difference between estimates of accumulation (i.e., snowfall) and mass loss through ablation (i.e., melt and subsequent run-off and sublimation) or ice discharge and calving. Each of these quantities involves a number of variables. Generally, snow accumulation is estimated from annual layering in ice cores, sometimes with interpolation between core sites using satellite microwave measurements or shallow radar sounding (e.g., Rotschky et al., 2004), or from atmospheric modeling (e.g., Box et al., 2004). Ice discharge is the product of ice flow velocity and ice thickness, with velocities measured *in situ* or remotely, preferably near the grounding line where velocity is almost depth independent. Ice thickness is measured by airborne radar, seismically, or from measured surface elevations assuming hydrostatic equilibrium for floating ice near grounding lines. Generally, meltwater run-off (large near the Greenland coast but small or zero elsewhere) is from model estimates calibrated against surface observations where available (e.g., Box et al., 2004; Hanna et al., 2005). Rignot and Kanagaratnam (2006) applied the mass balance approach around Greenland.

Mass budget calculations involve the often small difference of two large numbers, and small errors in either mass loss or gain terms can result in large errors in estimated total mass balance. These errors are difficult to assess for the Greenland Ice Sheet because of high temporal and spatial variability (see section 2.2). Although broad interferometric SAR (InSAR) coverage and progressively improving estimates of grounding line ice thickness have substantially improved ice discharge estimates, incomplete data coverage and uncertainty regarding velocity-depth relationships and meltwater run-off limit accuracy of total discharge estimates. The mass budget uncertainty for the Greenland Ice Sheet is a result of all these factors.

Moreover, accumulation estimates are based on data from the past few decades, and there are indications that snowfall in Greenland may be increasing with time (e.g., Hanna et al., 2005). Similarly, it is becoming clear that glacier velocities can change substantially over quite short time periods (see section 2.3). Both of these would add to estimated mass budget errors.

Thomas et al. (2000) gave a regional estimate for the mass budget for central Greenland, above 2000 m elevation, using discharge velocities from repeat GPS measurements made at stake markers 30 to 40 km apart along a traverse at 2000 to 3000 m elevations, ice thickness from radar sounding, and snow accumulation rates from shallow ice cores. Results show that between the mid-1970s and mid-1990s, central Greenland was in overall mass balance, but with regions of thickening, particularly in the southwest, and of thinning, particularly in the southeast.

A more complete picture, including near-coastal regions and individual drainage basins, comes from balancing net surface mass balance (section 2.2) against ice discharge rates derived from SAR-derived ice velocities and ice thicknesses from radar or surface elevations near grounding lines, where the ice is floating (Rignot and Kanagaratnam, 2006). SMB for this work was derived from SMB anomalies calculated from Hanna et al. (2005) scaled and applied to modulate a surface accumulation pattern from shallow ice cores combined with ECMWF (re)analyses (section 2.2).

Results show significant mass loss since the mid-1990s from most drainage basins, with very large losses after 2000 from Jakobshavn Isbræ, Kangerdlugssuaq Glacier and Helheim Glacier and from nearly all glaciers along the east coast south of Helheim (Joughin et al., 2004; Rignot and Kanagaratnam, 2006; Luckman et al., 2006; Howat et al., 2007, 2008). Velocity time series for these glaciers show up to a 100% increase after 2000, consistent with their negative mass balance.

2.4.2.2. Repeat altimetry

Rates of surface elevation change (δS/δt) reveal changes in ice sheet mass after correction for changes in depth / density profiles and bedrock elevation, or for hydrostatic equilibrium if the ice is floating. Satellite radar altimetry (SRALT) data have been widely used (e.g., Davis et al., 2000; Johannessen et al., 2005; Zwally et al., 2005), together with laser altimetry from airplanes (Krabill et al., 2000), and from NASA's ICESat (Zwally et al., 2002a; Thomas et al., 2006). Modeled corrections for isostatic changes in bedrock elevation (e.g., Peltier, 2004) are small (i.e., a few mm/y) but with errors comparable to the correction. Those for near-surface snow density changes (Arthern and Wingham, 1998; Li and Zwally, 2004; Helsen et al., 2008) are larger (i.e., a few cm/y) and also uncertain.

SRALT data provide the longest time series, with the first measurements in 1978. Resulting estimates of δS/δt can be misleading, however, because derived elevations refer to the radar reflection horizon. The depth of this horizon below the actual ice sheet surface is affected by near-surface characteristics, such as snow wetness. Consequently, changes in these characteristics can affect SRALT-derived estimates of δS/δt (Thomas et al., 2008), particularly in Greenland where the area affected by summer melting has increased substantially since the early 1990s. In addition, because of the broad radar beam, SRALT data are unreliable over the sloping and undulating surfaces near the coast, where rates of elevation change are greatest (e.g., Brenner et al., 2007).

Airborne Topographic Mapper (ATM) and satellite (ICESat) laser altimeters provide data that are easier to validate and interpret because footprints are small (i.e., about 1 m for airborne laser, and 60 m for ICESat) and there is negligible laser penetration into the ice. However, clouds limit data acquisition, and accuracy is affected by atmospheric conditions and particularly by laser pointing errors. Moreover, existing laser data are comparatively sparse, being limited by practical limitations on the number of ATM surveys and the ICESat orbit separation of up to 40 km in southern Greenland.

ATM elevation estimates are accurate to ~10 cm along survey tracks (Krabill et al., 2002) with similar errors estimated for ICESat surveys, but increasing over steeper slopes (Brenner et al., 2007). Because there are large gaps in both ICESat and airborne coverage, δS/δt values are generally supplemented by degree-day estimates of anomalous melting (Krabill et al., 2000, 2004). This increases overall errors and probably underestimates total losses, because it does not take into full account the dynamic thinning of outlet glaciers.

In summary, δS/δt errors cannot be quantified precisely for SRALT data, because of the broad radar beam and time-variable penetration. For laser data the difficulty in quantification comes because of sparse coverage. If the SRALT limitations are real, they are difficult, if not impossible, to resolve. Laser limitations result primarily from poor coverage, and can be resolved by increasing coverage to include all regions of rapid change around the Greenland coast. All altimetry mass balance estimates include additional uncertainties due to changes in the density profile of near-surface snow and (small) rates of basal uplift. If elevation changes are caused by recent changes in snowfall, the appropriate density (ρ) may be as low as 300 kg/m^3; for long-term changes, it may be as high as 900 kg/m^3. This is of most concern for small rates of surface elevation change, where the safest assumption is that density $\rho = 600 \pm 300$ kg/m^3, implying a ±50% mass balance uncertainty. Slobbe et al. (2009) used this conservatively large range for the entire ice sheet for an ICESat-only analysis, finding results consistent with other laser analyses.

Large and rapid changes, commonly found near the coast, are almost certainly caused by changes in melt rates or glacier dynamics, and the appropriate density of the lost or added volume approaches that of pure ice (ρ ~900 kg/m^3). Moreover, densification rates are sensitive to snow accumulation rates, temperature, and wetness, with warm conditions favoring more rapid densification (Arthern and Wingham, 1998; Li and Zwally, 2004; Helsen et al., 2008). Thus, the recent Greenland warming probably caused ice sheet surface lowering simply from this effect alone. Finally, the rate of basal uplift is inferred from models and has uncertain errors.

An early estimate of surface elevation change in southern Greenland using SRALT showed either little change or that much of the variability could be attributed to decadal changes in accumulation rates as measured in ice cores (Davis et al., 2000; McConnell et al., 2000). Other estimates of the Greenland Ice Sheet volume change from SRALT altimetry (Johannessen et al., 2005; Zwally et al., 2005) show thickening at higher elevations, particularly in the south, comparatively modest thinning nearer the coast, and a small positive mass balance overall.

However, as previously discussed, time-variable radar penetration and 'blurring' of near-coastal results by the wide radar beam make these results suspect (Thomas et al., 2008). Estimates derived by comparing ATM surveys 1993/1994 and 1998/1999 (Krabill et al., 2000) show substantial thinning near the coast and an overall mass loss of about 55 Gt/y between surveys. Comparison of elevations from these surveys with later values from ICESat at locations where flight tracks crossed ICESat orbits show that mass losses have increased to about 80 Gt/y between 1998/1999 and 2004 (Thomas et al., 2006). These laser-based estimates probably represent lower bounds because the sparse coverage 'misses' outlet glaciers that are thinning dynamically. Annual thinning rates of tens of meters on outlet glaciers in southeast Greenland are shown clearly from repeat stereo satellite coverage (Howat et al., 2007, 2008; Stearns and Hamilton, 2007).

2.4.2.3. Temporal variations in Earth's gravity field

Of all the techniques that are used to monitor the changing mass of the Greenland Ice Sheet, only gravity measurements are directly sensitive to changes in that mass; all other techniques convert changes in other quantities, such as melt, ice flux or surface elevation change, into changes in mass. Since 2002, the GRACE (Gravity Recovery and Climate Experiment) mission, joint between NASA and Deutsche Forschungsanstalt für Luft und Raumfahrt (DLR), has measured Earth's gravity field and its temporal variability. GRACE comprises two satellites, following each other in the same orbit 200 km apart, that carefully monitor the distance between themselves (Tapley et al., 2004a,b).

As the first satellite approaches a more massive object on the surface, it accelerates away from the trailing satellite, which then accelerates as it approaches the same object, closing the increased distance between the two orbiters. A detailed record of the changes in this inter-satellite distance over time provides a direct indication of changes in mass at the Earth's surface. Other accelerations are monitored with on-board accelerometers. GRACE has produced time series of these small changes in gravitational acceleration that start in 2002 and still continue (Tapley et al., 2004a,b). The spatial scale of the changes in gravity that GRACE can detect is determined by the height of the orbit, the spacing of the two satellites, and the accuracy with which the range rate between the two satellites can be determined.

After removing the effects of other loading, high latitude data, like those for Greenland, contain information on temporal changes in the mass distribution of the ice sheet and the underlying bedrock. Because of its high altitude, GRACE makes coarse-resolution measurements of the gravity field and its changes with time and the resulting mass balance estimates are also at coarse resolution – hundreds of kilometers. Despite this limitation, this technique has the advantage of covering the entire Greenland Ice Sheet, which is extremely difficult using other techniques. GRACE estimates include mass changes on the small ice caps and isolated glaciers that surround the Greenland Ice Sheet, regions not included in mass balance estimates from other techniques.

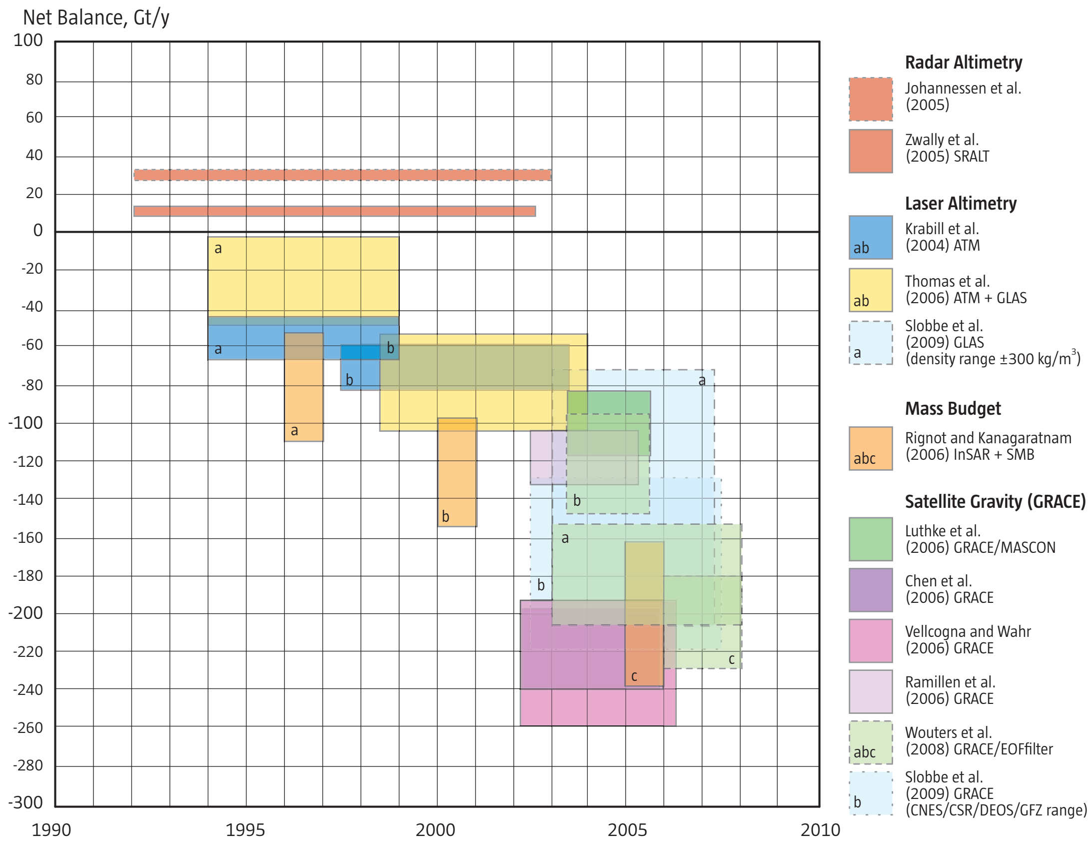
There are two different analysis schemes for GRACE data that have been applied to determining changes in the mass of the Greenland Ice Sheet. The first uses products derived by the GRACE mission that represent the global gravity field, expressed as a spherical harmonic expansion to a high order, differencing fields from each 30-day period to detect changes in ice sheet mass over time (Velicogna and Wahr, 2005, 2006; Chen et al., 2006). This technique requires convolving the harmonic expansion of the global gravity field with a filter representing either part or all of the Greenland Ice Sheet to isolate the changes in gravity that are due to changing ice mass. The gravity fields produced by the GRACE team have been updated since the time of publication of these papers; results may change with this revised input data. Ramillien et al. (2006) used a similar technique, but began with a different input dataset, which is a version of the Earth's geoid (i.e., another representation of the gravity field) derived for 10-day periods by a group at Centre National d'Etudes Spatiales (CNES). Slobbe et al. (2009) used a common analysis scheme on gravity fields produced by four different centers, and reported a significant range of resulting net balance rates.

Corrections applied in these analyses include compensation for changes in global patterns of water storage on land, ocean mass, and atmospheric mass. In addition, a correction must be made for longer-term changes in crustal mass distribution due to glacial isostatic adjustment (GIA) (crustal uplift that is a continued response to past deglaciation, or post-glacial rebound). Each of these analyses uses different input data sets and modeling strategies to make the corrections. Chen et al. (2006) used an optimized filtering technique to fit mass change time series on a 1 ° by 1 ° grid with high resolution aided by use of knowledge about where changes were occurring (i.e., in outlets near the ice margin identified by Rignot and Kanagaratnam, 2006). Ramillien et al. (2006) discussed the impacts of errors from each of these corrections. Barletta et al. (2008) conducted an extensive analysis of the impact of different factors on the GIA corrections to GRACE estimates in Greenland and Antarctica. They did make a somewhat simplified estimate of the rate of mass loss in Greenland from GRACE data to illustrate these impacts. This result is not included in Figure 2.13 or Table 2.3, but is similar to the others in the table.

The second type of GRACE analysis scheme uses the range rate measurements between the two satellites directly, modeling the local changes in mass required to produce the observed changes in these rates during different overpasses of Greenland (Luthcke et al., 2006). This work also makes corrections for GIA using published paleo ice sheet reconstructions, but not the smaller effects from atmospheric and oceanic mass variations.

Error sources can include measurement uncertainty, leakage of gravity change signal both from regions surrounding the Greenland Ice Sheet, and for some analyses, globally, and causes of gravity changes other than ice sheet changes. Of these, the most serious are the gravity changes associated with vertical bedrock motion (i.e., GIA). These are inferred from models of deglaciation history and its impact on vertical crustal motion, primarily post-glacial crustal uplift; the errors are also modeled and remain highly uncertain. Results from all gravity-based analyses show mass loss over their periods of observation, with rates ranging from -100 Gt/y to more than -200 Gt/y. The wide spread in results requires further investigation, but several studies using

Figure 2.13 *Results from the Recent Large Area Total Balance Measurements discussed in the text, placed into common units and displayed versus the time intervals of the observations. The heights of the boxes cover the published error bars or ranges in mass change rate over those intervals. Source: Mark Fahnestock, University of New Hampshire. See Table 2.3 for details.*



different analysis techniques show continued and accelerated change over the GRACE time period.

2.4.3. Synthesis and state-of-the-art

The results for each of the techniques are shown in Figure 2.13, plotted as boxes that encompass the length of the measurement period along the time axis, and the published range of the mass change rate over that interval. Data used for Figure 2.13 are summarized in Table 2.3.

With the exception of measurements from SRALT, estimates of the Greenland Ice Sheet total mass balance discussed here are negative, and are becoming more negative with time, with the caveat that many of these records are relatively short. The trend and timing are largely consistent with increased ice discharge from outlet glaciers around the edge of the Greenland Ice Sheet. Where these measurements do not agree within published errors, the sources of the discrepancies may be attributable to shortcomings in spatial sampling, complex sets of corrections, and input data that continue to evolve as satellite missions mature; limitations in measurement techniques are better characterized, and surface conditions on the Greenland Ice Sheet respond to a changing climate.

The complexity of measuring or estimating total balance has led to the development and exploitation of a number of measurement strategies. With respect to the Greenland Ice Sheet, these strategies have been developed and are maturing into a system that will one day be capable of tracking well constrained and internally consistent mass changes on a sub-annual basis. The extension of present measurement time series, combined with follow-on instruments and new airborne and satellite campaigns (e.g., Cryosat II), will help solve problems associated with limited coverage, slope and surface change induced error, and disparities in analysis of common data sets.

This represents a rapid evolution in capability that is presently matched by the rapid evolution of the Greenland Ice Sheet mass itself. One lesson that may be taken from the present state of knowledge is that multiple approaches and observations of mass input and mass loss processes have been required to make the progress that has been made to date.

Table 2.3 *Rate of total mass change in the Greenland Ice Sheet.*

Source	Technique	Time period	Rate of total mass change, Gt/y	Range (or stated error), ± Gt/y	Notes
	Radar Altimetry				Satellite radar altimetry is affected by surface slope near ice margins and changes in penetration
Zwally et al. (2005)	SRALT (ERS 1-2) + limited ATM	1992-002 (10.5 y)	11	± 3	
Johannessen et al. (2005)	SRALT (ERS 1-2) low elevations sparsely sampled	1992-2003 (11 y)	30	± 3	Densities of 0.9, 0.5, 0.4, 0.3, 0.3 Gt/km^3 assumed to convert area-weighted height change in elevation intervals (<1.5, 1.5-2, 2-2.5, 2.5-3, >3 km) into mass
	Laser Altimetry				
Krabill et al. (2004)	ATM (airborne laser altimetry)	1993/4-1998/9	-55	± 3	
		1997-2003	-73	± 11	
Thomas et al. (2006)	ATM	1993/4-1998/9	-4 to -50		High end of range likely due to limited coverage at low elevations
	ATM/ICESat	1998/9-2004	-57 to -105		High end of range likely due to limited coverage at low elevations
Slobbe et al. (2009)	ICESat (GLAS)	2003/2-2007/4	-139 ± 68		Used large density range (± 300 kg/m^3) for error bounds
	Mass Budget				
Rignot and Kanagaratnam (2006)	InSAR (ice motion) 1996 + balance anomaly estimate (Hanna et al., 2005)	1996	-83	± 28	
	InSAR (ice motion) 2000 balance anomaly estimate (Hanna et al., 2005)	2000	-127	± 28	
	InSAR (ice motion) 2005 + balance anomaly estimate (Hanna et al., 2005) (extrapolated)	2005	-205	± 38	
	Satellite Gravity (GRACE)				
Luthke et al. (2006)	GRACE MASCON – (change in gravitational acceleration on each orbital pass fit to forward model of mass history of local basins)	July 2003- July 2005	-101	± 16	Gain of 54 Gt/y above 2000 m and loss of 155 Gt/y below 2000 m
Chen et al. (2006)	GRACE – averaging filter applied to monthly spherical harmonic gravity field product	April 2002- Nov 2005	-219	± 21	
Velicogna and Wahr (2006)	GRACE – averaging filter applied to monthly spherical harmonic gravity field product	April 2002- April 2006	-227	± 33	
Ramillien et al. (2006)	GRACE – averaging filter applied to 10-day gravity solutions	July 2002- March 2005	-118	± 14	
Wouters et al. (2008)	GRACE EOF decomposition of monthly spherical harmonics iteratively fit with forward model of mass change of basins (based initially on R&K and iterated)	February 2003- January 2008	-179	± 25	
		July 2003- July 2005	-121	± 27	Same period as Luthke et al. (2006) Similar elevation dependence seen
		January 2006- January 2008	-204	± 25	
Slobbe et al. (2009)	GRACE using products from CNES (Centre National d'Etudes Spatiales), CSR (Center for Space Research, University of Texas-Austin), DEOS (Delft Institute of Earth Observation and Space Systems, Delft University of Technology, Delft), and GFZ (GeoForschungsZentrum, Helmholtz-Zentrum Potsdam)	2002/7- 2007/6	-128 to -218		Range in net balance is range of results using gravity products from four different centers, rather than estimated error

2.5. Summary and outlook

Compared with processes that take place at the base of the ice sheet, those that influence the surface mass balance are relatively well understood and considerably easier to observe. Despite this, published reconstructions of the individual components that make up the surface mass balance differ significantly. The differences can be as large as the inferred increase in mass loss due to changes in ice dynamics over the past decade. The uncertainties in the surface mass balance arise, largely, as a result of the paucity of relevant, spatially extensive *in situ* observations. To reduce uncertainties to a level useful for mass budget calculations and reliable surface mass balance predictions will require a concerted effort to collect more targeted *in situ* data, especially from the percolation and ablation zone. Improvements in process understanding and modeling of key processes such as refreezing, blowing snow and sub-grid scale effects will also be needed. Although there are differences between reconstructions, they all indicate a negative trend over the last ~15 years. This is largely due to an increase in run-off since about 1995, resulting in a marked reduction in the surface mass balance since then. Also of note is that 2007 had the lowest surface mass balance of any year over the 50-year record for which reliable estimates exist.

Much of the ice loss from the Greenland Ice Sheet occurs by ice discharge into the surrounding ocean. This has been estimated at 50% of the mass loss on average over the last ~50 years, but recently large variability in the flow of many large outlet glaciers has been documented. The temporal variability of Greenland's ice mass balance has a large contribution from variability of flow in its outlet glaciers. Observations during the past decade have shown that ice discharge can increase by a factor of two within a few years and, in some cases at least, that this can also be reversed. These changes are coincident with observations of warming in the ocean and atmosphere, as well as the disappearance of near-coastal sea ice. While the general mechanisms of fast flow are reasonably well understood, the necessary tools for predicting future behavior have not been developed. This is primarily due to a lack of understanding of the ice-ocean coupling and the role of surface water in determining the rates of ice flow.

The total mass balance of the ice sheet (i.e., the sum of surface mass balance and ice discharge) was exceptionally difficult to determine prior to the development of the present suite of observational and analysis techniques. While it is clear from the spread of published results that estimation remains a challenge for these techniques, it is important to note that nearly all the approaches show very similar trends, indicating clearly that the Greenland Ice Sheet is losing significant mass, and has been doing so at an accelerating rate over the past ten years. This conclusion is supported by all approaches sensitive to the ice margins: (1) the change in ice discharge from outlet glaciers combined with increasingly negative surface mass balance; (2) surface lowering in outlet glaciers measured directly by laser altimeters; and (3) the reduction in the mass of the ice sheet measured by the GRACE satellite gravity mission. This reduction of mass is consistent with what would be expected to happen in a warming Arctic and the diverse, independent suite of approaches used to document this change in total mass balance provides confidence in the result.

The challenge of more accurately determining the present and future rate of mass change of the Greenland Ice Sheet, by reducing the uncertainties inherent in the different approaches, will require improvements in measurements, more consistent and complete observational time series, better analysis schemes, and an improved understanding of the physical processes involved in recent rapid changes.

This chapter has attempted to show the state-of-the-art with respect to observations and understanding of the present-day, and recent past, behavior of the Greenland Ice Sheet. It is clear that, although the state-of-the-art has greatly advanced over the past decade or so, major gaps and challenges remain. During this period, unpredicted and striking variability in ice dynamics has been detected from satellite observations as well as a marked decrease in surface mass balance coincident with a decreasing trend in total mass balance. Serious limitations in the ability to model and monitor these processes, however, are also apparent from the disparity between published results.

Progress will need to come from many sources, including:

- Improving acquisition of more comprehensive and complete time series suitable for ice velocity determination, including seasonal variability. This will be predominantly from satellite and airborne remote sensing.
- Acquisition of improved measurements of surface elevation change by satellite and aircraft.
- Better measurement of ice thickness in rapidly flowing glaciers for improved ice discharge estimation.
- Improvements in analysis strategies and measurement capabilities for all processes which determine surface mass balance.
- More, and better, *in situ* measurements of refreezing and ablation.
- Better understanding of links between outlet glaciers, surface melt, and marine conditions.
- Perhaps, most importantly of all, the need to ensure that the time series of observations is secure into the future, and extended back in time with the aid of historic and paleo-proxy data.

3. Predictions and sensitivity

The Greenland Ice Sheet response to climate change is too complex to be predicted through simple theory or analytical equations. Surface mass balance depends on the interaction of weather systems with the ice sheet orographic influences and surface energy balance. Spatial and temporal patterns of ice sheet evolution therefore feed back onto the meteorological fields over the ice sheet. The regional- and synoptic-scale weather patterns that affect the ice sheet are also evolving as part of global-scale climate change and accompanying shifts in sea ice, ocean conditions, the hydrological cycle, and atmospheric dynamics. These systems are all interlinked and the interactions can be described only by numerical weather or climate models (Box 3.1).

Similarly, the evolution of the ice sheet depends on the climate fields that dictate the surface mass balance (see section 2.2), on the oceanic influences that affect the rates of iceberg calving and basal melting at the ice-ocean interface, and on the dynamic ice sheet processes that govern the flux of ice to the ice sheet margin. Numerical models have been developed to describe ice sheet dynamics – the gravity-driven deformation and sliding of ice. However, some processes are poorly understood and unresolved in the models, particularly those associated with basal sliding. Furthermore, surface mass balance and ocean-ice interactions need to be prescribed to simulate ice sheet response to climate change. Although efforts to couple ice sheet, atmosphere, and ocean models to describe the co-evolution of these systems are still in early stages, this is the best available route for physics-based forecasts of how the Greenland Ice Sheet will change in the coming decades and centuries.

Box 3.1 Numerical models of the oceans, atmosphere and ice sheets

Numerical models of the oceans, atmosphere, and ice sheets take similar approaches to simulating the large-scale evolution of each system. All are based on the principles of mass, momentum, and energy conservation, with these conservation equations governing mass and energy transfer within or between these systems. The models dissemble the oceans, atmosphere, and ice sheets into 3-D grid-cell networks and solve the governing equations to predict the system state in each grid cell. Neighboring cells interact to simulate the fluxes of water, air, ice, energy, and climatically active variables such as salt and water vapor.

The whole system is driven by boundary conditions such as global topography and bathymetry, the Earth's rate of rotation, and temporal and spatial patterns of solar input. The resulting fluid motions and energy exchanges provide a representation of the climate system. Integrating the system forward in time, system interactions can give internal weather/climate variability or can simulate climate system response to a change in boundary conditions, for example, increases in solar activity or greenhouse gas concentrations.

The resolution of the simulation is determined by the size of the 3-D grid cells used to discretize the oceans, atmosphere, and ice sheets. Although computational limitations are continually receding, simulating climate dynamics still taxes the world's most advanced computational facilities. This places limits on the available simulation capabilities. For century-scale climate change scenarios, typical grid resolutions are 1° latitude-longitude for the oceans, 2° to 3° for the atmosphere, and 0.2° (~20 km) for ice sheets.

3.1. Climate and meteorological modeling for Greenland

The response of the Greenland Ice Sheet to changing environmental conditions is the result of regional and hemispheric non-linear interactions between variable solar radiation, atmospheric composition, atmospheric and ocean circulation, ice sheet dynamics, cloudiness, precipitation, near-surface air temperature and the concentrations of aerosols suspended in the atmosphere and deposited onto the ice sheet surface.

Therefore, a key scientific focus of ongoing research is to clarify the robustness of Arctic climate change and the feedback processes responsible for the large climate variations and to quantify the extent to which Arctic climate change is due to regional or global processes. The interplay between observations and regional and global climate models is very important in this respect. Recent observations and climate modeling results have highlighted the Arctic as a region of particular vulnerability to global climate change. The Arctic has warmed by 0.46 °C since 1979, a value which is twice as large as the global warming due to the polar amplification. Superimposed on this trend, the Arctic climate system shows pronounced decadal-scale variability.

Temperature changes in the Arctic are linked to natural trends in modes of climate variability termed the Arctic Oscillation (AO) and the Pacific Decadal Oscillation (PDO) (see Serreze and Francis, 2006). These trends in the large-scale teleconnection pattern are connected with trends in Arctic cyclones. Intensification of storm tracks impacts on the global general circulation (in addition to the radiative greenhouse gas forcing) due to its nonlinear influence on meridional heat, humidity and momentum transports and forcing of quasi-stationary planetary waves. Global teleconnection patterns can be affected by regional feedbacks in the Arctic, as shown by Dethloff et al. (2006) as a result of albedo effects by sea ice and snow.

Chylek et al. (2006) analyzed Greenland temperature records to compare the current (1995-2005) warming period with the previous (1920-1930) Greenland warming period. Although there has been a big increase in temperature over the past decade (1995-2005), a similar increase and one which took place at a faster rate occurred during the early part of the 20th century (1920-1930). This was a period when carbon dioxide (CO_2) or other greenhouse gases could not have been a cause. The current (1995-2005) temperature increase seems to be within the natural variability of the Greenland climate. However, this does not exclude an anthropogenic cause for this recent period of warming.

Box and Cohen (2006) found systematic patterns of tropospheric and stratospheric temperature change around the Greenland Ice Sheet in radiosonde data spanning the 1964-2005 period. The recent 21-year period (1985-2005) is marked by statistically significant lower to mid-tropospheric warming and mid-stratospheric cooling. The 1995-2005 warming period has strongly influenced the longer term (1964-2005) temperature variations, resulting in a pattern of overall tropospheric warming in the +1.5 to +3.5 K range. Pre-1980s winter tropospheric cooling dominates the 1964-2005 period above the 500 hPa altitude.

Various researchers have analyzed atmospheric changes over the past 100 years and linked these data to changes in the Greenland Ice Sheet mass balance (see Chapter 2). Hanna and Cappelen (2003) analyzed data from eight stations in coastal southern Greenland for the period 1958-2001 and found a significant cooling over most of this period. The cooling is significantly, inversely correlated with a positive phase of the NAO over the past few decades. Hanna et al. (2001) showed that accumulation depends primarily on precipitation. Changes in the frequency and intensity of transitory cyclones are of paramount importance for precipitation trends.

Box et al. (2009) used meteorological station records and regional climate model output to develop a continuous 168-year (1840-2007) reconstruction of the Greenland Ice Sheet near-surface temperatures. They found that the annual whole-ice sheet 1919-1932 warming trend was 33% greater than the 1994-2007 warming. The recent warming was, however, stronger along western Greenland in autumn and along southern Greenland in winter. In contrast to the

1919-1932 warming, the 1994-2007 warming of Greenland has not surpassed the Northern Hemisphere anomaly. Greenland would need an additional 1.0 to 1.6 °C warming to become in phase with the hemispheric pattern. Two multi-decadal low temperature periods (1861-1919 and 1963-1984) in Greenland coincide with periods of major volcanic eruptions.

3.1.1. Regional climate modeling in Greenland

IPCC AR4 models incorporate many improvements compared to their predecessors, but there are still shortcomings (see Stroeve et al., 2007). While some studies suggest anthropogenic forcing may favor a positive Northern Annular Mode (NAM), there is evidence that climate models underestimate NAM-like variability (see Stenchikov et al., 2006). Maslanik et al. (2007) analyzed three types of atmospheric circulation pattern that appear most significant in terms of Arctic Basin winds and ice transport. The 'light ice' phases of these patterns include decreased mean sea level pressure (SLP) in the North Atlantic (an 'NAO-like' pattern resembling the positive phase of the NAO), a low pressure cell within the Arctic Basin (a 'central Arctic' pattern), and a dipole pattern of high pressure over the Canadian Arctic paired with low pressure over the Siberian Arctic. Winds and ice transport patterns that favor reduced ice cover in the western and central Arctic have continued since the late 1980s, but the AO index is not a reliable indicator of these patterns.

Box and Rinke (2003) ran the 1998 annual cycle and 1991 to 1998 summer simulations of the Greenland Ice Sheet surface climate with the HIRHAM regional climate model of the Arctic. They compared the model output with meteorological and energy balance observations from 15 Greenland Climate Network automatic weather stations and found that the model could reproduce the monthly average surface climate parameters, to a large extent within model and observational uncertainty. But, systematic model biases were identified.

Results from RCMs are sensitive to the choice of integration domain, the horizontal and vertical resolution and parameterizations, and all contribute to the uncertainties of RCM simulations. There is no exact choice of domain any more than there is for resolution or parameterizations, which means that domain must be treated as yet another element contributing to the uncertainty of RCM simulations. RCM simulations are also sensitive to lateral boundary conditions and to the surface conditions prescribed. The climate scenario projections come from interpolated atmosphere-ocean general circulation model (AOGCM) simulations and contribute to the overall uncertainties of RCM simulations and projections. The RCM results can vary depending on the AOGCMs used, in terms of large-scale atmospheric circulation patterns and their teleconnection. RCM results are often based on only one specific AOGCM and are for just one selected scenario. A multi-model ensemble approach could increase the confidence of future projections of extreme climate. Accurate simulation of surface mass accumulation over the ice sheets requires a spatial resolution which is currently not available from AOGCM simulations. Therefore high-resolution RCMs (i.e., PMM5 and MAR) have been run for shorter time slices driven by AOGCM boundary conditions and combined with an ice sheet mass balance model as described by Gregory and Huybrechts (2006) and Fettweis et al. (2008).

The following model improvements are recommended for any regional climate simulation of Greenland:

- Use of an accurate ice sheet topographic dataset.
- Accurate albedo parameterization.
- More realistic surface layer model, including thermal conductivity for snow / firn.
- Accurate planetary boundary layer momentum exchange.

These recommendations would improve the simulation of regional effects of Greenland's climate due to finer-resolved orography and land-sea contrasts, better-resolved non-linear interactions between the large-scale and mesoscales, better simulation of hydrodynamic instabilities and synoptic cyclones, and better description of hydrological and precipitation processes. Better representation of the northward development of the storm track, and invasion of the Arctic Basin by cyclonic systems from the lee of Greenland, which varies greatly with the NAO, would have a major influence on the Arctic sea-ice movement and melting in a coupled regional climate system. At higher resolution, the air-sea interaction changes, the air-sea heat flux is enhanced and localized in the lee of Greenland, and the wind stress increases.

Dethloff et al. (2004) quantified the influence of Greenland's topographic effect on the atmospheric winter and summer circulation in the Arctic with the high-resolution regional atmospheric model HIRHAM4. Removing the influence of Greenland from the model had a pronounced effect on the atmospheric winter circulation of the Arctic. There was a northeastward shift in the storm tracks over the North Atlantic and an increase in synoptic activity over Alaska. The pronounced 'precipitation minus evaporation' changes connected with shifts in the synoptic storm tracks during winter would have important consequences for the atmospheric freshwater input into the Arctic Ocean and the Nordic Sea with the potential to cause variability in the Arctic Ocean dynamics.

As well as regional impacts, Greenland exerts strong hemispheric influences. Jung and Rhines (2007) showed that stationary waves in the Northern Hemisphere are influenced by Greenland, as argued from the extreme model simulations with and without Greenland by Junge et al. (2005). They also showed how the ability of the models to simulate the energy cascade from eddies to stationary waves can be simulated with increasing realism with increasing resolution (Figure 3.1). There is an interaction between the upper-tropospheric jet stream waveguide and synoptic systems in this region. Greenland lies north of the core of the tropospheric westerly winds. Strong standing waves, which extend well into the stratosphere, produce a trough / ridge system with a jet stream lying close to Greenland, a mean Icelandic low in its wake, and a storm track that interacts strongly with topography.

Dethloff et al. (2002) simulated the accumulation defined as 'Precipitation minus Evaporation' over Greenland with the high-resolution limited-area regional climate model HIRHAM4 applied over an Arctic integration domain. They compared this with a revised estimate of annual accumulation across Greenland taking into account information from a new set of ice core analyses, based on surface sample collections from the North Greenland Traverse. The minima are connected with a prevailing and blocking high pressure over the Greenland Ice Sheet and katabatic wind systems preventing humidity transports to central Greenland. Maxima of precipitation and accumulation occur at the southwestern and southeastern coasts of Greenland and are connected with cyclonic activity and the main storm tracks around Greenland. The central region of the Greenland Ice Sheet acts as a barrier blocking moving weather systems and prohibits cyclones moving from west to east across the region, thus preventing moisture transport.

Dorn et al. (2003) analyzed regional patterns of Arctic winter climate changes resulting from regime changes of the NAO using a regional atmospheric climate model. The model was driven using data for positive and negative phases of the NAO from a control simulation as well as from a time-dependent greenhouse gas and aerosol scenario simulation. In some regions, the climate changes associated with the NAO were clearly stronger than those attributed to enhanced greenhouse gases and aerosols, indicating that projected global changes in atmospheric composition and internal circulation are competing with each other.

Knowledge of the future NAO trend at decadal and longer time scales appears to be important in terms of a regional assessment of climate scenarios in the Arctic. These will depend on interactions at global spatial scales, as shown by Brand et al. (2008). They used the AOGCM ECHO-GiSP with simplified stratospheric chemistry in a conceptual study on the impact of interactive stratospheric ozone chemistry on the tropospheric circulation. The results showed a clear sensitivity of the tropospheric circulation dynamics to the stratospheric chemistry. With interactive stratospheric chemistry enabled, the model tends to the negative phase of the AO mode. The AO may be considered as the full hemispheric equivalent of the NAO, which is often con-

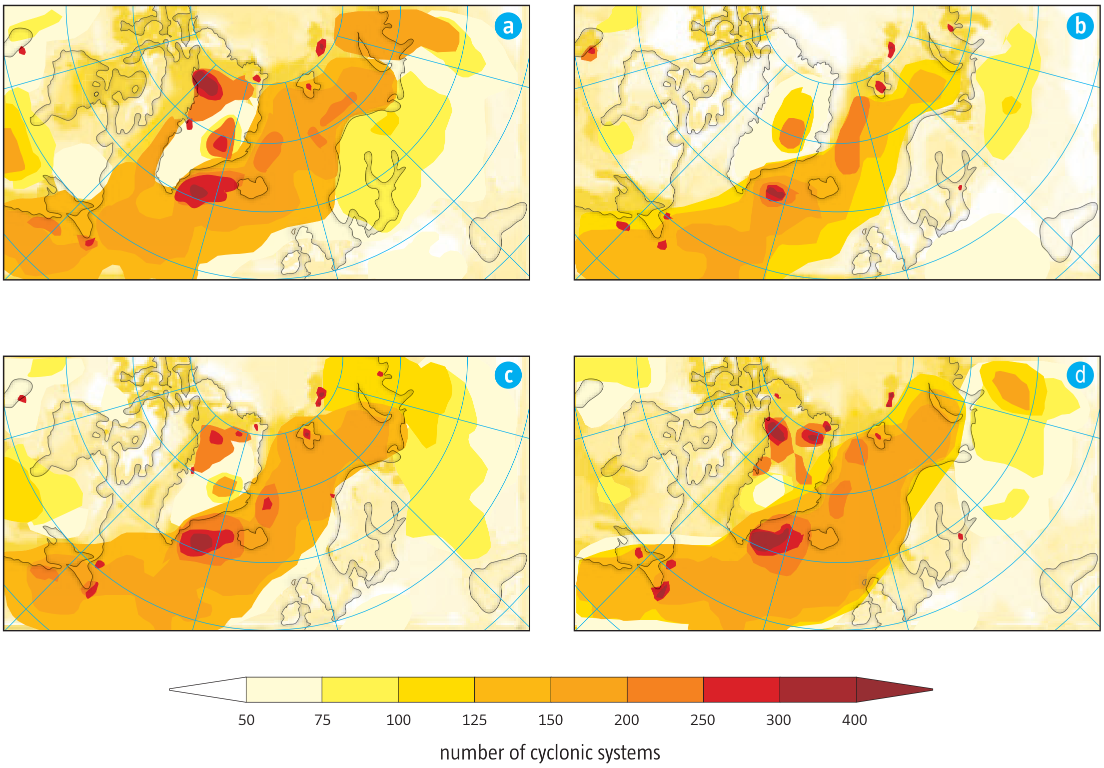


Figure 3.1 *Number of cyclonic systems from re-analysis data and models with various model resolutions. (a) ERA-40 re-analysis data, (b) T95L60 (grid interval of about 200 km), (c) T159L60 (grid interval of 120 km), and (d) T255L40 (grid interval of 80 km). Results are based on all winters (DJFM) for the period 1982-2001. Results are based on only those systems that persisted longer than 12 h. Note the improving agreement with ERA-40 data as resolution exceeds that for T95. Cyclonic activity peaks both on the eastern and northwestern sides of Greenland. Source: Jung and Rhines (2007).*

sidered a more relevant phenomenon when dealing with the free (upper) atmosphere.

Water vapor is the most abundant and most radiatively important greenhouse gas. It can transport large amounts of latent heat, strongly influencing the dynamics of the atmosphere, and is an important link between the various components of the hydrological cycle. Rinke et al. (2008) presented regional Arctic Total Water Vapor (TWV) trends for 1958-2001. A widespread increase in TWV was obvious in all four seasons, with the greatest increase in TWV during the summer. The positive TWV trend accelerated over the Arctic and Atlantic regions between 1988 and 2001.

Hanna et al. (2006) calculated annual and monthly snow accumulation for the Greenland Ice Sheet from the 40-year ECMWF Re-Analysis (ERA-40). Results showed the central and northern plateaus to be too 'dry' by 10% to 30%, whereas some parts of the interior south seemed, conversely, to be too 'wet'. This implied significant deficiencies in the version of the ECMWF model used to produce ERA-40, partly due to relatively low resolution (1.125° latitude × 1.125° longitude) (see also section 2.2).

Hanna et al. (2008) attributed significant increases in Greenland summer temperatures and Greenland Ice Sheet melt and run-off since 1990 to global warming. The significant increases in observed Greenland margin summer temperatures and modeled run-off, the new 2003 and 2005 observed temperature and snowmelt records, and the highly significant correlation between recent Greenland and Northern Hemisphere temperatures since the early 1990s, all suggest that a response of the Greenland Ice Sheet to global warming may well be emerging, and now less strongly linked with modes of natural variability such as the NAO (see also section 2.2).

This more recent signal can be set against a background of natural variability, including regional changes in atmospheric circulation related to the NAO. This is in agreement with a study by Bromwich and Wang (2008) who analyzed the spatial and temporal behavior of the atmospheric general circulation in both polar regions based on the ECMWF 40-year re-analysis, updated through the end of 2005 by the addition of ECMWF operational analyses and found that the NAO has been a key modulator of the atmospheric circulation in the North Atlantic sector, especially in winter, and has been the dominant control on moisture transport into the Arctic Basin. Greenland climate and mass balance, and ocean warming have been suggested to be linked to NAO as well (Holland et al., 2008) and potentially be drivers for recent rapid changes of Greenland outlet ice streams (see also section 2.3).

Saha et al. (2006b) investigated the effects of different land surface schemes on the large-scale circulation in the Arctic by coupling the regional climate model HIRHAM4 with two different land surface models (ECHAM4-LSM and NCAR-LSM). Mean sea level pressure was increased over most of the domain during summer with maxima centered over the central Arctic and north Canada, whereas during winter the increased maxima occurred over the Kara Sea and the Laptev Sea. Although regional circulation changes were obvious over land areas, the main changes occurred over the Arctic Ocean. This shows that the local to regional changes in land surface processes not only have a direct impact on the local to regional circulation, but that this is also the case in remote regions. The mechanisms involved are thought to occur mainly via changes in surface albedo and surface turbulent fluxes. Indirectly, surface albedo changes can also modify the large-scale circulation (Dethloff et al., 2006).

Modeling sea ice in a realistic manner is still a great challenge, particularly with respect to the minimum ice extent at the end of summer. Dorn et al. (2007, 2008, 2009) investigated modified descriptions of ice growth, snow and ice albedo, and snow cover on ice used in the coupled regional atmosphere-ocean-ice model HIRHAM-NAOSIM. A series of sensitivity experiments were performed in order to assess the need for more sophisticated parameterizations of these processes in coupled regional and global models. These showed that the simulation of Arctic summer sea ice is very sensitive to the parameterization of snow and ice albedo as well as to the treatment of ice growth. The parameterization of the snow cover fraction on ice plays an important role in the onset of summer ice melt. This has a crucial impact on summer ice decay when more sophisticated schemes for ice growth and ice albedo are used. Dorn and co-workers showed that when using a harmonized combination of more sophisticated parameterizations, simulation of the summer minimum in ice extent can be considerably improved due to the more realistic representation of interactions between the atmosphere and sea ice influencing Greenland and the Arctic as a whole.

Saha et al. (2006a) examined the possible changes in future winter temperature and precipitation extremes in the Arctic using the regional climate model HIRHAM4. Under the B2 emissions scenario conditions, the frequency and intensity of future (2037 to 2051) extremes changed significantly compared to present-day (1981 to 1995) extremes. Extreme precipitations intensified and the number of extreme events changed significantly over East Siberia and the Barents Sea. Extreme warm and extreme cold temperatures became warmer with maxima over the Barents Sea and central Eurasia. Changes in the mean climate and its variability were modulating future winter extreme events. The model projected very significant changes in the frequency and intensity of extreme temperature and precipitation in the Arctic in three to five decades time due to increased greenhouse gases.

Generally, the Arctic is projected to be warmer and wetter according to the recent IPCC AR4 GCM studies (Kattsov et al., 2007). Using an RCM, Rinke and Dethloff (2008) estimated future changes over the Arctic by the end of the 21st century, based on the SRES A1B emissions scenario. They showed that the RCM projections include distinct regional patterns. Over the Arctic Ocean these are mainly associated with sea-ice retreat and over land are mainly associated with surface characteristics, such as orography and the agreement with the large-scale temperature and precipitation projections within the driving GCM is striking. Regional detail is evident along the North Atlantic storm track and, during summer, close to orographic obstacles. This is attributed to the higher resolution of the RCM.

Generally, GCMs project a decrease in SLP over the Arctic associated with a poleward expansion in the Hadley cell circulation, a poleward shift in the mid-latitude storm tracks with a corresponding increase in cyclonic circulation patterns over the Arctic, and a northeastward shift in the Icelandic low. Thus, during winter the SLP is projected to decrease significantly over the circum-Arctic domain. In the study by Dethloff et al. (2008), the maximum SLP reduction (> 6 hPa) is projected to occur with increasing frequency over the Barents Sea, the Kara Sea, and the Bering Strait associated with sea-ice reduction and the poleward shift in storm tracks. The Siberian High is projected to weaken in association with the warmer lower troposphere. This projection is in line with GCM projections and with the declining trend in the intensity of the Siberian High observed after 1977. In summer, the changes in SLP are projected to be more moderate. The regional pattern in SLP is characterized by a slight increase in the Beaufort Sea, Chukchi Sea and Bering Strait, with a significant decrease (up to 4 hPa) over the northern North Atlantic and the Barents Sea associated with sea-ice retreat and the related surface warming.

The RCM results can vary depending on the driving AOGCMs and are sensitive in terms of large-scale atmospheric circulation patterns and their teleconnection. Handorf and Dethloff (2009) analyzed the low-frequency variability of the mid-tropospheric atmospheric flow of the Northern Hemisphere during winter in terms of teleconnection patterns and atmospheric flow regimes under future climate projections. To assess the realism of state-of-the-art AOGCMs, multi-model simulations for present-day conditions, performed for the Fourth Assessment Report of the Intergovernmental Panel on Climate Change were analyzed. A comparison with observations revealed that state-of-the-art AOGCMs are able to describe low-frequency variability in terms of teleconnections and flow regimes realistically. Their analyses of simulations for future climate scenarios revealed changes in the strengths of the centers of action and two new climate regimes, as well as slight changes in the structure of some existing regimes.

Krinner et al. (2009) showed that compared to simulations with a simple fixed and constant sea-ice thickness of 3 m, a variable sea-ice thickness will induce increased heat flux from the underlying ocean towards the surface-atmosphere interface. The change is particularly strong where and when the sea ice is thin but dense and the snow cover is weak, i.e., in the marginal ice areas and at the start and end of the cold season. Under a changing climate, the prevalence of these conditions shifts in time (i.e., seasonally) and space. Remarkably, therefore, the maximum impact of a prescribed variable sea-ice thickness, as opposed to a fixed constant value of 3 m, occurs in the marginal Arctic seas in winter under present-day climatic conditions, but in spring in the central Arctic at the end of the 21st century (supposing that the climate change simulated by the climate models used here and the underlying SRES-A1B greenhouse gas emissions scenario are not completely unrealistic).

As a result, not only is the simulated polar climate at a given period significantly affected by the way sea-ice thickness is prescribed, but this is also the case for the simulated future polar climate. Climate change studies using atmosphere-only models, typically regional climate model studies carried out with the aim of regionalizing global-scale coupled model scenario runs using prescribed reconstructed sea-ice extents, should thus be carefully designed in this respect.

Atmospheric modeling over complex terrain is a great challenge. The major reason for this is the difficulty represented by the topographical relief. If not properly resolved, the atmospheric circulation will not, for an otherwise perfect model, be able to depict the real local circulation patterns. This, in turn, will lead to systematic deficiencies in the models' ability to capture regional / local climatological features.

The most obvious example is that of topographically lifted precipitation that causes enhanced precipitation on the upstream slopes of a mountain, while producing less precipitation on the lee side downstream. Only when the appropriate mountain ridges are captured by the resolution, will the precipitation patterns be accurately simulated. The importance of this effect at larger scales is not well understood.

3.1.2. Model limitations and challenges

In order to capture even the large-scale snow accumulation rates, it is essential for the atmospheric models to include a good representation of the mountains around the Greenland Ice Sheet. The detailed deposition of moisture in the mountains as well as over the ablation zone of the ice sheet will influence the amount of moisture transported further into the interior of the ice sheet. It is also of great importance to use an accurate elevation model for the ice sheet itself. Box and Rinke (2003) looked at how the ice sheet topography used in the standard version of the HIRHAM regional climate model compared to a more recent digital elevation model (e.g., Bamber et al., 2001), and found substantial differences. They considered these to be responsible for several degrees of systematic temperature bias, particularly in the ablation zone.

Figure 3.2 illustrates how the topography of Greenland is depicted with increasing resolution (based on Bamber et al., 2001). At the coarsest grid resolution (~150 km), which is typical for most advanced state-of-the-art global circulation models, the resolution does not allow for discrimination between the outermost part of the ice sheet and the coastal mountains. This basically implies that the atmosphere 'sees' all of Greenland as one big lump of ice. Increasing the grid resolution to 25 km improves the situation, but the gradients over the ice margins are not resolved even at this scale. Typically, the incremental rise in topography within the ablation zone is 300 to 500 m for a 25 km grid, which is potentially in conflict with the formulation of the vertical structure of the atmospheric model and clearly introduces systematic surface temperature errors that may exceed 3 °C. At a grid resolution of 12 km, this feature is almost resolved. But it is not until a resolution of about 5 km is applied that the surface temperature error no longer introduces serious inconsistencies, in cases where the modeled temperature field is used without any adjustment.

At present, it is still speculative as to whether a grid resolution of 5 km will also improve the precipitation climatology. Figure 3.3, however, compares the precipitation patterns from three simulations with the regional climate model HIRHAM with increasing resolution (Stendel et al., 2008; Philippe Lucas-Picher, Danish Meteorological Institute, pers. comm. 2008). The three HIRHAM realizations have a grid distance of ~80 km, 25 km and 5 km, respectively. The large-scale patterns are similar, but whether the precipitation is falling in the coastal margins, in the ablation zone or even in the accumulation zone does not appear to be resolved unless using a fine grid resolution.

In order to provide realistic driving conditions for an ice sheet model, a necessary requirement for the atmospheric model must, therefore, be that it resolves the features that are of most relevance in controlling the mass balance of the ice sheet. If this cannot be provided directly, additional adjustments to the data are necessary. This, in turn, raises questions about internal consistency of the energy and water cycle balances.

Representing the detailed regional conditions of Greenland in a climate model is also affected by the ability of the model to simulate realistically the energy and moisture cycles in the surrounding seas. Clearly, part of the moisture supply for the precipitation systems influencing the ice sheet originates from the regional seas surrounding Greenland. Because the oceanic circulation around Greenland is complex and poorly resolved by

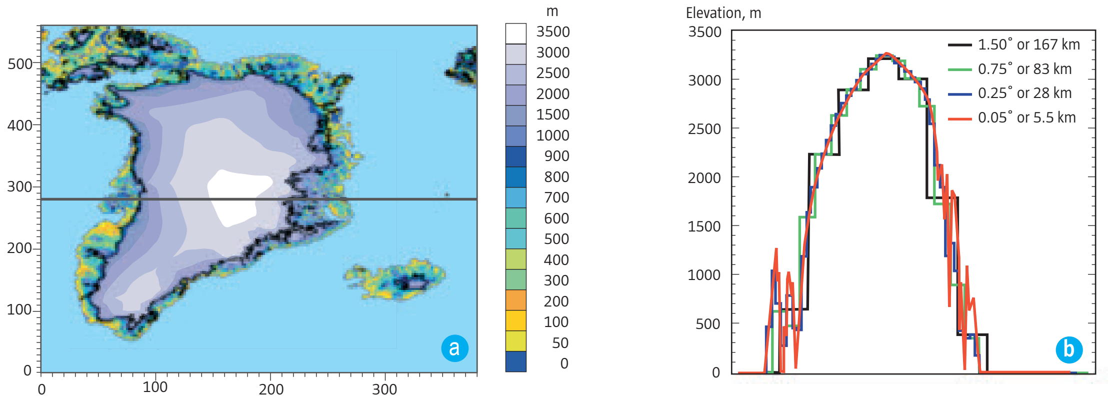


Figure 3.2 *(a) Topography of Greenland. The ice sheet extent indicated by black contour, (b) elevation as seen by models along W-E transect (indicated on the adjacent map) at GCM scale [***black** *line] (~150 km), high-resolution GCM scale [***green** *line] (~75 km), high-resolution RCM [***blue** *line] (~25 km), and very high resolution [***red** *line] (~5 km). Source: Courtesy of Philippe Lucas-Picher, Danish Meteorological Institute.*

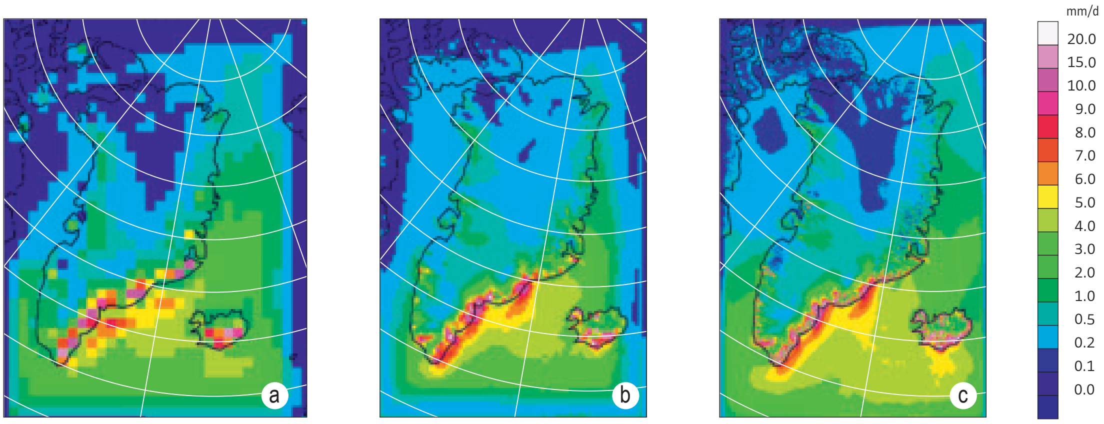


Figure 3.3 *Resolution dependency of simulated winter precipitation over Greenland (mm/d). (a) RCM resolution (~80 km), (b) RCM resolution (~25 km), (c) RCM very high resolution (~5 km). Source: Courtesy of Philippe Lucas-Picher, Danish Meteorological Institute.*

coarse-resolution coupled climate models, it is not surprising that detailed sea-ice conditions are often poorly represented.

At best, only large-scale sea-ice conditions can be depicted. Stendel et al. (2008) gave an example of a climate change downscaling experiment, where the driving GCM does have a relatively good representation of present-day sea-ice conditions, both with respect to seasonal variability and to interannual variability. However, details are still not matching observed conditions to a level providing confidence that changes are indeed results of a realistic timing of changes in circulation and moisture support.

When ice sheet models are forced with atmospheric model output from GCMs, it is always an issue as to how realistic the simulated sea-ice temperatures and sea-surface temperatures actually are.

The IPCC concluded in its Fourth Assessment Report (IPCC, 2007) that mean temperature within the Arctic is likely to increase at a higher rate than mean global temperature, confirming results from previous IPCC assessments and the Arctic Climate Impact Assessment (ACIA, 2005). For the IPCC SRES A1B emissions scenario, the annual mean Arctic temperature increase by the end of the 21st century is projected to be about twice that of the global mean increase (5-7 °C *vs* 2.5-3.5 °C) (IPCC, 2007; Christensen et al., 2007). Figure 3.4 shows an extract from an evaluation of 21 model simulations of global change under the A1B emissions scenario (Christensen et al., 2007), highlighting three models, the NCAR PCM, GFDL CM2.0, and MPI ECHAM5 as well as the 21-model mean. It is clear that the three individual models qualitatively show the same climate change response, but that the magnitude of the change differs by several degrees with the PCM model showing the lowest and the ECHAM5 model the highest degree of warming, while the GFDL CM2.0 model is close to the ensemble mean.

It is interesting to note, however, that the projected climate signals are to some degree caused by quite different mechanisms. Figure 3.5 shows an extract of an analysis of the performance of 14 model simulations for the period 1958-2000 (Walsh et al., 2008).

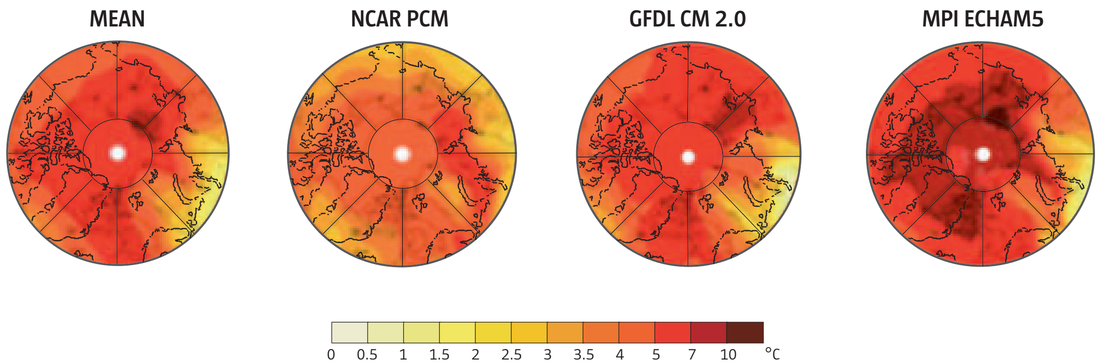


Figure 3.4 *Maps of composite (based on 21 models) and three individual model simulated annual mean temperature changes (2080-2099 vs 1980-1999) for the IPCC A1B SRES scenario. Source: IPCC (2007).*

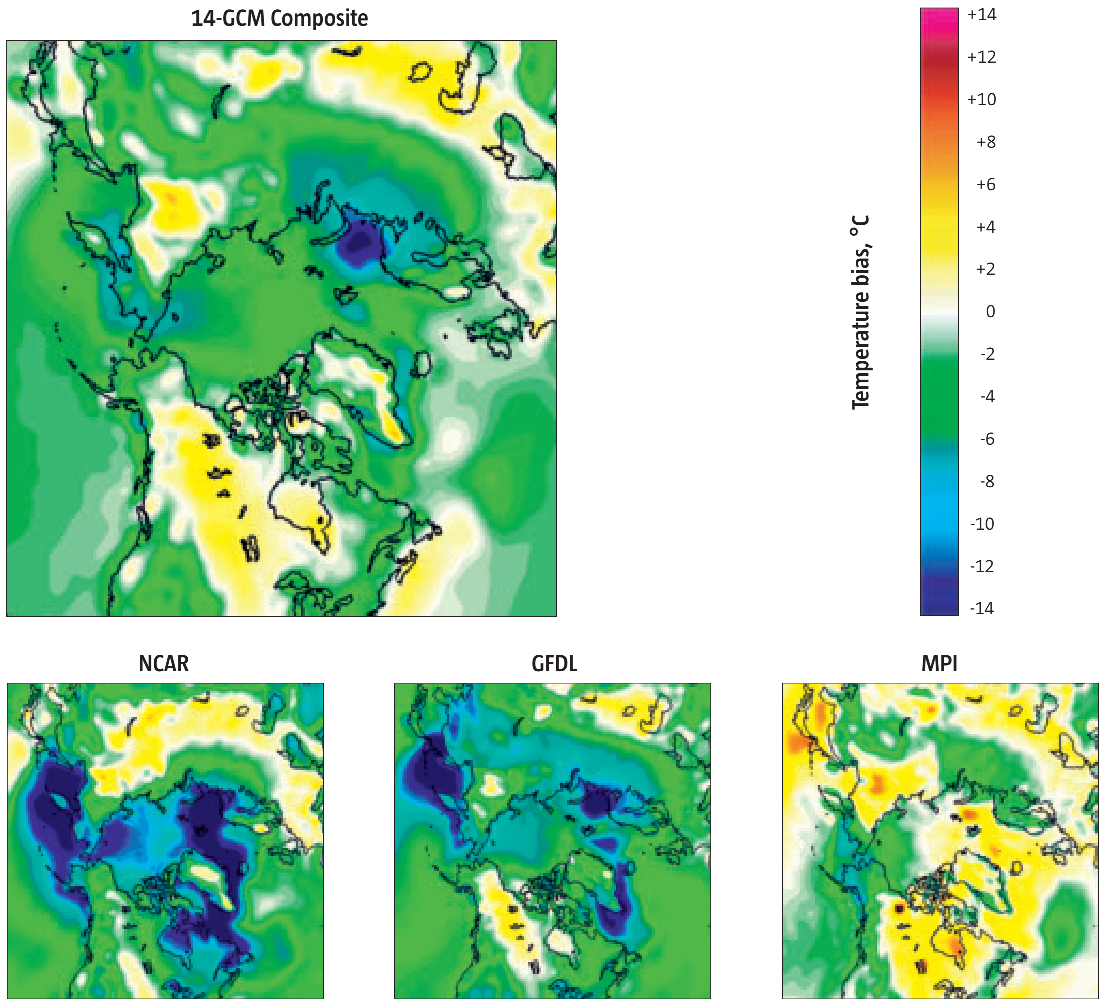


Figure 3.5 *Maps of composite (based on 14 models) and three individual model temperature biases for winter (1958-2000). Source: Walsh et al. (2008).*

Again, the ensemble mean behavior is shown with the same three individual models. A common feature for most of the models, reflected by the ensemble mean, is a clear cold bias in the Barents Sea due to a tendency to simulate too much sea ice, with the MPI model a clear exception. The implication is that only the MPI model simulates reasonably well sea-ice coverage in this region. At the same time, the greatest warming by the end of the study period is simulated exactly over this region in the ensemble mean as well as by the individual models.

In the NCAR and GFDL models, this partly reflects the bias in present-day sea-ice conditions, while this apparently cannot be the case in the MPI model because the present-day sea ice appears to be captured with some realism. It should also be noted that, in general, the greatest warming occurs in the area with too much ice (strong cold bias) under current conditions, and in the NCAR model in particular, even though here winter data are compared with the annual mean.

This section has shown that, to varying extents, results at the regional scale are clearly subject to systematic errors in present-day simulations. Using an ensemble of models masks this deficiency. The role of the ocean as a driver of the climate in the Arctic is obvious, and one should be concerned about interpreting a climate change signal, when most of the signal is apparently due to systematic error in simulating present-day conditions.

As the models discussed above have indicated, maps of warming must be carefully analyzed and, without further analysis, data cannot be used in a region with non-linear feedbacks such as the presence and absence of sea ice. Thus, it is also likely that the simulated warming and precipitation change signals are strongly influenced by these systematic errors, not only in the vicinity of the imperfections themselves, but also farther away due to non-linear interactions in the atmospheric and oceanic circulations.

3.2. Modeling Greenland Ice Sheet dynamics

3.2.1. Objectives of numerical ice sheet modeling

The future evolution of the Greenland Ice Sheet depends on the extent of climate change over the ice sheet, in particular the specific impacts on surface mass balance as well as past climate forcing. As discussed in sections 2.3 and 2.4, overall ice sheet mass balance depends strongly on the rate of ice discharge. In particular, dynamical mass loss is governed by the flux of ice to the ablation area. This includes both terrestrial margins, where ice is removed through melting and run-off, and marine margins, where ice loss occurs through iceberg calving and melting of ice that is in contact with the ocean. High ice velocities lead to greater transport of ice to the ice sheet margins and higher rates of mass loss through each of these processes.

Ice sheet dynamics are difficult to predict or extrapolate from present-day observations; ice sheet velocities do not remain constant on decadal and century timescales. There are unanswered questions as to how ice flux will respond to changes in ice sheet geometry, oceanic boundary conditions, and surface meltwater reaching the bed. All of these changes are expected in the coming decades and centuries.

Numerical ice sheet models have been designed to describe first order ice sheet physics and to provide prognostic solutions for ice sheet evolution in response to such changes. Numerical models are required because the flux of ice at a particular point depends on a complex array of physical parameters, including ice sheet thickness, surface and bed slopes, ice temperature (which influences the effective viscosity of the ice), ice fabric, valley walls, regional ice flow, and basal friction (a function of bed geology, basal ice temperature, and subglacial hydrological conditions). By simulating these processes and features of the ice sheet system, physically-based estimates of ice flux and ice sheet response to climate change are possible.

3.2.2. The physical basis of ice sheet models

Detailed descriptions of ice sheet physics are provided by van der Veen (1999) and Paterson (1994). Box 3.2 provides a simplified overview of the theoretical basis of ice sheet models. The ice sheet is discretized into a 3-D array of grid cells and the equations that govern ice thickness, velocity, and temperature evolution are solved in each grid cell. Atmospheric and oceanic forcing of the ice sheet are introduced as boundary conditions.

The Greenland Ice Sheet is still adjusting to the last cycle of glaciation and deglaciation, particularly with respect to ice temperatures at depth, where most of the ice deformation occurs. Simulations of future evolution of the ice sheet therefore require a ‘spin up’ simulation that takes Greenland through one or more glacial cycles (Huybrechts et al., 1991; Letréguilly et al., 1991). The climate forcing for these simulations is taken from Greenland ice core records. Climate forecasts are then imposed for studies of future ice sheet changes, generally as perturbations from present-day climatology.

When considering longer time scales, adjustment of the Earth’s crust to ice loading and unloading must be included in the model as this represents a potentially important feedback through the coupling between surface mass balance and surface elevation (e.g., Oerlemans, 1980). Different models have been adopted for estimating isostatic adjustment, with for example, Marshall et al. (2000) assuming a local, damped return to isostatic equilibrium, while Le Meur and Huybrechts (1998) incorporated a self-gravitating spherical visco-elastic Earth model.

For whole-ice-sheet simulations over one or more glacial cycles the horizontal grid spacing is typically 20 km or more, with some 30 layers in the vertical (e.g., Huybrechts, 2002). Grid sizes of 5 to 10 km are tractable, but each factor of two reduction in grid size increases the computational demand by an order of magnitude. There are also theoretical and pragmatic limits to resolution. The shallow-ice-approximation that is used in most whole-ice-sheet models is based on an assumption that horizontal grid dimensions are greater

Box 3.2 Theoretical basis of ice sheet models

Three-dimensional models applied to the Greenland Ice Sheet to date all make use of the shallow-ice approximation, in which the gravitational driving stress is locally balanced by drag at the glacier base (Nye, 1957; Hutter, 1983). The gravitational driving stress is τd (z) = ρ g(hS – z) ∇hS, where ρ is the ice density, g is gravitational acceleration, hS is the glacier surface elevation, and ∇hS is the surface slope. At the glacier bed, τd (hb) = ρ gH ∇hS, for ice thickness H. Glen's flow law relates ice deformation rates to the stress field in the ice. In the shallow-ice approximation, the ice velocity (u) associated with vertical shear deformation (d), i.e., ud, follows ud ∝ ∇hS n H n+1, where n = 3 is the exponent in Glen's flow law.

Ice deformation rates are sensitive to ice temperature, with the effective viscosity of ice varying by a factor of about 1000 over the range of temperatures found in the Greenland Ice Sheet (Marshall, 2005). Ice sheet models therefore simulate the 3-D ice thermodynamics – advection and diffusion of energy and strain heating due to ice deformation – to model the temperature distribution in the ice sheet.

In addition to internal deformation, ice can flow via basal motion where the bed is at the pressure-melting point, through some combination of subglacial sediment deformation and decoupled sliding over the bed. Basal flow rates, u_b, often exceed the ice motion associated with deformation by one or several orders of magnitude. The vertically-averaged velocity, $\bar{u}$, is calculated from the sum of these two contributions: $\bar{u} = \bar{u}_d + u_b$. Large-scale basal flow generally requires a layer of pressurized subglacial meltwater at the bed. Models make some allowance for basal flow, usually through a local sliding 'law' relating basal flow rates to gravitational shear stress at the bed. There is little knowledge of subglacial hydrological conditions, and the physical processes that determine basal water pressure are not yet modeled or parameterized in ice sheet models. For this reason, sliding rates and the spatial-temporal variability in basal flow are highly uncertain in current models.

Given an estimate of the vertically-averaged ice velocity, the equation for conservation of mass describes the rate of change of ice thickness at each point on the ice sheet:

$$\frac{\partial H}{\partial t} = -\nabla \cdot (\overline{u} H) + b$$

The first term on the right describes the divergence of the ice flux and b is the local mass balance rate (accumulation minus ablation; basal melting can be included in this term but is, generally, negligible compared to surface accumulation and ablation). This is usually considered over one year, i.e., expressed as meters per year of ice-equivalent gain or loss of mass. Given measurements or climate model predictions of b, this equation can be integrated forward to simulate the evolution of ice thickness at all locations on the ice sheet. This is the basis of ice sheet modeling.

than the ice thickness. Further, available bedrock maps for the ice sheet, derived from airborne radar surveys, are currently available only at 5-km resolution. Issues associated with model resolution are discussed below.

Parameterization of surface mass balance is a key parameterization in the models. Most ice sheet modeling efforts to date prescribe the surface mass balance from monthly or annual surface air temperatures and precipitation rates. These can be based on observed (present-day) meteorological patterns or meteorological fields downscaled from a climate model. In general, the actual mass balance fields (accumulation minus melt) simulated by climate models are not accurate enough for coupled ice sheet-climate modeling, in part because general circulation models are too coarse to adequately resolve the ice sheet ablation areas (see section 3.1). This makes it difficult to estimate ablation through a direct surface energy balance, so degree-day methods, driven solely by temperature fields, are generally adopted for melt modeling, following Reeh (1991) and Huybrechts et al. (1991).

Interest in decadal-scale ice sheet changes and the development of improved regional-scale meteorological models (e.g., Box et al., 2006) should soon permit direct estimates of surface mass balance from meteorological models for future predictions, rather than parameterizations based on temperature and precipitation fields. This is desirable because such models are more physically-based and can capture the spatial variability in climate fields as ice sheet geometry and surface properties change; parameterizations tuned to present-day conditions become less reliable as both the climate and the ice sheet depart from their present-day state. This is discussed in more depth in sections 2.2, 3.1, and 3.3.

3.2.3. Modeling of the Greenland Ice Sheet

To date, Greenland Ice Sheet simulations are all based on the shallow-ice approximation and the physics described in Box 3.2 (Letréguilly et al., 1991; Huybrechts et al., 1991, 2004; Ritz et al., 1997; Greve, 1997, 2000; Huybrechts and de Wolde, 1999; Marshall and Cuffey, 2000; Tarasov and Peltier, 2002; Gregory and Huybrechts, 2006). Basal flow is typically assumed to be active wherever the basal ice is at the pressure-melting point and is prescribed as a function of the gravitational driving stress.

Figure 3.6 gives an example of modeled *vs* observed topography for the Greenland Ice Sheet on a 5-km grid. Modern-day ice sheet area and volume can be simulated to within a few percent, but there are often systematic regional biases in modeled ice thickness (Figure 3.6c). The Northeast Greenland Ice Stream (see Figures 2.11 and 2.12) is present in the model (Figure 3.7b), but the modeled flow rates in the ice stream are too slow, contributing to the ice thickness anomaly in this region. The ice margin position is poorly simulated in northern Greenland, leading to thickness anomalies of several 100 meters. The model is anomalously thin in southern-central Greenland, for reasons that are unclear. It may be related to a bias in the climate forcing or modeled rates of basal flow that are excessive, driving a thinner, lower-sloping ice dome.

The velocity field over the whole ice sheet indicates high flow rates on the ice sheet flanks

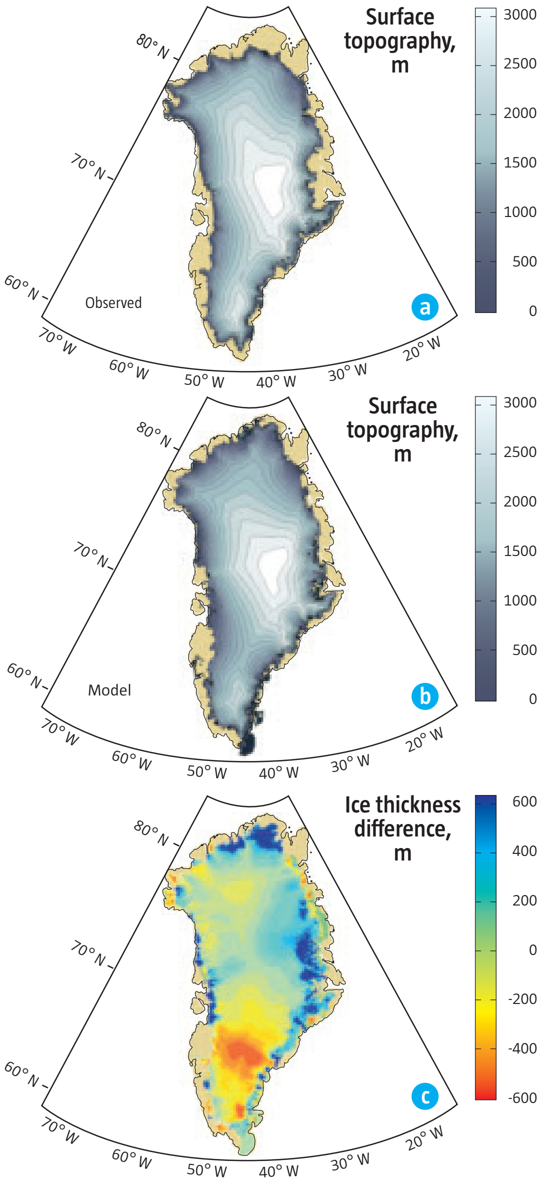


Figure 3.6 *Model reconstructions of the present-day Greenland Ice Sheet on a 5-km grid. (a) Observed and (b) modeled ice sheet surface topography. (c) Modeled minus observed ice sheet thickness. Source: Unpublished results, based on the Greenland Ice Sheet model described by Marshall and Cuffey (2000).*

and areas of concentrated ice flux where they are expected (Figure 2.11). These regions correspond to major topographic channels and drainage outlets, which are resolved in this simulation. However, the model fails to capture the high rates of discharge in major outlet glaciers such as Jakobshavn Isbræ and Helheim Glacier; simulated velocities in some of these systems are an order of magnitude too low. This is most likely to be related to inadequate resolution of the fjord geometry, discussed further below.

Basal flow in this model is parameterized as a function of gravitational driving stress and is enabled wherever the ice sheet bed is at the pressure-melting point. Figure 3.7 plots modeled present-day basal ice temperatures and ice sheet velocities, including contribu-

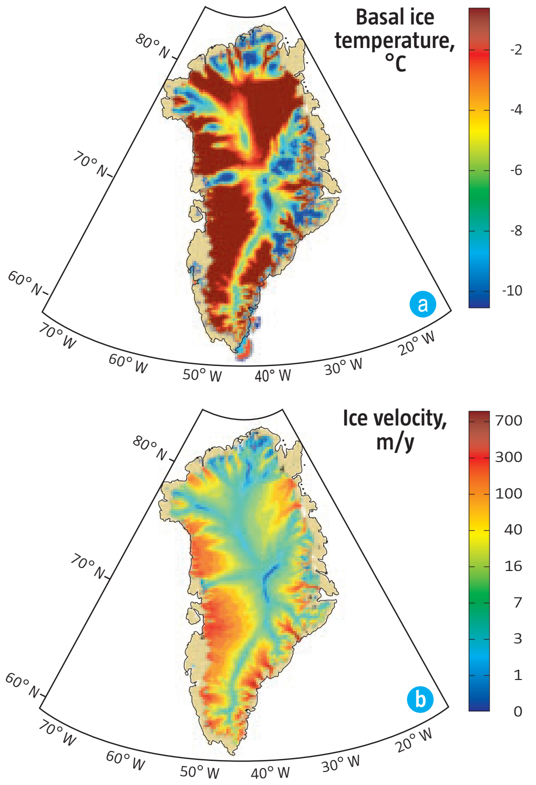


Figure 3.7 *Model reconstructions of the present-day Greenland Ice Sheet (a) basal ice temperatures below the pressure-melting point, and (b) surface ice velocity. Source: Unpublished results, based on the Greenland Ice Sheet model described by Marshall and Cuffey (2000).*

tions from basal flow in warm-based sectors of the ice sheet. Few data are available to validate or test these predictions from the model.

Deep borehole data confirm that central and eastern Greenland are generally cold-based, due to the relatively thin ice that drapes the underlying mountains and the advection of cold ice to the bed along the ice divide. On the ice sheet flanks, deformational heating warms the ice to produce large areas that are at the melting point. In the major topographic channels, this can produce a thick temperate ice layer which has a low effective viscosity, introducing a positive feedback that accelerates deformation and ice flow in these channels.

These models have limitations, elaborated below, but they also capture many large-scale features of the ice sheet dynamics in Greenland. Bearing the model limitations in mind (including a weak representation of fast-flow processes), Figure 3.8 illustrates the modeled response time of the Greenland Ice Sheet to simple warming scenarios. Warming of 2 to 8 °C is introduced as a linearly increasing temperature perturbation over the period 2000-2100, relative to late 20th century baseline temperature fields over the Greenland Ice Sheet as compiled by Ohmura (1987). The temperature anomaly is held constant from 2100 to 3000 AD. Figure 3.9 plots ice sheet configurations at 3000 AD for the 4 °C and 8 °C scenarios.

Most of the ice volume loss in these scenarios comes from southwestern Greenland, with the ice dome in southern Greenland collapsing under sufficient warming. This result has been well established in the modeling by Huybrechts and others, as illustrated in Figure 3.10 from Alley et al. (2005). These results correspond to three different scenarios of future atmospheric CO_2, with values in excess of 750 ppm inducing summer warming of more than 5 °C in Greenland, sufficient to excite collapse of Greenland's southern dome on a timescale of several centuries.

These results reflect the plausible impact of atmospheric warming on increased melt and drawdown of the Greenland Ice Sheet. There is an accelerating ice loss in late stages of the retreat, when the interior of the Greenland Ice Sheet falls below the equilibrium line altitude. Due to simplified fast-flow and ice-marginal physics in the model, as well as one-way climate forcings (i.e., missing climatic feedbacks of ice sheet retreat), the time-scale of modeled Greenland Ice Sheet retreat in Figures 3.8, 3.9 and 3.10 is highly uncertain. Most of the missing feedbacks and processes are believed to give a systematic bias to underpredict ice sheet sensitivity to climate warming; hence, it is expected that Greenland Ice Sheet retreat will proceed more quickly than forecast for a given warming scenario. However, how quickly this could transpire is not well-constrained, because no models with the requisite physics are available to assess this. This is picked up again in section 3.3. The next section addresses several of the outstanding challenges for the current ice sheet models.

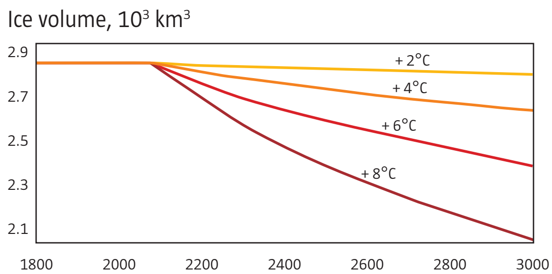


Figure 3.8 *Modeled Greenland Ice Sheet volume response to different climate warming scenarios. Warming scenarios are introduced as a linear temperature perturbation from 2000-2100 AD, with warming held constant at that level (2, 4, 6, or 8 °C) from 2100-3000 AD. Source: Unpublished model results by Marshall (2008). For illustrative purposes only; not intended as a forecast for the ice sheet.*

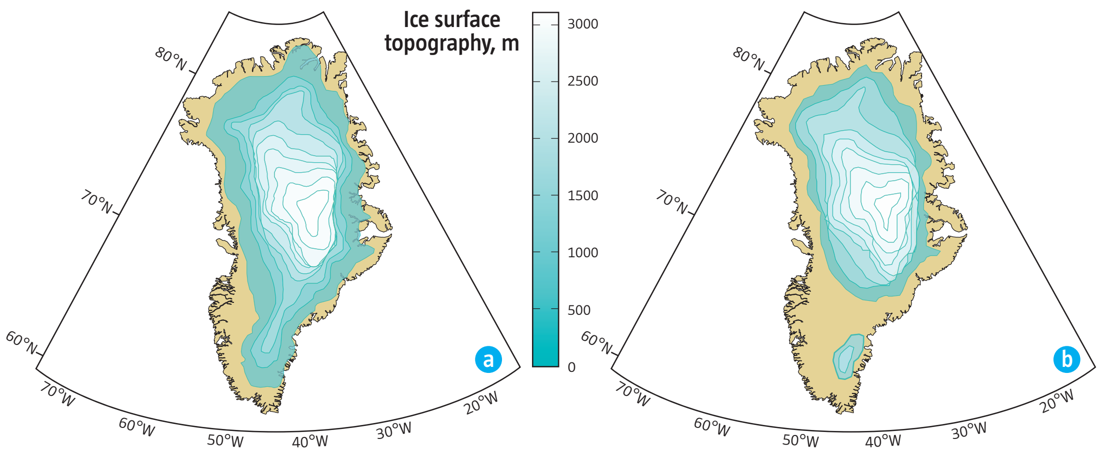


Figure 3.9 *Modeled Greenland Ice Sheet surface topography in 3000 AD for (a) the 4 °C and (b) 8 °C warming scenarios of Figure 3.8. Source: Shawn Marshall, University of Calgary.*

3.2.4. Modeling challenges

The modeling studies cited above have a general skill in simulating Greenland Ice Sheet dynamics, but some features and processes are difficult to capture. These include ice sheet-ocean interactions, ice stream dynamics, basal flow processes (sliding and sediment deformation at the glacier bed), and coupling of ice marginal dynamics and inland ice dynamics. These features are critical to questions of ice sheet sensitivity and response time to climate change.

The Fourth Assessment Report of the IPCC summarized this well:

> *Dynamical processes related to iceflow not included in current models but suggested by recent observations could increase the vulnerability of the ice sheets to warming, increasing future sea level rise. Understanding of these processes is limited and there is no consensus on their magnitude. (IPCC, 2007:17).*

An expanded discussion of shortcomings of existing ice sheet models was provided by van der Veen and ISMASS (2007). Shortcomings with existing Greenland Ice Sheet models fall into three categories: (1) those related to physical processes; (2) those associated with spatial

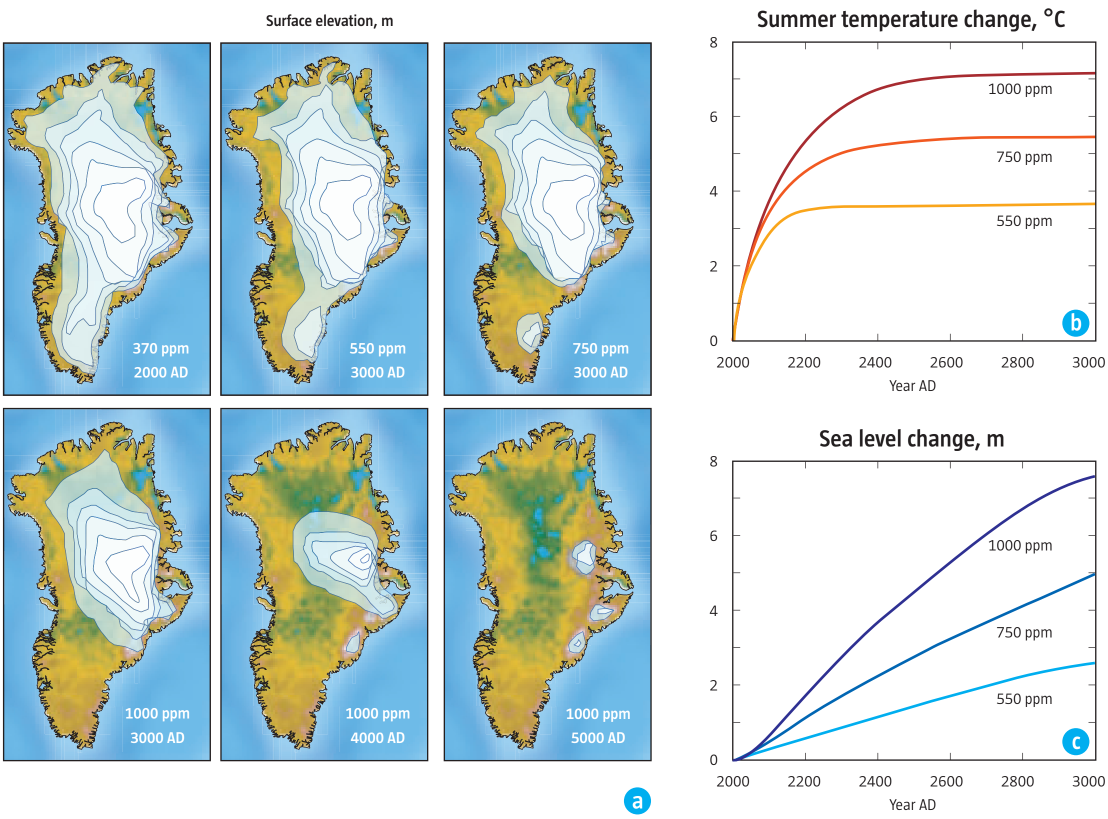


Figure 3.10 *Greenland Ice Sheet simulations for 3000-5000 AD under different greenhouse gas scenarios (a) surface elevation, (b) surface temperature change, and (c) sea level change. Source: Alley et al. (2005).*

resolution; and (3) uncertainties in specification of boundary conditions, including ice-ocean and ice-atmosphere interactions. These issues are elaborated in the following sections, ranked according to a subjective judgment of their relative importance to modeling Greenland Ice Sheet ice dynamics. These processes and considerations have different levels of importance in different parts of the ice sheet, and in some cases the process is not understood well enough to assess how important it will prove to Greenland Ice Sheet response to climate change.

3.2.4.1. Uncertainties associated with physical processes

3.2.4.1.1. Basal flow and hydrology

The extent and physical controls of glacier motion due to basal flow remain poorly known and difficult to quantify (Clarke, 2005). This is true across a wide spectrum of ice masses, including mountain glaciers, outlet glaciers in Greenland and other polar icefields, and Antarctic ice streams. Basal flow is generally the main mechanism for fast flow (except for enhanced creep rates in deep fjords), so must be well quantified to provide a realistic model of ice sheet dynamics.

Elaborate theories have been constructed to describe the sliding of ice over variously shaped obstacles and glacier beds with different roughness characteristics, but it is unclear whether the models are relevant to large-scale basal flow. The models describe salient physical processes in the subglacial environment, including regelation and stress enhanced softening of glacier ice, which allow it to flow around basal obstacles, and the free slip of basal ice over air- and water-filled cavities in the lee of bedrock obstacles. While these processes are likely to contribute to integrated basal slip over large areas of the glacier bed, at least for hard beds (i.e., bedrock rather than deformable till), the governing physics operate at scales of centimeters to meters. Basal motion at scales of tens of meters to tens of kilometers, as observed in Antarctic ice streams and ice marginal regions of Greenland (Joughin et al., 2008a), is more relevant to overall ice flux.

Furthermore, motion associated with regelation and stress-softening around bedrock obstacles increases non-linearly with shear stress (Paterson, 1994:157), whereas basal flow in many settings (e.g., Antarctic ice streams) is associated with low values of gravitational driving stress: Thin, low-sloping ice with subglacial water pressures near flotation, implying low values of basal shear stress over large regions (Bindschadler et al., 2000). Summer speed-ups in valley and outlet glaciers and the ice marginal region of the Greenland Ice Sheet provide another illustration of this. These speed-ups are associated with negligible change in surface geometry (i.e., gravitational driving stress), but are instead driven by reductions in basal friction due to the influx of surface meltwater to the bed (e.g., Iken and Bindschadler, 1986; Copland et al., 2003; Das et al., 2008; Joughin et al., 2008a).

The mechanics of large-scale basal motion therefore appear to be governed by the extent of the glacier bed that experiences ice-bed decoupling due to high subglacial water pressures. High water pressure promotes basal flow by effectively floating the glacier above the bed, reducing basal traction. This applies to both basal sliding and subglacial sediment deformation, although the rheological properties and supply of basal sediments also need to be considered for the latter.

The need to understand hydrological influences on basal flow and to incorporate these processes into models is pressing, considering recent evidence that surface meltwater can drain to the bed through hundreds of meters of cold ice in marginal areas of the Greenland Ice Sheet (Zwally et al., 2002b; Das et al., 2008) and smaller-scale polar icefields (Boon and Sharp, 2003). Similar to the long-standing observations from mountain glaciers, this influx of meltwater is capable of triggering transient speed-ups, increasing the discharge of ice (Zwally et al., 2002b; Joughin et al., 2008a). This presents a direct mechanism by which climate change can influence ice dynamics. Coupled with trends of increasing spatial extent and intensity of the melt season in Greenland (Tedesco, 2007), accelerated summertime flow is expected in future decades in the marginal zones of the Greenland Ice Sheet.

The magnitude of the speed-up and the extent to which ice sheet drawdown will propagate inland are less clear. Joughin et al. (2008a) presented observations over a 300-km wide region in southwest Greenland which indicate that summer speed-ups of the major outlet glaciers, including Jakobshavn Isbræ, contribute only a few percent to the total annual discharge. Speed-ups on the slow-moving ice sheet flank (i.e., away from the major outlet glaciers) are relatively greater, increasing annual discharge by as much as 25%, but this region does not account for much of the total dynamic discharge of the Greenland Ice Sheet. Therefore, this mechanism represents a pathway for ice sheet response to climate forcing, but one which may be relatively stable (i.e., non-catastrophic).

These influences on ice dynamics are generally absent in the large-scale models that predict Greenland Ice Sheet response to climate change. Physics-based models of ice sheet hydrology are needed to simulate these processes. In a flowline model of the western flank of the Greenland Ice Sheet, Parizek and Alley (2004) simulated the supraglacial hydrology and englacial drainage to the bed, assuming that surface drainage induces speed-ups as seen by Zwally et al. (2002b). Extrapolating their results to the whole ice sheet, their most conservative scenarios gave 12% to 38% more volume loss from the Greenland Ice Sheet over the next several centuries due to surface-meltwater induced lubrication, with the effect increasing with warming. Parizek and Alley (2004) also suggested that the effect could be much higher, with the upper range of their modeling giving a tripling of the ice volume loss in Greenland, relative to models that exclude surface meltwater effects.

These models assume that basal flow is proportional to water supply at the bed, but subglacial water storage and basal water pressures are what really matter for ice-bed coupling and basal flow. The physics that governs the relationship between subglacial water pressure, bed geology, bed roughness, and basal traction still needs to be elucidated and parameterized in a way that is physically valid and testable in models.

A subglacial hydrological model is also required to provide a realistic estimate of spatial and temporal variations in basal water pressure. These have begun to be developed and coupled with models of glacier dynamics (Arnold and Sharp, 2002; Flowers and Clarke, 2002; Johnson and Fastook, 2002; Flowers et al., 2005; Marshall et al., 2005), but subglacial hydrological conditions are notoriously difficult to observe, so models are difficult to test and constrain at large scales. To date, most models of subglacial hydrology treat the subglacial water system as a distributed water sheet, with water fluxes driven by hydraulic potential gradients. Conduit physics and the nuances of regime shifts between distributed and channelized drainage systems need to be better understood and included in the models in order to capture spatio-temporal patterns of basal water pressure and their influence on glacier dynamics. This has yet to be attempted in Greenland.

3.2.4.1.2. High-order stresses (horizontal stress coupling)

The shallow-ice approximation estimates ice motion as a function of the local gravitational driving stress, with the explicit assumption that non-local effects are negligible. This is known to be a poor approximation in certain parts of the ice sheet, such as the ice divide, ice streams, ice shelves, and ice sheet margins. In these settings, higher-order stress terms such as horizontal shear stress and longitudinal stress gradients, sometimes known as non-local stress coupling effects, become important to ice deformation. For instance, horizontal shear stresses at ice stream margins and next to valley walls play an important role in holding back ice flow, while longitudinal strains (stretching) represent the primary mode of ice deformation in ice shelves and floating ice tongues.

Longitudinal stress coupling also provides a mechanism for the transmission of ice-marginal forcings to the interior of the ice sheet. Current thinning of Jakobshavn Isbræ is accompanied by rapid retreat of the calving terminus (Sohn et al., 1998; Joughin et al., 2004). Joughin et al. (2004) proposed that the velocity increase resulted from development of large rifts in the floating tongue, followed soon thereafter by complete collapse of this tongue. Associated reductions in buttressing would allow increased discharge from the grounded portion of the glacier (Thomas, 2004). On the other hand, Pfeffer (2007) proposed that thinning could have decreased the height above buoyancy, thereby reducing the effective basal water pressure. This, in turn, would lead to greater sliding speeds, especially if the ice thickness approached floatation. To adequately incorporate these processes into numerical models and to test the validity of these competing hypotheses, it is necessary to incorporate stress transmission across the grounding line.

Models of glacier dynamics with higher-order stresses have become more common in studies of valley glaciers as well as regional ice sheet applications (Albrecht et al., 2000; Schneeberger et al., 2001; Pattyn, 2002, 2003; Saito et al., 2003). These advances are relatively new, and have yet to be applied to whole-ice sheet simulations in Greenland. High-order models such as that of Payne et al. (2004) and Price et al. (2008) demonstrate the importance of longitudinal stress coupling for accurate simulation of ice sheet response to perturbations at the ice sheet margin, i.e., inland propagation of changes due to ice-marginal thinning or acceleration.

3.2.4.1.3. Iceberg calving

Ice sheet grounding lines, floating ice tongues, and iceberg calving processes are poorly represented in current glaciological models. Existing models generally determine the position of the calving terminus (i.e., marine ice front) by either prescribing an ice limit based on a maximum water depth (e.g., the continental shelf break), or by freely simulating floating ice tongue advance based on a simple parameterization of iceberg calving, typically prescribing calving rates to be proportional to water depth. Zweck and Huybrechts (2005) examined several different treatments of calving in large-scale models; all are simple parameterizations that do not describe the actual process(es) and should be considered to be statistical-empirical models that aim to capture the bulk loss of ice through this mechanism.

The physical controls of iceberg calving are not well understood and may not be deterministic. Iceberg calving requires fractures to initiate and propagate at the ice front. The process has been well-observed, but the environmental controls of calving rates are difficult to isolate; ocean and air temperature, surface meltwater, sea-ice conditions, and tidal flexure at the ice front may all have an influence on calving, but there is no simple relationship. Sohn et al. (1998) observed an increase in calving rate following the onset of spring melting and found a strong correlation between surface melting on the Greenland Ice Sheet, the melting of fjord ice, and the rapid increase in calving rate during summer. Their observations are insufficient, however, to resolve conclusively whether the presence of abundant surface meltwater enhances the ability of surface crevasses to penetrate the ice thickness, or whether the break-up of confining fjord ice reduces the restraining force acting on the calving front.

Recent evidence indicates the likely role of regional ocean warming in destabilizing the floating ice tongue of Jakobshavn Isbræ (Holland et al., 2008). This is also consistent with tidewater-style retreat of Helheim Glacier on the southeast coast of Greenland (Nick et al., 2009). Ocean-ice sheet coupling is lacking in the current models, but clearly needs to be embraced for predictive models of ice sheet response to climate warming.

3.2.4.2. Uncertainties associated with model resolution

Whole ice sheet models for Greenland now operate at scales of 5 to 20 km, but scale problems still arise because of the small spatial scale of most of the outlet glaciers where drainage to the sea is concentrated. For example, Jakobshavn Isbræ, which drains ~7% of the total area of the Greenland Ice Sheet, is less than 10 km wide over its entire ~100 km length (Echelmeyer et al., 1991). This glacier overlies a narrow (6 to 10 km) and deep (~1500 m below the surrounding basal landscape) subglacial trough (Clarke and Echelmeyer, 1996; Gogineni et al., 2006; Lohoefener et al., 2006) that cannot be resolved with a horizontal grid spacing of 10 km or more. This limits the capacity of models to simulate the enhanced creep and high rates of flow that are observed in Jakobshavn Isbræ and other major outlet glaciers that drain the Greenland Ice Sheet.

At the same time, detailed and accurate bed topography upstream of the grounding line or calving front is needed to assess the possible irreversibility of terminus retreat for Greenland's major outlet glaciers. Howat et al. (2007) suggested that the rapid retreat of the Helheim Glacier terminus occurred as the calving front retreated into deeper water and continued until reaching a reversed bed slope, similar to the behavior of tidewater glaciers (Nick et al., 2009).

Even if spatial resolution is improved sufficiently to capture small-scale basal topography, narrow channels no more than a few kilometers wide and up to 1500 m deep challenge the assumptions in the shallow-ice approximation. As noted by Clarke and Echelmeyer (1996), these bedrock troughs are similar to valley walls in that the near-vertical faces produce lateral drag that partially opposes the driving stress – horizontal shear stresses become important in taking up the gravitational driving stress. Also, an overlying temperate layer of fast moving ice is embedded within more slowly moving ice, which may provide additional resistance, similar to the situation on West Antarctic ice streams. As a start, an effective shape factor could be introduced (Clarke and Echelmeyer, 1996) to account for the additional flow resistance, but no rigorous tests have been conducted to evaluate the applicability of this approach.

Additional uncertainty is introduced by the well-known (if rarely discussed) sensitivity of modeled ice margin dynamics to grid resolution (Abe-Ouchi and Blatter, 1993). With a fixed grid and a free boundary, the standard treatment in continental ice sheet models, the advance and retreat of the ice sheet margin proceed by discrete 'jumps' to adjacent grid cells. This process is sensitive to grid resolution because the flux of ice into a neighboring cell is a function of the local surface slope. This sensitivity combines with mass balance elevation feedbacks to make ice margins more mobile under higher resolution; it is simpler for the ice sheet to advance as grid size $\Delta x \rightarrow 0$. Strategies involving adaptive grids (e.g., Price et al., 2008) or sub-grid tracking of the ice sheet margin are needed to give convergent behavior that is insensitive to model resolution.

3.2.4.3. Uncertainties associated with boundary conditions

3.2.4.3.1. Geothermal heat flux

Pronounced subglacial topography such as that of Jakobshavn Isbræ intensifies the local geothermal heat flux by as much as 100% (van der Veen et al., 2007), potentially introducing an important feedback between subglacial topography and fast flow. On a broader scale, significant variations in crustal thickness have been inferred from airborne gravity, with the thinnest crust underlying the Northeast Greenland Ice Stream and other major drainage routes (Braun et al., 2007). Where the crust is relatively thin, geothermal heat flow may be expected to be greater than the continental average (Leftwich et al., 2007). High geothermal heat flux in northeast Greenland is also believed to play a role in providing the temperate, lubricated bed that enables fast flow in this ice stream (Fahnestock et al., 2001). The extent to which spatial variations in heat supplied to the glacier base affect the regional flow regime remains to be resolved (Greve, 2005). This could explain discrepancies in areas where basal meltwater has been detected but where models incorrectly predict the basal ice to be frozen to the bed.

3.2.4.3.2. Climatic forcing

Ice sheet models require distributed (spatially-resolved) atmospheric fields to provide surface mass balance fields. To examine climate change scenarios, temporal evolution of atmospheric conditions needs to be characterized. This appears as a boundary condition in ice sheet models, and is prescribed simplistically in most ice sheet model experiments. Uncertainties in spatial-temporal climate evolution may well dominate uncertainties in ice dynamics modeling for prediction of Greenland Ice Sheet response to climate change. This is discussed in more detail in section 3.3.

In addition to atmospheric fields, ocean and sea-ice conditions directly impact on marine outlet glaciers (e.g., Holland et al., 2008; section 2.3). Marine influences are parameterized through the ocean heat flux that is prescribed as a basal boundary condition for floating ice; this is generally held fixed in current models, e.g., a uniform flux of 2 W/m^2. Plus, ocean conditions can be incorporated in calving parameterizations, but this has been rarely done, in part because of unclear mechanistic links. Except for a small number of regional studies, ocean and ice sheet models are largely uncoupled and ocean variability is not considered in ice sheet modeling. Challenges in ice sheet modeling are revisited in the summary comments of section 3.4, where the main sources of uncertainty for predicting Greenland Ice Sheet sensitivity to climate change are discussed.

3.3. Future climate and ice sheet scenarios: Response of the Greenland Ice Sheet to a warmer climate

3.3.1. Objectives of coupled ice sheet-climate modeling

Ice sheet models, as described in section 3.2, can investigate the ice-dynamical response to idealized climate change scenarios, such as a uniform cooling or warming. However, patterns of change in climate fields (e.g., temperature, precipitation, radiation, and winds) will not be uniform over the area and elevation range of the Greenland Ice Sheet. Realistic climate forcing is required to provide more detail on ice sheet evolution in the coming decades and centuries. Ice sheet evolution over long timescales also introduces a number of feedbacks that need to be captured in future climate forecasts, such as changing elevation and albedo fields over the ice sheet. Coupled ice-sheet climate models are needed to capture these feedbacks.

The next section describes the coupling approaches adopted in ice sheet-climate modeling. Model predictions of Greenland Ice Sheet response to future warming are then summarized, followed by a discussion of some of the limitations of current models and priorities for future development.

3.3.2. Coupling approaches

3.3.2.1. One-way and two-way coupling

Both one-way and two-way approaches have been adopted for coupling atmospheric climate and ice sheet models. In one-way coupling, output from the climate model provides input for the ice sheet model (e.g., surface mass balance fields), but the ice sheet model output does not force the climate model. This allows existing climate change scenarios (e.g., the ensemble of IPCC climate model forecasts) to be applied to ice sheet simulations. To increase realism of the exchange, relationships between surface elevation, latitude, and surface conditions have been prescribed to allow climate forcings to change as the ice sheet surface evolves (Huybrechts et al., 1991). This includes elevation-temperature (lapse rate) feedbacks, precipitation rates that increase with warming temperatures, roughly in accord with a Clausius-Clapeyron relationship, and a modification of local precipitation rates as a function of local surface slope (Ritz et al., 1997).

In two-way coupling, climate model output forces the ice sheet model, while output from the ice sheet model (e.g., ice topography, albedo fields) serves as input to the climate model. Two-way coupling is desirable as the simulation more realistically represents the physical system and the climate feedbacks associated with temporal evolution of the ice sheet. However, as two-way coupling is more challenging than one-way coupling and there are, generally, discrepancies of resolution between climate and ice sheet models, most ice sheet modeling experiments to date have used one-way coupling.

The model results in Figures 3.7, 3.8, 3.9, and 3.10 are examples of ice sheet simulations with one-way climate forcing. Surface mass balance is estimated from temperature and precipitation fields, using a degree-day methodology to estimate surface melt and the fraction of precipitation to fall as snow (Reeh, 1991). Modern precipitation and temperature fields are taken as the baseline and temporal variability is prescribed as a perturbation based on ice core reconstructions, climate model forecasts, or simple scenarios (e.g., a step warming of 2 or 4 °C). Many Greenland

Ice Sheet modeling studies have adopted this approach (Huybrechts et al., 1991, 2004; Ritz et al., 1997; Huybrechts and de Wolde, 1999; Greve, 2000; Alley et al., 2005; Gregory and Huybrechts, 2006).

While several parameterized ice-climate feedbacks are included in these numerical experiments, the one-way approach may omit key physical processes that affect the circulation of the atmosphere and ocean, which in turn would modify temperature and accumulation patterns on the ice sheet. A small number of groups have begun to interactively couple ice sheet models with climate models of varying degrees of complexity and resolution. Huybrechts et al. (2002), Fichefet et al. (2003), and Driesschaert et al. (2007) forced an ice sheet model with an AOGCM and examined how Greenland Ice Sheet retreat may affect oceanic circulation in the coming decades. These studies use one-way ice-sheet forcing from the atmospheric model. Other studies focus on ice sheet-atmosphere feedbacks (Toniazzo et al., 2004; Gregory et al., 2004), while a small number of studies explore fully-coupled ice sheet and AOGCM models (Ridley et al., 2005; Mikolajewicz et al., 2007b; Vizcaíno et al., 2008).

A number of climate-system and ice sheet feedbacks identified in coupled model studies are discussed in section 3.3.3. Specific predictions of Greenland Ice Sheet retreat and sea-level rise in response to climate change are summarized in section 3.3.4.

3.3.2.2. Climate downscaling

The experiments described above require atmospheric fields to be downscaled to the ice sheet model grid for mass balance modeling. This is due to the mismatch that exists between even high-resolution AOGCMs running at 1.25 ° (~150 km) grid resolution and ice sheet models (10 to 20 km). Grid downscaling is thus required and commonly takes the form of a simple interpolation.

This step requires a good understanding of altitude-surface mass balance relationships in different climate regions of the ice sheet. Constant and spatially-uniform temperature lapse rates (linear temperature decreases with elevation) are usually used to downscale temperatures, often with different summer and annual lapse rates (Huybrechts et al., 1991). This does not account for the effects of inversions, katabatic wind induced cooling at low elevations on the ice sheet, or surface energy balance effects that govern near-surface (*vs* free-air) temperatures, so a uniform lapse rate is a simplification. Precipitation is more difficult to downscale, as it can vary over small spatial scales, particularly in association with orographic forcing of precipitation at the ice sheet margin. Most modeling studies to date are based on perturbations to the modern-day spatial pattern of precipitation in Greenland.

Under interactive coupling, the influence of significant changes in ice sheet topography and boundary conditions are better accounted for. For example, processes such as orographic precipitation and the energy balance influence of increasing areas of open water and ice-free land can be included as feedbacks in the model, rather than an assumption that modern-day climate patterns will hold.

Because of the uncertainty associated with interpolation of meteorological fields, the downscaled climate is typically parameterized into positive-degree-day (PDD) input in order to simulate snow / ice melt; a full surface energy balance is not generally attempted. PDD approaches have proven reasonable for melt modeling in Greenland, but downscaling or interpolation of temperature or PDD fields does not ensure conservation of energy between the applied surface climatology and the climate model. There is a similar concern for downscaled precipitation fields. Conservation of mass with respect to larger-scale climate model precipitation and moisture fields is not straightforward and is often disregarded. Conservation of energy and mass warrant serious attention, particularly in two-way coupling experiments where there is a possibility of cumulative biases leading to 'drift' in transient simulations.

3.3.2.3. Mass balance estimation in coupled models

For the reasons outlined above, most Greenland Ice Sheet modeling studies to date estimate surface mass balance from degree-day calculations as a function of climate (i.e., temperature and precipitation) perturbations. As was discussed in detail in section 2.2, surface mass balance fields have also been simulated directly from climate model re-analyses (Hanna et al., 2002, 2005, 2008), remote sensing studies (Mote, 2003), and regional climate modeling (Box et al., 2006; Fettweiss et al., 2007, 2008). Like the reconstructions by Hanna and co-workers, regional climate models make use of re-analyzed climatology to provide the boundary conditions for atmospheric simulations. The main difference is the nature of accumulation and melt modeling in the two approaches; Hanna and co-workers use a degree-day approach, while regional climate models distribute precipitation based on atmospheric dynamics and estimate melt from a surface energy balance.

Table 2.1 gives a summary of simulated surface mass balance terms from the historical period for several of these efforts. These studies provide detailed estimates of Greenland Ice Sheet mass balance through the second half of the 20th century, but future forecasts cannot be carried out with re-analyzed climatology. Regional climate models offer a potential solution to this, as they can be trained / calibrated on the historical period and forced by global-model fields for future simulations. This has yet to be attempted for Greenland.

Regional climate models have the inherent capacity to represent the physical processes of how snowfall and melt relate to the local and regional surface characteristics and energy budgets, and these models are likely to form the basis for the next generation of Greenland Ice Sheet predictions. However, climate models exhibit systematic biases, e.g., land surface temperature too warm or too cold, in response to biases in model surface energy budget. These errors have been attributed largely to cloud-radiation bias, e.g., too much or too little downward longwave radiation. Terrain biases are another typical source of systematic bias, commonly arising from inaccurate representation of the terrain or from the need to smooth the surface topography at the 'steep' ice sheet margin to avoid atmospheric dynamical instabilities, e.g., gravity waves. Through comparison with *in situ* observations, calibration procedures can be developed and uncertainties can be quantified (e.g., Box et al., 2004, 2006). However, a careful assessment of the stationarity of model biases is needed before these models can be applied with confidence to future forecasts.

3.3.2.4. Ice sheet-ocean coupling

Regional-scale coastal and sub-ice shelf oceanographic models are also needed to improve ice-climate coupling, something that has been explored on a regional scale in Antarctica, but has not been considered in Greenland. Other than an influence of global sea level on calving rates, ice-ocean feedbacks have not been examined on the scale of the whole ice sheet; ocean temperatures and heat fluxes, sea-ice conditions, and other potential controls of iceberg calving and melting at the ice-ocean interface are invariant in current models. This point is turned to in section 3.4.

3.3.3. Feedbacks in coupled models

Simulations with two-way coupling of the ice-ocean-atmosphere system reveal both positive and negative feedbacks on ice sheet retreat in a warmer world. Huybrechts et al. (2002) and Fichefet et al. (2003) examined the coupled response of a 3-D thermo-mechanical model of the Greenland Ice Sheet and an AOGCM to future climate warming (IPCC SRESB2 scenario, mid-range greenhouse gas forcing). Simulated ice sheet melt is put into the ocean circulation model. Modeled melt rates are modest in the 21st century (global sea level rise of 4 cm in response to a warming of 4.5 °C over Greenland), but Fichefet et al. (2003) concluded that the freshwater addition to the ocean from ice sheet melt would contribute to weakened Atlantic meridional overturning circulation by the end of the 21st century, a feedback that lowers temperatures over East Greenland and partly counteracts warming in this region.

In a subsequent study, Driesschaert et al. (2007) concluded that Greenland Ice Sheet melting only significantly impacts on the Atlantic meridional overturning circulation in the 21st century under severe future warming scenarios. This is echoed in the results of Mikolajewicz et al. (2007b) and Vizcaíno et al. (2008), who used a coarse-resolution AOGCM interactively coupled with an ice-sheet model at 80-km resolution to explore millennial-scale future climate and ice sheet changes in Greenland. They also predicted that weakened

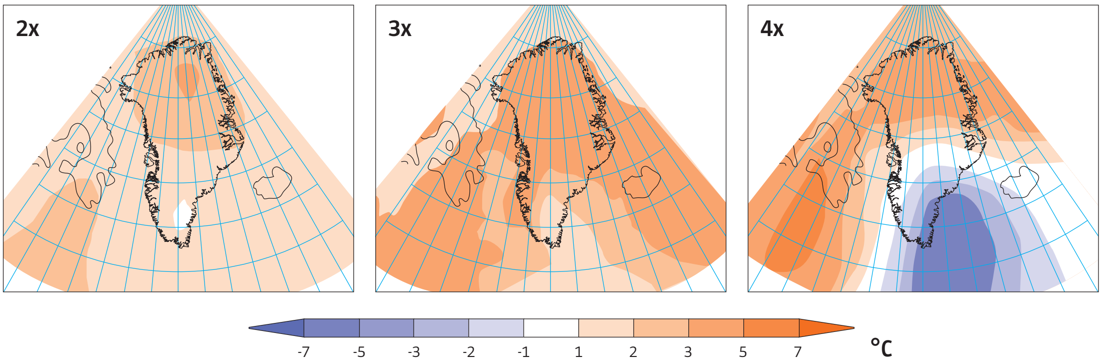


Figure 3.11 *Modeled changes in summer temperature over Greenland as a result of 2×, 3×, and 4× pre-industrial carbon dioxide scenarios. These snapshots correspond to the period 2900-3000 AD and include ice-dynamical feedbacks. Source: Vizcaíno et al. (2008).*

overturning circulation will reduce the degree of warming over southern Greenland (Figure 3.11), buffering Greenland Ice Sheet melt. However, changes in the hydrological cycle of the North Atlantic region are the main reason for this; Greenland Ice Sheet run-off is about 10% of the total freshwater forcing of the system (Mikolajewicz et al., 2007b).

Vizcaíno et al. (2008) predicted that a doubling or tripling of atmospheric CO_2 would result in 2 to 5 °C of summer warming in Greenland, with greater warming in northern Greenland (Figure 3.11). Simulations with 4 × CO_2 trigger a shutdown in North Atlantic deepwater formation, which creates a dramatic reversal in the projections and a cooling of up to 3 °C in southern Greenland.

Atmospheric feedbacks in response to Greenland meltback have been examined extensively through the productive collaboration of Philippe Huybrechts, Jonathan Gregory, and colleagues, using Huybrechts' ice sheet model and the Hadley Centre AOGCM. Toniazzo et al. (2004) examined the effect of removing the Greenland Ice Sheet on regional atmospheric circulation and climate. A lower (or absent) ice sheet reduces the persistent anticyclonic flow over the ice sheet, particularly in winter months, causing a net increase in precipitation in eastern and northern Greenland. Low elevation, reduced albedo, and the development of vegetation (e.g., boreal forest) cause greatly intensified summer warming and seasonal melting. No perennial snow is predicted. This prevents the ice sheet from re-growing in today's climate; it is an artifact of the Pleistocene, preserved by its own elevation and regional cooling influence.

Ridley et al. (2005) examined two-way coupling for a 4 × CO_2 climate. In this analysis, the ice sheet model output forces the climate model through changes in ice topography, surface albedo, and freshwater run-off. Ridley et al. (2005) concluded that long-term trends are primarily sensitive to topographic feedbacks (i.e., elevation-temperature effects) that are reasonably well captured in the one-way forcing parameterizations of, for example, Huybrechts et al. (1991). There are some additional feedback effects that appear to slow down ice loss. Warming of the ice-free region at the ice sheet margin, which intensifies under climate warming (and with ice sheet retreat) generates a convective circulation that cycles warm air to the interior of the ice sheet, while enhancing the intensity of katabatic return flows at the surface. This cools the lower elevations, reducing melt rates at the ice margin and helping to preserve the ice there.

Ice dynamics introduce another negative feedback in the models during early stages of the ice sheet retreat. As described by Huybrechts and de Wolde (1999) and Ridley et al. (2005), higher ablation at the margins and increased snowfall at high elevations combine to give greater balance gradients and steeper ice sheet slopes. This produces an increased ice flux which buffers drawdown at the ice sheet margin, maintaining thicker ice and reducing the elevation-temperature feedbacks. This reduces the initial rate of ice sheet retreat, although it is not sustainable as drawdown in the interior regions eventually causes an expansion of the ablation area, with elevation and albedo feedbacks that hasten ice sheet retreat.

3.3.4. Predicted Greenland Ice Sheet response to warming

Bearing in mind the missing ice-dynamical processes and ice sheet-climate coupling challenges, a number of insights are available from simulations to date. It is clear that the Greenland Ice Sheet is very sensitive to climate warming. In steady-state, a regional warming of more than 3 °C is likely to be sufficient to precipitate ice sheet retreat in southern Greenland (Huybrechts et al., 1991; Alley et al., 2005). This is illustrated in Figure 3.10; the southern dome retreats by 3000 AD in these simulations at CO_2 levels between 550 and 750 ppm, which correspond to a summer warming of 4 to 5 °C in Greenland.

Based on an ensemble of models, including high-resolution (~1 °) climate model representations of temperature and precipitation patterns over the ice sheet, Gregory and Huybrechts (2006) concluded that an average annual regional warming of 4.5 °C would push the Greenland Ice Sheet into a negative surface mass balance. The threshold temperature is higher than in earlier studies (e.g., 2.7 °C in Huybrechts et al., 1991; 3 °C in Gregory et al., 2004), in part because summer warming over Greenland, ΔT_S, is forecast to be less than the mean annual warming, ΔT_A: Gregory and Huybrechts (2006) reported an average value $\Delta T_S / \Delta T_A = 0.81$. Summer temperature is the critical parameter for Greenland's mass balance, as this is the main driver of annual surface melt and run-off.

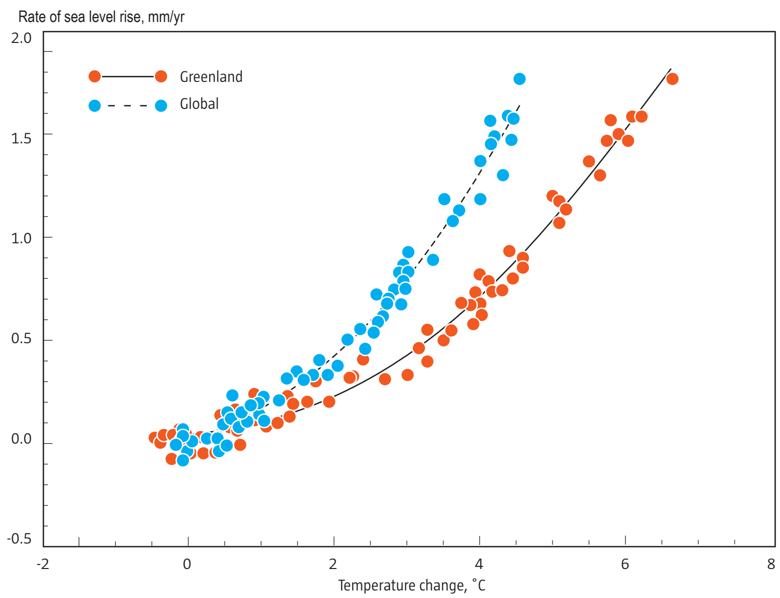
A warming of 4.5 °C in Greenland corresponds to an average global warming of 3.1 °C in the ensemble of models explored by Gregory and Huybrechts (2006). The relationship is plotted in Figure 3.12, along with estimates of rates of sea level rise due to surface mass balance losses as a function of mean annual warming. There is an average polar amplification factor of ~1.5 over Greenland, less than the average value for the Arctic. Part of the moderate polar amplification factor here may be due to the cooling influence of weakened deepwater formation in the Nordic Seas that is predicted by many AOGCMs in the coming decades (see for example Figure 3.11). This is addressed in more detail in section 4.1.

The critical warming value of 4 to 5 °C for Greenland to go into a state of negative surface mass balance is often quoted as a threshold for 'irreversible' ice sheet decline. Dynamical ice losses would place the Greenland Ice Sheet into a negative total mass balance well before this surface mass balance threshold was met – this threshold appears to have been reached in the 1990s and 2000s with a regional warming of close to 1 °C (section 2.4). However, the temperature threshold for surface mass balance to become negative is considered to be terminal because ice sheet thinning would provide a positive feedback to ice sheet decay. If this proceeds too far, ice sheet drawdown would make it difficult to reverse the decline, even with temperature stabilization. However, the timescale for such a decline is long – many centuries to millennia – so it is difficult to judge irreversibility on this timescale.

Charbit et al. (2008) argued that 3000 Gt of cumulative CO_2 emissions could lead past this threshold for an irreversible Greenland Ice Sheet retreat, such that anthropogenic activity in the 21st century could commit the world to several centuries of sea level rise. Ridley et al. (2009) examined this question through coupled modeling with climate restored to its pre-industrial state for several different Greenland Ice Sheet configurations. They argued for a potential two-stage destabilization of the ice sheet in Greenland, with reductions to 80 to 90% of the current ice sheet volume inducing ice loss in southern Greenland and an irreversible sea level rise of about 1.3 m over several centuries. If the ice sheet retreats further before climate is stabilized (i.e., under greater or sustained warming), such that the ice sheet falls below half of its current volume, Ridley et al. (2009) predicted an irreversible sea level rise of about 5 m.

Such a decline is consistent with reconstruc-

Figure 3.12 *Rates of sea level change as a function of mean global temperature change and the temperature change over Greenland for future climate simulations in the ensemble of AOGCMs examined by Gregory and Huybrechts (2006). Source: Gregory and Huybrechts (2006).*



tions of a reduced Greenland Ice Sheet in the last interglacial period (Cuffey and Marshall, 2000), in response to orbitally induced spring and summer warming at high northern latitudes. Other ice sheet modeling studies predict a less severe ice sheet decline at this time (Huybrechts et al., 1991), due to different assumptions about the ice core isotope-temperature relationship. Miller et al. (2006) and Otto-Bliesner et al. (2006) estimated a summer warming of about 5 °C near Greenland at the peak of the Eemian warming, although it is not clear how long this warming persisted. Ice sheet models predict an almost complete collapse of the ice sheet if warming of this magnitude persists for several thousand years, but the modeled retreat takes several millennia – changes over the first few centuries are modest.

For future warming, the extent of modeled Greenland Ice Sheet retreat by 2100 is primarily a function of the change in surface mass balance, hence the climate scenario. Ice dynamics plays only a minor role over this time scale in the current models, for two main reasons: (1) elevation-induced surface mass balance and ice-dynamic feedbacks are minor over a one-century timescale, and (2) ice dynamics models lack a direct connection with climate (e.g., ocean or surface-meltwater forcing). The latter is consistent with ice-sheet model simulations of the last century; there is little to no interannual variability in modeled ice dynamics. This is partly because ice dynamics in the current generation of ice sheet models is sensitive only to local gravitational driving stress (see section 3.2), which changes slowly.

Under the IPCC IS92a climate change scenario, Huybrechts and de Wolde (1999) simulated a contribution to sea level rise from the Greenland Ice Sheet of 10.6 cm by 2100. Using a more sophisticated climate treatment, Huybrechts et al. (2004) gave a reduced estimate, with an average 21st century rate of sea level rise of 0.45 to 0.49 mm/y from an ensemble of climate models: a net sea level rise of 4.5 to 4.9 cm by 2100 (Figure 3.13). Taking a range of future climate scenarios into consideration, reference model estimates of Huybrechts et al. (2004) broaden to 2 to 7 cm by 2100. Alley et al. (2005) gave a similar value for 2100, 5 ± 4 cm. Greenland Ice Sheet retreat has also been studied by van de Wal and Oerlemans

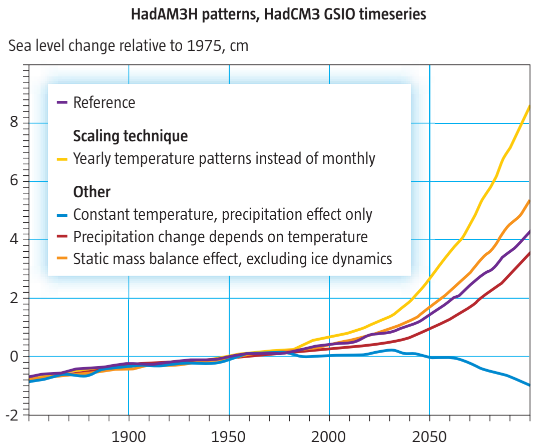


Figure 3.13 *Modeled sea level rise from Greenland Ice Sheet retreat, 1860-2100, driven by modeled high-resolution (1.125°) temperature and precipitation patterns over Greenland, with temporal climate perturbations from the Hadley Centre AOGCM. The different lines denote different treatments of climate. Source: Huybrechts et al. (2004).*

(1997), with a simpler (2-D, vertically-integrated) ice sheet model but using an energy balance model to calculate surface melt. They estimated 5 to 11 cm of surface mass balance driven sea level rise from Greenland by 2100.

The analysis of Gregory and Huybrechts (2006) is the most detailed examination to date of the expected future contributions to sea level rise from warming-driven changes in Greenland's surface mass balance. Figure 3.14 plots the phase space of simulated sea level rise as a function of changes in temperature and precipitation. Decadal changes in temperature and precipitation in a suite of AOGCM simulations of the 21st century are plotted with dots in this Figure, indicating projected temperature increases of up to 8 °C along with precipitation increases of up to 50%. The net effect on surface mass balance is negative in the models; higher melt rates trump increases in accumulation. The corresponding rate of sea level rise due to declining surface mass balance ranges from 0 to 2 mm/y for a warming of 0 to 8 °C.

Based on this collection of model studies, the direct surface mass balance effect of expected warming in Greenland is likely to be a sea level rise of 5 to 10 cm by 2100. Within the range of anticipated climate change, ice volume losses in excess of 1 m of sea-level equivalent from the Greenland Ice Sheet require several centuries in all published modeling studies to date. The complete demise of the ice sheet requires 3000 years or more under most future climate change scenarios. The most severe climate warming simulated by Greve (2000), i.e., 12 °C over Greenland, precipitates a full collapse (and a 7 m sea level rise) in 1000 years. Predictions by Greve (2000) for more moderate warming scenarios

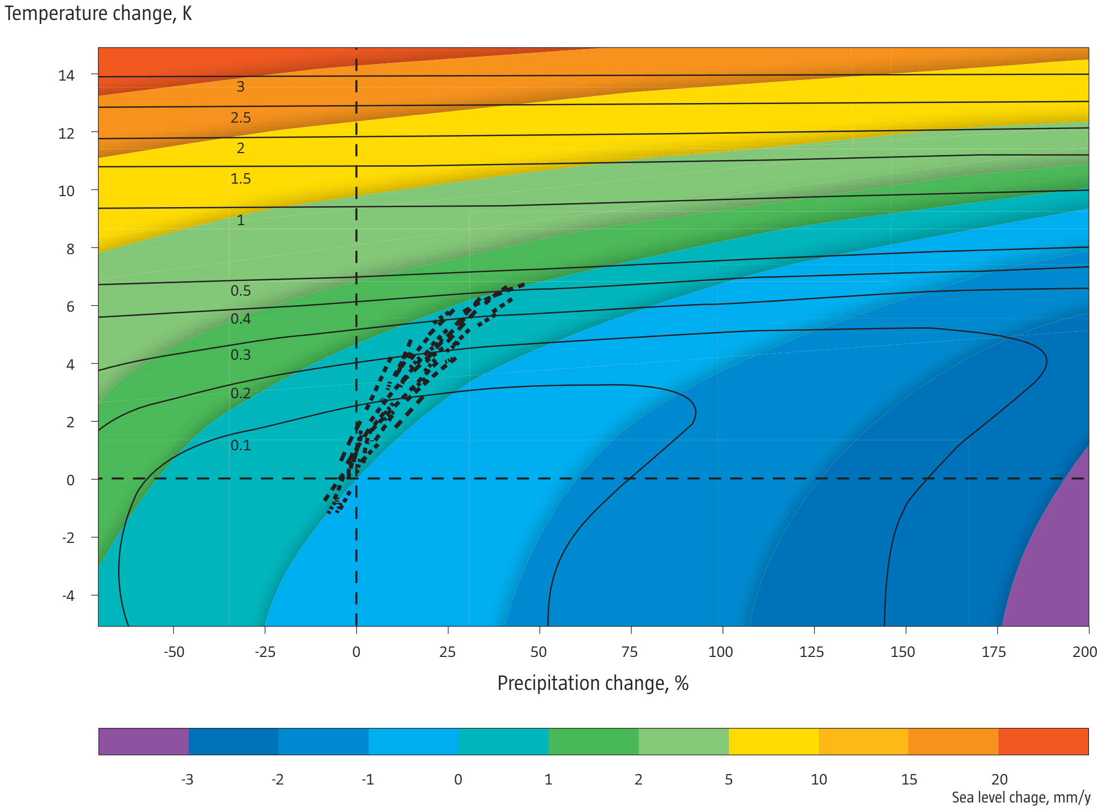


Figure 3.14 *Modeled rates of sea level change as a function of temperature and precipitation changes over Greenland. The rate of sea level rise is indicated by the color, while the contour lines denote the standard deviation in an ensemble of AOGCM forecasts. The points indicate the results of decadal temperature and precipitation shifts in future simulations from the ensemble of models. Source: Gregory and Huybrechts (2006).*

of 3 °C and 6 °C give a sea level rise of 0.72 m and 2.4 m after 1000 years. Contributions by 2100 are about 5% percent of this. Huybrechts and de Wolde (1999) found a similar rate of extreme ice sheet retreat in their most severe climate change scenario, 8 × CO_2, with an average rate of sea level rise of 60 cm per century and about 80% of the Greenland Ice Sheet melting within 1000 years under these conditions. Rates of ice loss increase non-linearly with time in these simulations, as elevation-climate feedbacks strengthen.

While the forecasts are reasonably well bounded for this century, models have a conservative bias for ice sheet response to climate change, due to missing processes in the glaciological models (see section 3.2) and a number of missing climatic feedbacks. There is no sensitivity to changing conditions at the ocean interface in current models, i.e., no capacity for outlet glacier acceleration.

Pfeffer et al. (2008) probed this by considering an extreme case where all of the marine-based outlet glaciers of the Greenland Ice Sheet double their discharge to the ocean. By taking the cross-sectional area of all marine outlet glaciers and assuming a constant ice thickness at these flux gates, Pfeffer et al. (2008) calculated that a two-fold increase in outlet glacier velocity would contribute 9.3 cm to sea level rise by 2100. This is an extreme case: All of the outlet glaciers would need to double their velocity immediately and remain at this higher rate of flow for the rest of the century, without significant thinning. This, therefore, provides a plausible upper bound on ice dynamical discharge in Greenland.

Pfeffer et al. (2008) added this term to the sea level rise associated with surface mass balance losses to give an upper bound for Greenland mass loss. However, the Greenland Ice Sheet's 'baseline state' and current estimates of future sea level rise from Greenland (e.g., Huybrechts and de Wolde, 1999; Greve, 2000; Huybrechts et al., 2004; Gregory et al., 2004) include contributions from dynamical ice losses (i.e., iceberg calving). Current models have negligible 21st century variability in this term and muted sensitivity to climatic forcing, relative to recent observations, but there is still a background level of ablation from this mechanism. Hence, it might be appropriate to consider only the extra iceberg discharge associated with a two-fold speed-up of marine outlet glaciers in placing an upper bound on Greenland's contribution to sea level rise. Using the fluxes calculated by Pfeffer et al. (2008), this gives 4.7 cm of sea level rise by

2100. If this is based, instead, on the estimated baseline ice discharge from Greenland in the early 1990s, 350 km^3/y (Table 2.2), the additional discharge from a doubling of the ice flux would give a 21st century sea level contribution of 8.8 cm. This can be added to the expected surface mass balance-driven losses of up to 10 cm to give an upper bound of ~19 cm of sea level rise from the Greenland Ice Sheet by 2100. This is not a true upper bound. The value could be higher, if climate change during this century is more severe than expected in Greenland, or as a result of additional climate feedbacks that are lacking in the current models.

3.4. Summary and outlook

The complexities of climate and ice sheet models are such that much work is still needed to refine model physics, improve model resolution, and assimilate observational data to better constrain and test simulations of the meteorology, mass balance, and dynamics of the Greenland Ice Sheet. Great strides have been made in coupling of ice sheet and climate models in recent years. Regional climate models now offer the possibility of grid-matching and a full surface energy balance calculation, to help reduce uncertainties associated with climate downscaling and some of the simplistic assumptions in current parameterizations of surface mass balance.

A number of climate forcings and ice sheet-climate processes are still missing in current models, particularly at the ice-ocean interface. This is pressing in light of evidence that oceanic forcing is responsible for some of the recent, rapid changes in the Greenland Ice Sheet mass balance and ice dynamics (Holland et al., 2008). Most fast-flow processes in ice sheets are not well represented in current models, and potential climatic excitation of fast flow, through either the ocean or atmosphere, has not been explored. This means that the parameter space for Greenland Ice Sheet response to climate change is not fully explored; models cannot give reliable bounds on this.

The kinematic approach of Pfeffer et al. (2008) probably offers the best available upper bound on the ice-dynamical contributions to mass balance losses in Greenland. For a proper dynamical simulation of these effects, the ice-dynamical processes discussed in detail in section 2.3 need to be captured by predictive models. This is important to improved estimation of the total ice sheet mass balance, given the potential for large-scale climate forcing of ice-dynamical ablation mechanisms. At present, ice sheet models are not sensitive to surface meltwater production or to ocean and sea-ice conditions, all of which have begun to change rapidly in the Arctic.

Further development of AOGCMs and regional equivalents is also needed. Little besides what has been reviewed here has been done to couple climate and ice dynamical models. Thus, more coupling experiments represent an obvious area of future work, particularly with respect to regional patterns of climate downscaling and atmospheric/mass balance feedbacks of changing ice sheet geometry.

Other climatic feedbacks on ice sheet mass balance are not fully explored, such as the impact of decreasing ice sheet albedo as the ablation zone of the ice sheet expands spatially and seasonally. Degree-day melt models typically use two melt factors, one for snow and one for ice, rather than the continuum of behavior that is evident in nature. The positive feedbacks of a wetter, lower-albedo snow surface may not be being captured in present model forecasts. More physics-based melt models (e.g., from energy balance calculations on high-resolution grids) will provide improved estimates of the threshold temperature at which the Greenland Ice Sheet switches into a state of negative surface mass balance.

These limitations make it difficult to project the extent of sea level rise that can be expected from the Greenland Ice Sheet in this century and beyond. Current models suggest 5 to 10 cm by 2100, but conservative biases in the models mean that it may be in excess of this. If climate warming excites systematic, widespread acceleration of the outlet glaciers in Greenland, sea level rise could be as much as 20 cm this century. Extreme warming scenarios (e.g., 10 °C) are needed to extract more ice than this from Greenland by 2100, although long-term ice sheet losses may result from instabilities that are triggered in the coming decades.

4. Impacts caused by changes in the Greenland Ice Sheet

4.1. Ocean circulation

Freshwater from the Greenland Ice Sheet melt as well as iceberg calving into the sea have various effects on the waters of the Arctic region (i.e., Arctic Ocean with continental shelves, Nordic Seas, and Labrador Sea) as well as on the World Ocean. The Greenland Ice Sheet run-off dilutes the marine surface water and reduces its salinity and density, thereby impacting global sea level rise and two types of ocean circulation: thermohaline circulation and estuarine circulation. All are important for the marine climate of the Arctic region although they are affected in different ways.

4.1.1. Sea level change

4.1.1.1. Global sea level change

The Greenland Ice Sheet represents, if entirely melted, a global average eustatic sea level rise of about 7 m. In a warmer climate with more precipitation, accumulation, ablation and ice discharge will increase but model studies suggest that the net balance of these factors will be towards a diminishing ice sheet (Gregory and Huybrechts, 2006). Depending on the future temperature evolution, the potential disappearance of the Greenland Ice Sheet or its transformation to a much reduced inland ice cap, will take in the order of millennia (Gregory et al., 2004).

Expected sea level changes by the end of the 21st century, due to negative surface mass balance of the Greenland Ice Sheet, can be estimated by surface mass balance models driven by climate models and are in the range 0.01 m to 0.12 m according to the IPCC Fourth Assessment Report (Meehl et al., 2007).

Since 2007, global sea level has been observed to rise at a faster rate than projected by Meehl et al. (2007). Motivated by this apparent under-estimation of sea level rise from dynamical modeling, Rahmstorf (2007) suggested an alternative semi-empirical approach where a statistical model is fitted to historical time series of global sea level and temperature. By then driving the statistical model with temperature scenarios Rahmstorf obtained total global sea level changes of 0.5 to 1.4 m for the end of the 21st century; larger than by Meehl et al. (2007). This work has, however, been criticized by several authors (Holgate et al., 2007; Schmith et al., 2007; von Storch et al., 2008; Rahmstorf, 2008). An improved semi-empirical model by Grinsted et al. (2009) estimated the likely sea level rise to be in the range 0.7 to 1.6 m at the end of the 21st century.

4.1.1.2. Regional sea level change

The increased release of freshwater from the Greenland Ice Sheet will, together with the expected increase in net precipitation over the North Atlantic area, cause changes in the salinity and temperature distribution (Curry et al., 2003), which, through a dynamic response, will cause regional changes in sea surface height. This dynamic response can be either thermohaline in nature, such as weakening of the Atlantic meridional overturning circulation, or an adjustment of gyre circulations due to changes in the density distribution.

A quantitative assessment of the effect of freshwater release from the Greenland Ice Sheet does not exist at present. Landerer et al. (2007) analyzed a standard IPCC scenario run (i.e., without melting from the Greenland Ice

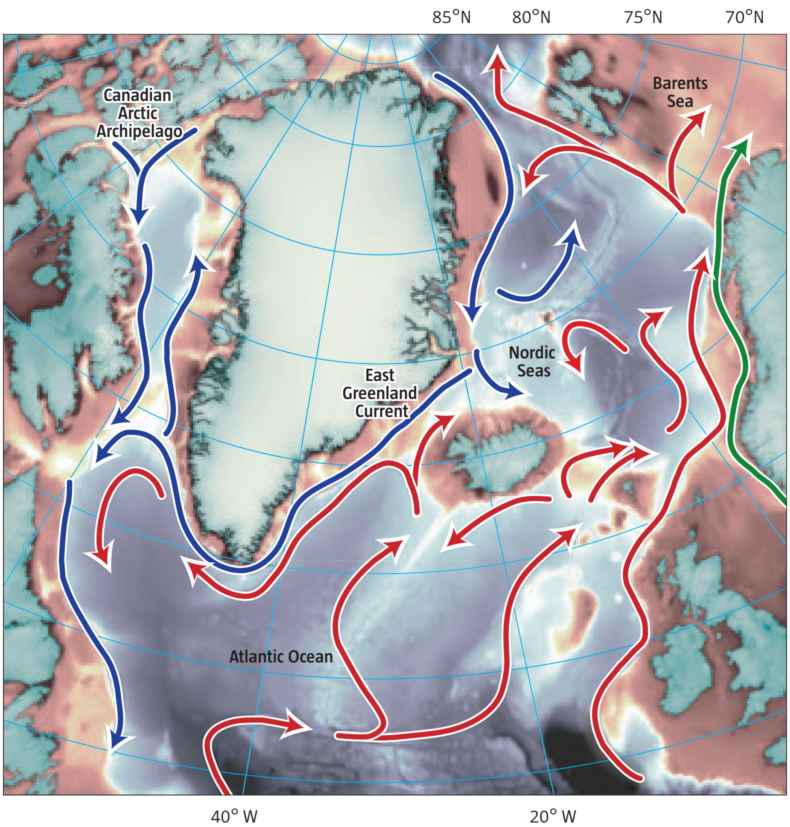
Sheet) and discovered regional signals, most prominent a sea level increase of 0.3 m in the Arctic Ocean and a minor increase (i.e., up to 0.1 m) along the Greenland coast. In the center of the subpolar gyre, Landerer et al. (2007) found a decrease in the order of 0.1 m, which could be due to either changed density distribution or to changed wind forcing, or both.

Moving mass from the ice sheet to the ocean causes changes in the gravity field (Mitrovica et al., 2001). This has remained unnoticed to date, but causes noticeable regional differences in sea level rise. Diminishing the Greenland Ice Sheet is expected to produce the largest sea level increase in the South Atlantic and the North Pacific and a decrease in the North Atlantic and around Greenland. This does not take vertical isostatic land lift into account, which takes place on a longer time scale.

4.1.2. North Atlantic circulation – deep convection and climate impact

Through the thermohaline ventilation, the Arctic region is a significant source to the North Atlantic Deep Water as part of the global thermohaline circulation. Therefore, increased run-off from Greenland has the potential to affect the circulation of the Arctic region and the global ocean.

Figure 4.1 *Ocean currents in the North Atlantic. Source: Danish Meteorological Institute.*



Off Greenland (Figure 4.1), the East Greenland Current flows southward along the shelf towards the Atlantic (Holfort et al., 2008). Together with the outflow through the Canadian Arctic Archipelago, this comprises a low-salinity surface outflow from the Arctic (Melling et al., 2008). The inflow, to compensate for the estuarine outflow, consists of warm and saline Atlantic water concentrated to the eastern part of the north Atlantic and which cannot be distinguished from the direct wind- and thermohaline-driven surface inflow to the Nordic Seas (Hansen and Østerhus, 2000). This inflow enters the Arctic in three distinct branches: a branch to the west of Iceland, a branch between Iceland and the Faroe Islands, and a branch between the Faroe Islands and Shetland (Østerhus et al., 2005; Hansen et al., 2008).

Through heat loss to the atmosphere, the inflow is cooled as it flows northward, and this increases its density. Brine rejection from freezing of seawater also has this effect whereas freshwater input decreases density. The overall effect is to separate the inflow into an upper low-density water mass and a subsurface high-density water mass that is then returned to the North Atlantic (Figure 4.2).

The low-density water mass is returned as surface outflow in the East Greenland Current and through the Canadian Arctic Archipelago, which also carries most of the Bering Strait inflow of Pacific water to the Arctic Ocean (Rudels, 1989). The high-density water mass returns to the North Atlantic through several current branches, termed 'overflows' (Saunders, 2001), and passes through a number of deep channels across the Greenland-Scotland Ridge.

In terms of volume transport, the overflows export roughly twice as much water to the North Atlantic as the surface outflows of the East Greenland Current and the Canadian Arctic Archipelago (Figure 4.2). The formation of overflow water by thermohaline ventilation involves two steps: density increase and sinking. For the density to increase substantially the water has to be in contact with the atmosphere and this occurs by heat loss in the western part of the Nordic Seas and in the Barents Sea (Simonsen and Haugan, 1996). The sink-

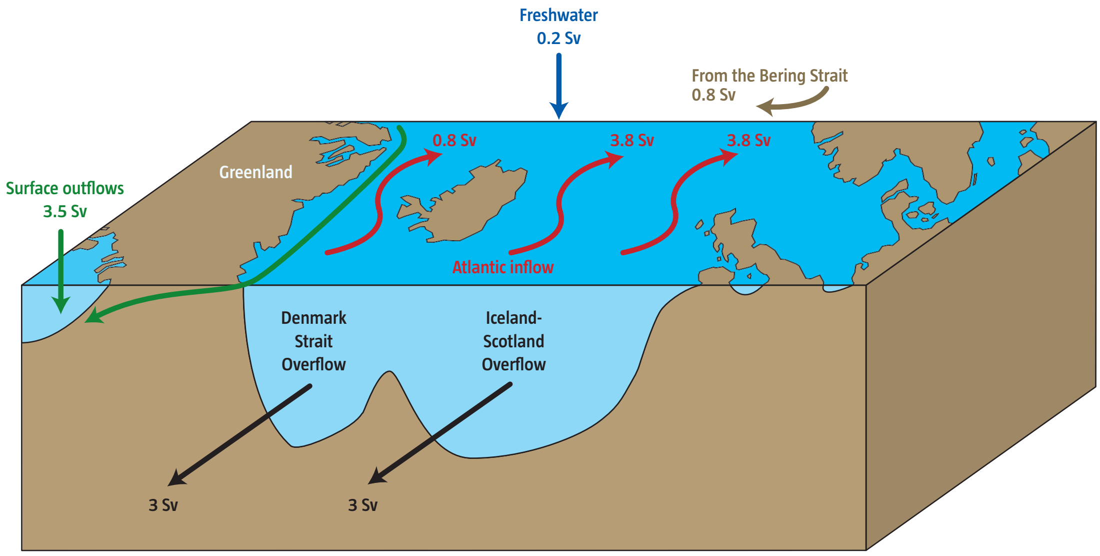


Figure 4.2 *The water budget of the Nordic Seas and Arctic Ocean with volume transports in Sv (1 Sv = 1 Sverdrup unit = 10^6 m^3/s). 'Atlantic inflow' across the lower end of the three red arrows on the graphic. Source: Hansen et al. (2008).*

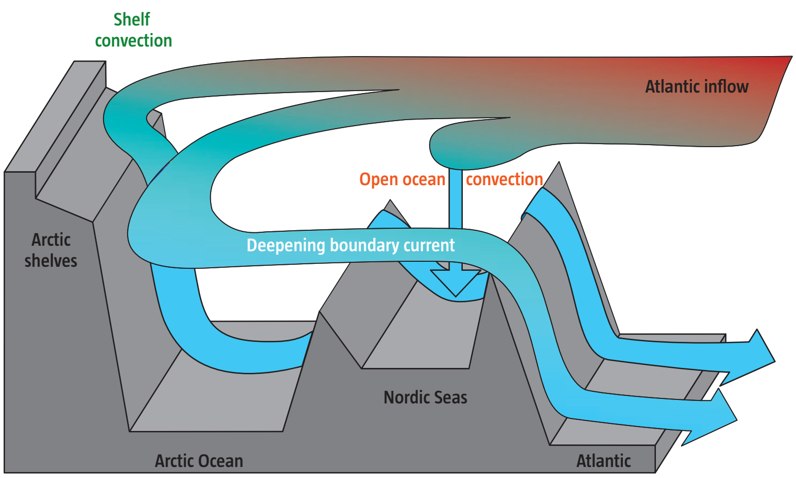
ing may occur through open ocean convection (Marshall and Schott, 1999) in the Greenland Sea and Iceland Sea, but this is only one of the mechanisms (Hansen and Østerhus, 2000) (Figure 4.3) and convection in the Greenland Sea is not likely to be the main source of overflow water (Eldevik et al., 2009).

The overflow water is cold, much of it at temperatures below 0 °C, and is relatively saline compared to other cold water masses in the ocean. Thus, it is one of the densest water masses found anywhere in the World Ocean but, immediately after crossing the Greenland-Scotland Ridge, it mixes strongly with, and so entrains, ambient upper ocean water masses (Dickson and Brown, 1994; Østerhus et al., 2008). The resulting water mass is warmer and less dense, but also has a larger volume transport. Most of this water flows into the Labrador Sea where it contributes to the North Atlantic Deep Water together with the Labrador Sea Water formed locally by convection (Lazier, 1980; Yashayaev et al., 2008). North Atlantic Deep Water is thus from three sources: overflow water, Atlantic water entrained into the overflows, and Labrador Sea Water. Formation rates for some of these are debated (Haine et al., 2008), but there seems to be general agreement that they are of similar magnitudes.

North Atlantic Deep Water is carried south by the deep branch of the Atlantic Meridional Overturning Circulation (AMOC) and is the main deep water source of the World Ocean, complemented by Antarctic Bottom Water formed by dense outflows from the Antarctic shelf regions. The AMOC is associated with heat transport on inter-hemispheric scales towards the ventilation areas where heat is released to the atmosphere. This transport of heat by the ocean is modest compared to the atmospheric heat transport (Trenberth and Caron, 2001) and its importance for climate has been challenged (Seager et al., 2002).

However, convincing evidence suggests that, through climate interactions, this ocean transport indeed plays a disproportionate role for fundamentals of the global climate (Rhines and Häkkinen, 2003; Rhines et al., 2008). This also finds support from tailored 'what if' climate model experiments, where artificial shutdown of the North Atlantic Deep Water production has been shown to strongly cool the regional climate of the Nordic Seas and Barents Sea but with cooling evident over the entire Northern Hemisphere (e.g., Vellinga and Wood, 2002; Sutton and Hodson, 2005).

Figure 4.3 *Schematic illustration of some of the processes that convert Atlantic inflow to overflow. Source: Danish Meteorological Institute.*



Apparently, this has occurred in other climatic periods, for example, during the deglaciation period where reduced thermohaline ventilation was likely to have been triggered by increased extreme freshwater supply from the melting glaciers to the ventilation areas (Tarasov and Peltier, 2005). Changes in the AMOC also have indirect impacts on the global climate. The formation of deep water and its subsequent spreading is crucial for maintaining not only the density but also the chemical structure and the biological conditions characterizing the World Ocean (Schmittner, 2005) with feedbacks also on atmospheric chemistry and climate.

A number of confounding factors, however, make quantifying the impacts of increasing run-off from the Greenland Ice Sheet difficult. First, run-off is not the only freshwater source in the Arctic region. In the present-day climate, total Greenland Ice Sheet run-off (0.018 Sverdrup [1 Sv = 10^6 m^3/s]; Dickson et al., 2007) is almost an order of magnitude smaller than the estimated freshwater (including sea ice) transport southward through Fram Strait (0.15 Sv; Carmack et al., 2008), which flows through the areas into which the Greenland Ice Sheet run-off is released. It is also much smaller than the freshwater component of the Canadian Arctic Archipelago outflow (Carmack et al., 2008).

Various changes in the freshwater cycle of high latitude lands and oceans have been reported for the period since the 1950s (Peterson et al., 2006). Changes have been widespread and perhaps dramatic in the case of the Great Salinity Anomaly of the 1970s but, by attribution, anomalous glacial melt has played a minor role, contributing only about 5% to the observed ocean freshwater storage in the North Atlantic and Arctic region. In the future, the Greenland Ice Sheet run-off may well, in relative terms, increase more rapidly than the other sources but even in the most extreme case, the projected peak rate of melting estimated from a coupled climate-ice sheet model (0.06 Sv; Ridley et al., 2005) remains considerably smaller than the contribution from other sources today (Dickson et al., 2007). In idealized anthropogenic climate change simulations, absolute changes in atmospheric moisture transport are also found to exceed changes in run-off (Mikolajewicz et al., 2007a; Vizcaíno et al., 2008), which tends to downplay the isolated role of the Greenland Ice Sheet run-off on ocean circulation.

Second, the effects of increased freshwater supply depend on other factors. The intensity of the estuarine circulation is determined mainly by the rate of mixing between freshwater and seawater thus depending primarily on stratification and winds (Stigebrandt, 2000). For the thermohaline ventilation, the amount of freshwater that can reach an area of deep ventilation is a crucial factor. Most of the Greenland Ice Sheet run-off should be able to reach the ventilation areas in the Labrador Sea, but only run-off from the east coast of Greenland, north of Denmark Strait, can directly affect the offshore regions of the Nordic Seas where overflow waters are produced.

In the present-day climate, even this run-off seems to affect these areas only weakly because most of it is carried south through Denmark Strait by the East Greenland Current (Dickson et al., 2007; Holfort et al., 2008), depending on the wind conditions over the East Greenland Current (Stigebrandt, 2000), which may well change. Although part of the run-off from the ice sheet may initially bypass areas of ventilation, it is recirculated in the sub-polar North Atlantic and will, on decadal scales, dilute the entire sub-polar North Atlantic and spread to the Nordic Seas (Curry et al., 2003; Curry and Mauritzen, 2005; Peterson et al., 2006). The recent freshening has indeed been traced into the deep overflows from the Nordic Seas (Dickson et al., 2002).

In anthropogenic climate change projections, the effect on the AMOC of increased freshwater supply to the ventilation areas in response to an intensified atmospheric hydrological cycle (Cubasch et al., 2001) may also be partly or completely compensated for by increased salinity of the Atlantic inflow. With enhanced evaporation at low latitudes, the source waters in the sub-tropical North Atlantic are becoming more saline (Curry et al., 2003) and this effect has been shown to stabilize the AMOC (Latif et al., 2000). The pathways of subtropical waters toward the ventilation areas, however, are dependent on the circulation of the subpolar gyre (Hátún et al., 2005), which again depends on Labrador

Sea convection (Häkkinen and Rhines, 2004). No such direct compensation, however, will be in effect considering an increased Greenland Ice Sheet run-off resulting from a negative mass balance of the ice sheet, or anomalous freshwater outflow from the Arctic Ocean due to retreating Arctic sea-ice cover.

Though these freshwater fluxes will certainly add to AMOC weakening under anthropogenic climate change, this effect, including changes due to intensification of the hydrological cycle, will possibly be masked by the direct effects of atmospheric warming on the AMOC. In a comparison of eleven different climate models, Gregory et al. (2005) found weakening of the AMOC in idealized global warming experiments to be induced mainly by changes in surface heat flux and to a lesser extent by changes in freshwater fluxes.

In the IPCC's Fourth Assessment Report, Working Group 1 reviewed the status of climate model projections for the AMOC (Figure 4.4) and found it 'very likely' that the strength of the AMOC will decrease in the 21st century (Meehl et al., 2007). The working group did not expect the AMOC weakening to lead to cooling over Europe and they considered an abrupt transition during the course of the 21st century to be very unlikely. However, none of the model studies above take the effects of ice sheets into account.

Very few studies exist where general circulation models have been coupled to dynamic ice sheet models (Fichefet et al., 2003; Ridley et al., 2005; Winguth et al., 2005; Mikolajewicz et al., 2007a,b; Vizcaíno et al., 2008). It may also be argued that only in the recent study by Mikolajewicz and co-workers (2007a) was the coupling complete. This model alone features a realistic climate allowing reasonable simulation of ice sheets. Nevertheless, these models yield similar meltwater input rates and most, although not all, reveal only a moderate effect of enhanced Greenland Ice Sheet run-off on the AMOC. Thus, it would be premature to conclude on the effects of the Greenland Ice Sheet run-off on the thermohaline circulation on the basis of these few studies.

Whereas fully coupled models will be required to convincingly project future melt rates of the Greenland Ice Sheet and project future global sea level rise, the freshwater impact on the AMOC by other and probably more dominating sources may be studied independently. This understanding is crucial for assessing the stability of the AMOC under anthropogenic climate change where a large spread in projected weakening for the 21st century among available models (Gregory et al., 2005) indicates a significant model uncertainty. In particular, differences in the projected changes in the hydrological cycle explain differences in the projections of the AMOC (Cubasch et al., 2001; Vellinga et al., 2008). This is one of the key questions to be addressed in the FP7-funded research project THOR (ThermoHaline Overturning – at Risk?) that was initiated in 2008.

It has been debated whether the projected AMOC decrease has already been initiated. Bryden et al. (2005) suggested that the Meridional Overturning Circulation at 24° N

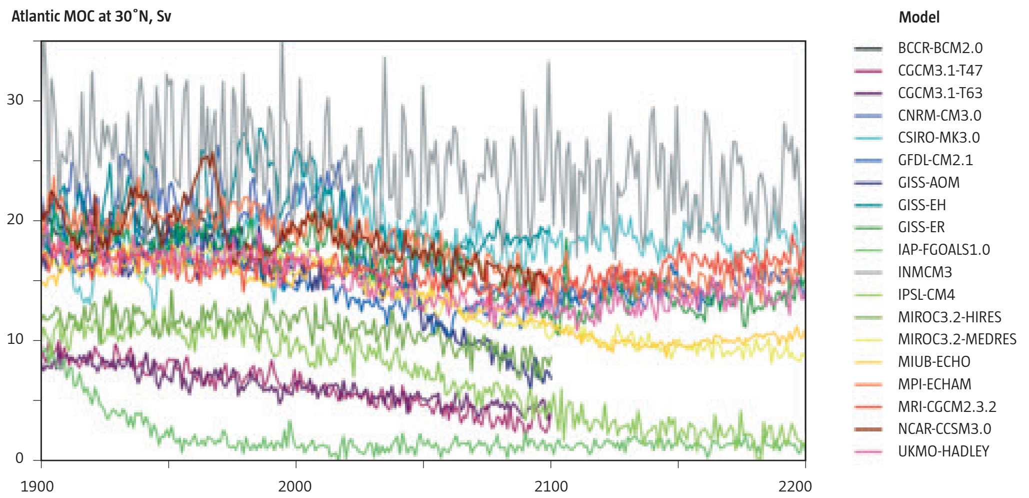


Figure 4.4 *Evolution of the Atlantic Meridional Overturning Circulation at 30° N in simulations with the IPCC suite of coupled climate models. Source: Meehl et al. (2007).*

in the Atlantic Ocean decreased by 30% between 1950 and 2000 with most of the decrease being due to declining overflows. This supports the suggestion of a similar, but smaller, decrease in the overflow transport through the Faroe Bank Channel (Hansen et al., 2001).

Direct measurements have demonstrated a high stability of this flow between 1995 and 2005 (Hansen and Østerhus, 2007) and Olsen et al. (2008) found no weakening of either this overflow branch or the total overflow transport from 1948 to 2005 in a study combining measurements with an ocean model. Concerns have also been raised about the methodology of the Bryden et al. (2005) study (Cunningham et al., 2007; Baehr et al., 2007; Wunsch, 2008).

Conversely, there is evidence in the Labrador Sea of a weakened convection since the mid-1990s, but the formation rate of Labrador Sea Water exhibits large variability (Haine et al., 2008) and attribution of this weakening to anthropogenic causes would be premature and, on the present observational basis (Peterson et al., 2006), it seems unlikely that enhanced Greenland Ice Sheet run-off has played a prominent role.

4.1.3. Sea ice

Sea-ice formation depends on the stability of the surface layer of the ocean. Changes to the surface layer incurred by increased meltwater run-off from the Greenland Ice Sheet in a warmer climate could therefore potentially influence sea-ice production.

4.1.3.1. Freshwater flux

An increased freshwater flux may lead to increased stability of the upper ocean, which could precondition greater sea-ice production. On the other hand, ice cover of the marginal seas around Greenland has reduced during the present global warming period according to satellite data from the period 1978 to 2008. The total effect of higher annual mean air temperatures and increased freshwater flux is not well predicted.

In the Greenland Sea, the role of local sea-ice production and advection was shown to potentially precondition the ocean for deep convection (Visbeck et al., 1995; Toudal and Coon, 2001). Brine rejection during sea-ice formation destabilizes the ocean and advection of the ice out of the area removes excess freshwater. After a period of ice formation and advection, the density gradient will be eroded and further cooling can generate deeper convection. With sufficient cooling, this mechanism can remove excess freshwater from the surface of the Greenland Sea and counteract an increased freshwater flux from Greenland in relation to deep water formation. However, as the mechanism is reliant on subsequent cooling during winter, it is not clear if it will prevail in a warmer climate.

Few examples of coupled global climate model predictions of the response of sea ice to an increased freshwater flux from the Greenland Ice Sheet have been published. Fichefet et al. (2003) showed indications of a possible feedback mechanism where a breakdown of the thermohaline circulation leads to a significant cooling over eastern Greenland and northern North Atlantic waters, amplified by the sea-ice albedo feedback related to an expansion of the sea-ice cover in the Greenland-Iceland-Norwegian seas. The cooling subsequently reduces the freshwater flux from Greenland.

Swingedouw and Braconnot (2007) found the strongest sea-ice effect to be associated with a marked cooling of the Barents Sea region resulting from a slowdown of the thermohaline circulation. They further showed that the impact of AMOC weakening on surface temperature is important in the Northern Hemisphere mid- and high latitudes, although atmosphere heat transport compensates part of the oceanic heat transport weakening.

4.1.3.1.1. Feedback of less ice on the state of ice shelves

The reduced summer ice cover at the Arctic Ocean margins (e.g., the Greenland Sea and Barents Sea) exposes the coastline of these regions to the influence of ocean swells. This may cause increased coastal erosion but may also directly influence the stability of floating glaciers, such as Storstrømmen and 79-fjorden, in northeast Greenland.

Reeh et al. (2001) studied the interaction between sea ice and glacier, in the case of the floating tongue of the 79-fjorden glacier, northeast Greenland (79°30’ N, 22° W). The

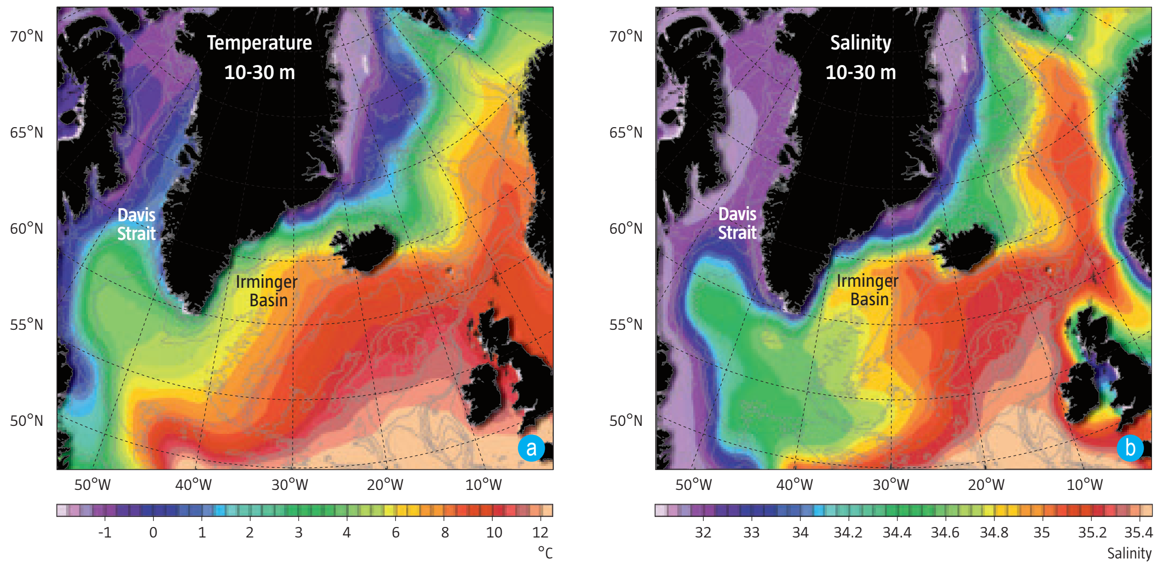


Figure 4.5 *Annual mean surface (10-30 m) temperature (a) and salinity (b) from World Ocean Atlas 2001. Note the change in scales at 4 °C for temperature and 34 for salinity. Source: Conkright and Boyer (2002). Public information.*

authors used information from glaciological and geological studies, expedition reports, aerial photographs and satellite imagery to document the position of the glacier front and fast ice conditions on millennial to decadal time scales. The study indicates that stability of the floating glacier margin is dependent on the presence of a protecting fast-ice cover in front of the glacier. In periods with a permanent fast-ice cover, no iceberg calving occurs, but after fast-ice break-up the glacier responds with a large ice discharge activity, whereby several years worth of accumulated glacier ice flux suddenly breaks away.

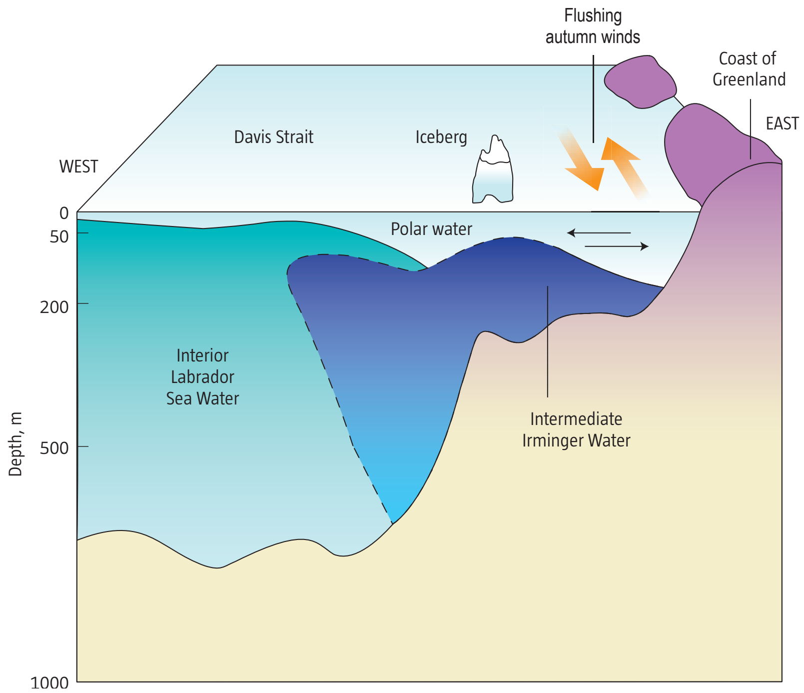
Climate-induced changes in sea-ice conditions in the Arctic Ocean with seasonal break-up of the nearshore fast ice could lead to the disintegration of floating glaciers. The present dominant mass loss by bottom melting would then be largely overtaken by grounding line calving of icebergs and the local influx of freshwater from north Greenland glaciers to the sea would be reduced and local iceberg production would increase.

4.1.4. Greenland coastal currents, including fjords, circulation and stratification

The ocean currents around Greenland are part of the cyclonic sub-polar gyre circulation of the North Atlantic and the Arctic region (Figure 4.1). The surface waters in southeastern and western Greenland are dominated by two very different water masses: warm and saline Irminger Water, a side branch of the North Atlantic Current, and cold and low-saline Polar Water originating from the Arctic Ocean.

These water masses meet in the northern Irminger Basin and in the Denmark Strait (Figure 4.5). The main branch of the Irminger Current turns west towards East Greenland where it meets the Polar Water, which is transported southward along East Greenland within the East Greenland Current. The strength of these two currents determines the hydrographic conditions around southeastern and West Greenland. The two water masses flow side by side forming large meanders at the front where intensive mixing takes place. Polar Water is found at the surface on the

Figure 4.6 *Sketch of the water masses off West Greenland in the Davis Strait region. Source: Valeur et al. (1997).*



continental shelf whereas Irminger Water is found over the continental slope and partly below the Polar Water on the deeper parts of the continental shelf. As they round Cape Farewell the Irminger Water subducts under the Polar Water (Figure 4.6).

Adjacent to the coast, the surface water is modified by run-off from Greenland whereby the stratification in the water column increases. During summer, this forms a freshwater baroclinic jet, named the East Greenland Coastal Current, which is only a few tens of kilometers wide but with maximum velocities exceeding 1 m/s (Bacon et al., 2002; Wilkinson and Bacon, 2005). The run-off originates from melting of the Greenland Ice Sheet as well as from precipitation. Kiilsholm et al. (2003) showed that by far the greatest value (i.e., precipitation minus evaporation) over Greenland is found in southeast Greenland, which could be the main driver behind the East Greenland Coastal Current. If so, the presence of significant, baroclinic coastal currents is limited to southwest Greenland waters, but this is still an open question, especially due to the lack of observations in northeast Greenland waters.

Recently, Bacon et al. (2008) speculated that the continuous melting of sea ice combined with a general wind field from the north could be the organizing principle for the East Greenland Coastal Current. A spot value of the freshwater transport (0.06 Sv) calculated by Bacon et al. (2002) would equate to 30% of the total annual Arctic freshwater gain. However, it seems unlikely that the East Greenland Coastal Current will keep its momentum throughout the year, as both run-off and air temperature have a clear annual cycle.

The classic time series of temperature and salinity on top of Fylla Bank (Figure 4.7) describes the observed hydrographic conditions off West Greenland. For the temperature curve, the most striking feature is the three cold periods centered around 1970, the early 1980s and the 1990s; all three related to atmospheric forcing (Buch et al., 2004). The cooling in the late 1960s was associated with a large pulse of freshwater that left the Arctic Ocean through Fram Strait and rapidly moved southward in the East Greenland Current and later northward in the West Greenland Current, and it was easily recognized in the Fylla Bank salinity (Figure 4.7). This event has been labeled the Great Salinity Anomaly (Dickson et al., 1988; Belkin et al., 1998), and Curry and Mauritzen (2005) estimated the freshwater flux to about 0.07 Sv. This value is in agreement with model simulations preformed by Haak et al. (2003) and Olsen and Schmidt (2007).

Interestingly, the freshwater anomaly associated with the Great Salinity Anomaly is similar to the expected freshwater anomaly due to the forecast maximum melting of the Greenland Ice Sheet. In other words, the natural variability observed today in freshwater fluxes is similar to the expected additional contribution from the Greenland Ice sheet. However, events similar to the Great Salinity Anomaly are expected in the future and more important, as noted by Vizcaíno et al. (2008), the atmospheric moisture transport is expected to increase by one order of magnitude higher than the freshwater increase due to melting of the Greenland Ice Sheet.

In a future climate with increased run-off due to increased precipitation and melting of the Greenland Ice Sheet, the East Greenland Coastal Current would be expected to increase. In a study of climate changes in Greenland waters and surrounding seas, high-resolution regional models were applied to calculate a scenario of the climate for the period 1950 to 2050 (Stendel et al., 2007; Kliem et al., 2007). The regional models were used

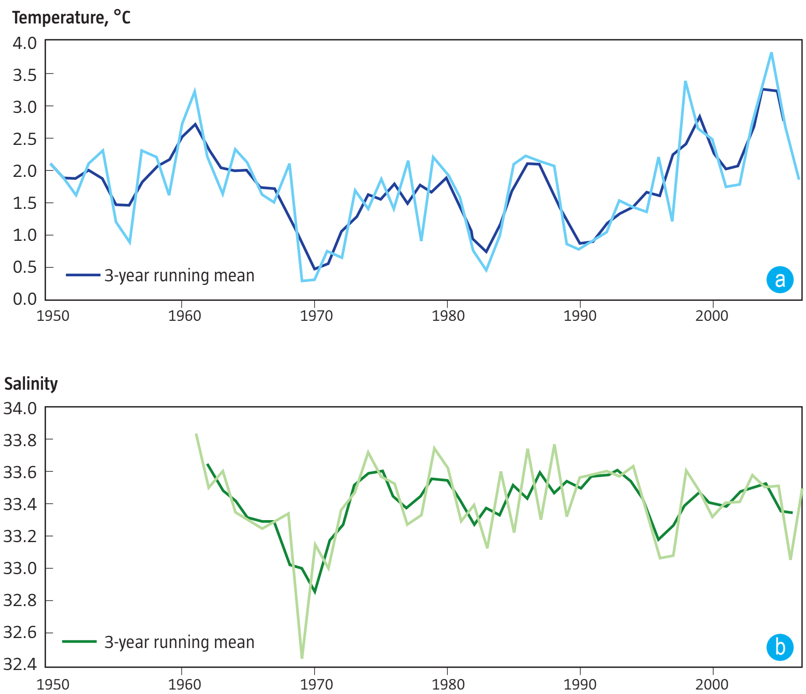


Figure 4.7 *Time series of mean temperature (a) and mean salinity (b) on top of Fylla Bank (0-40 m) in the middle of June for the period 1950-2007. Updated from Ribergaard (2007).*

to downscale a global coupled atmospheric-ocean general circulation model.

The ocean model was forced by the global circulation model simulation at the lateral boundaries and by a regional climate atmospheric simulation at the surface. Run-off was not included due to melting of the ice sheet directly in the simulations. However, an indirect effect was included by restoring the surface salinity towards the surface salinity of the global circulation model. Still, Vizcaíno et al. (2008) found that the increase in freshwater flux from the Greenland Ice Sheet is one order of magnitude smaller than the increase in the atmospheric moisture transport under anthropogenic greenhouse gas forcing.

The slight increase in temperature (see Figure 4.8) is due to a general atmospheric warming and not caused by the retreat of sea ice, as it is rarely present at Fylla Bank. Moreover, the simulations show a clear freshening at Fylla Bank that is most likely to be a result of the increased hydrological cycle. Note that the major salinity fluctuations are due to major GSA-like events. These natural variations are an order of magnitude higher than the general freshening.

4.1.4.1. Fjords

Deep fjords are characteristic elements of the Greenland coastline. They constitute a key element of the land-ocean interface as they connect the ice sheet to shelf waters around Greenland. In addition to tide, run-off and local winds are the main driving forces acting on the upper water masses in a fjord system (Svendsen et al., 2002). Frequently, there is a large climate gradient from the inner parts of the fjords to the outer parts. Also, meltwater from the Greenland Ice Sheet or rivers that drain into the inner parts of the fjords also causes a strong salinity gradient along the fjord axis. Typically, meltwater discharge takes place over a few months when temperatures exceed 0 °C and the discharge often occurs in pulses (Hasholt et al., 2008).

Where a glacier meets ocean water, mass is not just lost by the visually spectacular calving mechanisms. At ice shelves and floating tongues it is known from mass conservation considerations that sub-ice melting can occur at rates exceeding 10 m/y (Rignot and Jacobs, 2002; Rignot and Steffen, 2008). Also, Motyka et al. (2003) proposed a model of freshwater-driven fjord convection that brings in warm saline bottom water and transports cold fresh surface water out of the fjord. Recent observations of high water temperatures (3 to 10 °C) in the immediate vicinity of tidewater glaciers (Motyka and Truffer, 2007; Ritchie et al., 2008; Rysgaard et al., 2008) indicate that the ocean water around Greenland has a large heat content and a corresponding melting potential. Changes in the ocean have been observed to coincide with glacier changes in this system and elsewhere in Greenland (e.g., Howat et al., 2008; Holland et al., 2008).

Numerical simulations of the hydrographic conditions under a warmer climate scenario exist from a northeast Greenland fjord (Young Sund, 74° N) that is connected to the Greenland Ice Sheet via river run-off. However, no large floating glaciers are present in the fjord system. In a warmer climate, the run-off from land would increase and this would locally change the circulation and influence the biological production (Rysgaard et al., 2003; Bendtsen et al., 2007). The sensitivity of Young Sund to changing run-off shows that in the case of no run-off, the mixed layer thickness is about 9 to 11 m during the summer period due to a weak pycnocline in the upper part of the water column (Figure 4.9).

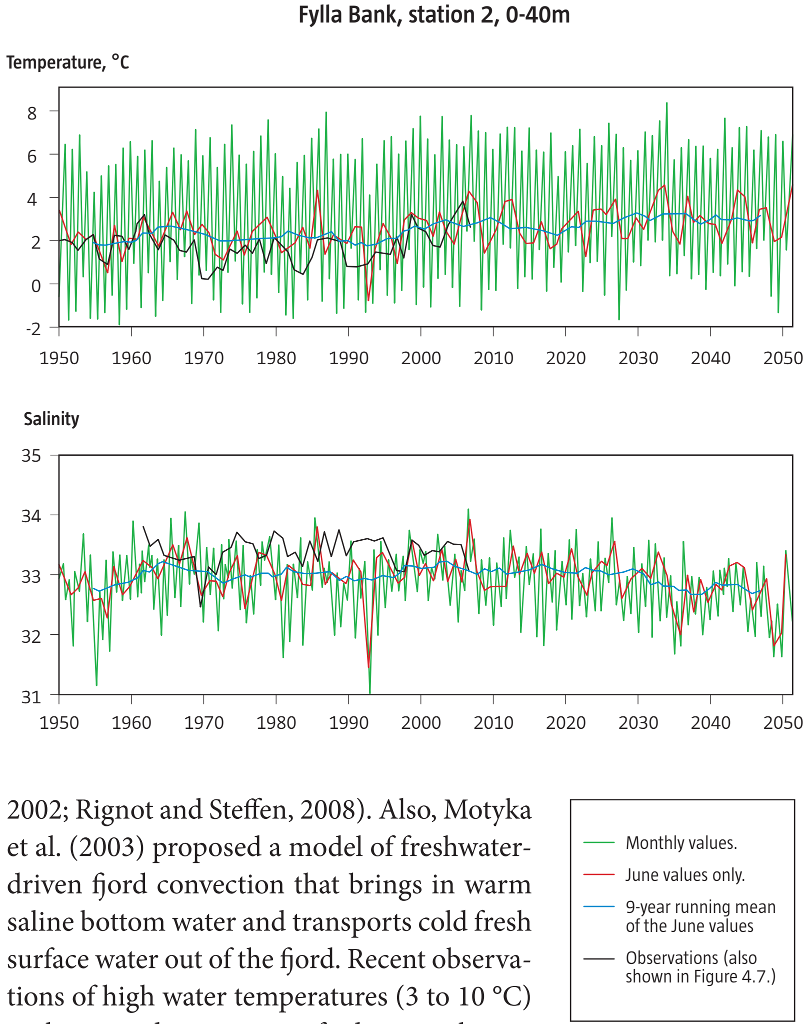


Figure 4.8 *Modeled time series of temperature and salinity on top of Fylla Bank (0-40 m). The atmospheric component was run without data assimilation. Therefore, only the overall size of the fluctuations and the mean values are comparable, but not individual years. Source: Kliem et al. (2007).*

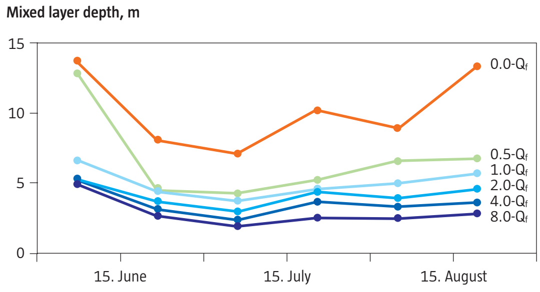


Figure 4.9 *Model solutions from a sensitivity study of the relation between the mixed layer depth in the central part of Young Sound and the freshwater discharge (Qf) where the present-day discharge (corresponding to 1.0) is scaled between 0 and 8. Source: adapted from Bendtsen et al. (2007).*

At a run-off corresponding to only 50% of the normal run-off, the halocline is quickly established, and the mixed layer is about 0.7-1.6 m deeper in July and August than under present day conditions where the mixed layer depth is about 4.0-5.2 m in July and August. At a run-off two times greater than in the reference case, the mixed layer becomes 0.3-1.1 m shallower during the summer season than in the reference case, and in the extreme case when run-off is 8 times greater, mixed layer depth decreases to about 2.2-3.7 m during the summer season.

These sensitivity studies show the importance of the run-off for controlling the depth of the mixed layer. Without a run-off, the mixed layer is controlled solely by wind-induced mixing and buoyancy fluxes at the surface. Even at a moderate run-off of only 50% of the present level, a freshwater controlled mixed layer is established quite early by the end of June, and is only slightly deeper than in the reference case. At the other extreme, when run-off is large, the mixed layer depth is controlled by the strength of the surface forcing on the system, i.e., wind and air temperature. Thus, the case with no run-off puts an upper limit on the mixed layer depth of about 11 m in the fjord during the summer season, and, correspondingly when run-off is 8 times the current level this puts a lower limit on the mixed layer depth of about 2 m.

In a future warmer climate scenario, the run-off period might increase significantly, and this would prolong the period in which the surface conditions are controlled by freshwater discharge. However, this would not be expected to decrease the mixed layer depth significantly below the solutions shown in Figure 4.9, because the balance between run-off and atmospheric forcing is established within a few weeks.

Whereas mixed layer depth would change little due to increased freshwater run-off, it may alter significantly the transport of saltwater into the fjord from the Greenland Sea due to increased estuarine circulation. Hence, in a future climate scenario primary production is expected to increase due to a combination of increased nutrient import to the fjord and improved light availability as a result of reduced sea ice (Rysgaard et al., 1999, 2003; Bendtsen et al., 2007).

4.1.5. Summary

Melting of the Greenland Ice Sheet will affect the physical marine environment on different scales, ranging from local scales to global scales. The most visible effect, which also is the one that attracts most attention in public debate, is the rising water level of the world ocean. The latest assessment from the IPCC (2007) predicts the water level to rise between 0.18 and 0.59 m by 2100. This forecast has been questioned by several scientists, however, because recent data show that global warming seems to accelerate the melting of the Greenland Ice Sheet, which will lead to an increased rise in sea level. Estimates are still rather uncertain but recent studies project a sea level rise of 0.5 to 1.6 m by 2100.

More locally, the most visible effect of the melting ice sheet on the marine environment may be the changes in salinity conditions especially in the Greenland fjords and coastal waters. These may result in stronger gradients between the relatively low-saline surface water and the more saline water below; stronger gradients will have a great impact on primary production.

Although the increased run-off of meltwater from Greenland will reach the areas of deep water formation in the North Atlantic (i.e., the Greenland Sea and Labrador Sea), the freshwater volume from the Greenland Ice Sheet is around one order of magnitude less than that from other freshwater sources (i.e., the Arctic Ocean), and so it seems to be of minor importance to changes in the global thermohaline circulation although this has not yet been fully proven.

4.2. The marine environment and marine ecosystems

Almost no research has focused on the effects of increased melting of the Greenland Ice Sheet on marine ecology. Present knowledge on the impact on marine ecology due to climate change in the Arctic region is mainly related to the melting or reduction of sea ice. As a result this section mainly describes changes in the marine ecology related to changes in sea-ice cover, but does serve as an introduction to the likely changes in marine ecology due to climate change in the Arctic region.

4.2.1. Primary and secondary production

Marine primary production is the basis of the Arctic food web. The magnitude of the annual marine primary production depends on the combination of sea-ice coverage, solar radiation, availability of nutrients and CO_2, and mixed layer depth. All these factors are directly or indirectly affected by climate change and changes in the Greenland Ice Sheet. The seasonality in these factors drives the characteristic annual patterns observed in the plankton succession in the sea. The seasonal stratification of the water column triggers the primary production by keeping the phytoplankton in the nutrient-rich and illuminated part of the water column, thereby enabling the spring phytoplankton bloom (Sverdrup, 1953). The initial concentration of 'new nutrients' supplies the spring bloom. Later, during summer, primary production is lower and dependent on nutrients regenerated through heterotrophic activity in the surface layer.

The spring phytoplankton bloom is the single most important event determining the secondary production capacity of Arctic marine food webs. The onset of the spring bloom varies between years depending on the duration of ice cover and the meteorological conditions. Measurements of annual pelagic primary production in the Arctic show a strong positive correlation between the open-water period and the magnitude of the annual primary production (Rysgaard et al., 1999). This correlation occurs at the local/regional scale and is driven by differences in the duration of the growth season. Large-scale differences in primary production across the Arctic are suggested to be primarily dictated by nutrient supply regimes induced by vertical mixing by winds and convection (Tremblay and Gagnon, 2009).

A warming climate may result in reduction of the Arctic sea-ice cover, a freshening of the surface layer due to increased melting and precipitation, and consequently a stronger stratification of the surface layer. This would impede the upward supply of nutrients thereby decreasing productivity. However, a prolonged exposure to light may enhance pelagic productivity (Tremblay and Gagnon, 2009). It is likely that the onset of the spring bloom will be more predictable, because the bloom will develop according to the light cycle rather than ice break-up. Also, earlier and stronger stratification of the surface layers will increase their heat trapping capacity. A direct effect of higher surface water temperatures will be a general acceleration of biological rates and, consequently, more nutrients and organic matter will be recycled in the surface layer.

A key group in the Arctic marine ecosystem is copepods of the genus *Calanus* (Figure 4.10). These crustaceans dominate the zooplankton (Madsen et al., 2001; Møller et al., 2006; Nielsen et al., 2007) and many fishes, seabirds and marine mammals are directly or indirectly dependent on copepods (Karnovsky et al., 2003; Laidre et al., 2007). Three species of *Calanus* co-exist in the Arctic. *C. hyperboreus* and *C. glacialis* are true Arctic species while *C. finmarchicus* is a temperate species associated with Atlantic waters. *C. finmarchicus* is however, abundant along the coast of Greenland (Hirche et al., 1994; Madsen et al., 2001; Nielsen et al., 2007).

The life cycle of *Calanus* is highly adapted to the Arctic environment. They spend the winter in hibernation in the deep waters. At springtime, when the phytoplankton bloom develops, they ascend to the surface water to exploit the abundant food supply. *C. hyperboreus* spawn in winter, and the nauplii larvae are thus ready to exploit the phytoplankton spring bloom. *C. glacialis* spawn just before spring bloom, while *C. finmarchicus* need to feed before they start reproducing. After the phytoplankton spring bloom, reproduction

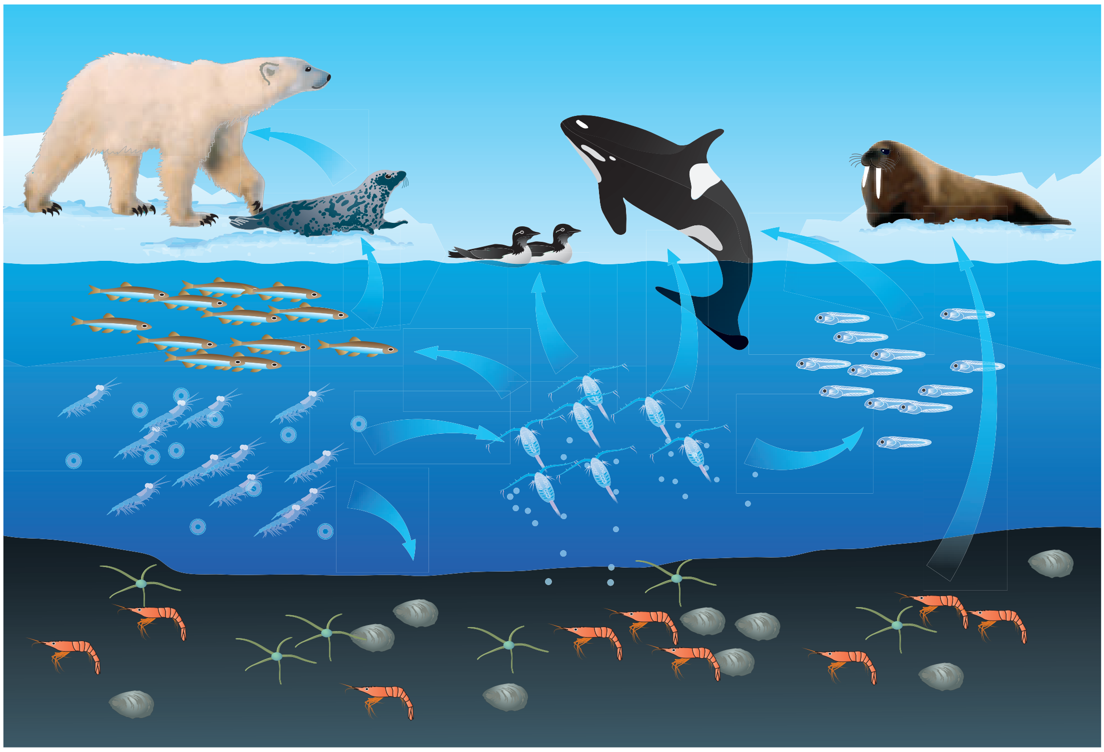


Figure 4.10 *The Arctic marine food web is based on and fuelled by lipids synthesized by diatoms and accumulated in the* Calanus*-dominated zooplankton. Source: Pers. comm. Torkel Gissel Nielsen, University of Aarhus.*

ceases and the copepods start to build up lipid reserves for the following winter and reproduction next spring, before descending to start hibernation around midsummer. When *Calanus* have left the surface layers, secondary production becomes dominated by smaller copepod species with limited lipid storage and unicellular zooplankton; the protozooplankton (Levinsen and Nielsen, 2002).

Through the food chain, the lipid level increases from 10-20% of dry mass in phytoplankton to 50-70% in *Calanus*. This lipid-based energy flux is one of the primary reasons for the large stocks of fish, birds and marine mammals in Arctic waters (Falk-Petersen et al., 2006). Noteworthy differences in lipid content are found between the different species of *Calanus*, reflecting their degree of adaptation to the unpredictable Arctic environment. *C. hyperboreus* is the largest species, containing 25 times more lipid than *C. finmarchicus*, while *C. glacialis* contain 10 times more lipid than *C. finmarchicus* (Scott et al., 2000).

The lipid reserves allow the copepods to withstand long periods of starvation (Lee et al., 2006). Their production is, however, heavily dependent on the timing of their ascent in relation to the spring bloom. A changing climate may therefore have profound effects on the composition and production of the copepod community. It has been suggested that a later break-up of the sea ice, resulting in a later spring bloom, could lead to copepods experiencing periods of starvation or low food quality, because they ascend too early. On the other hand, an earlier spring bloom could sediment to the bottom before the copepods arrive at the surface (Hansen et al., 2003). The success of the different copepod species in a changing climate depends on the capability of adapting to the concurrent changes in temperature and food conditions (Ringuette et al., 2002).

The temporal match of the spring bloom and copepod ascent is also crucial for the proportion of the phytoplankton that does sediment to the bottom. However, even if the copepods do match the spring bloom timing, their large fecal pellets also will sink, and a prolonged productive season will lead to increased total sedimentation. Plus, a prolonged open water period will augment the relative importance of food webs based on the smaller copepods and protozooplankton (Levinsen and Nielsen, 2002), with a much lower level of lipids than a *Calanus*-based food web (Figure 4.11).

Thus, a warming climate is likely to cause changes in the relative contribution of the *Calanus* species, and this may have profound effects on the organisms dependent on them.

An example of an organism directly affected by changes in the copepod community is the high Arctic alcid, Little Auk (*Alle alle*). Little Auk is dependent on the large lipid-rich *C. hyperboreus* and *C. glacialis* to raise their chicks (Karnovsky et al., 2003; Steen et al., 2007). A significant shift in the copepod community to dominance by the more Atlantic species, *C. finmarchicus*, would, therefore, cause a decline in the Little Auk population and a probable replacement in the food chain by pelagic fish (Falk-Petersen et al., 2006).

Overall, predictions of how the marine food web will respond to climate change are challenging due to the non-linear nature of ecosystems. Changes in the phytoplankton influence the copepods that, in turn, influence the phytoplankton through changes in grazing pressure. The same is true for the relationship between the copepods and their predators, as well as for relationships further up the food chain. Thus, more research is needed before the essential details of the effects of climate change on marine organisms at the base of the food web can be understood and reliable predictions of ecosystem changes can be made.

In summary, climate-mediated changes of the Greenland Ice Sheet and subsequent changes in physical properties of the marine environment may have large consequences for the succession, composition and production at the base of Arctic marine food web. These changes will propagate up the food web and ultimately affect populations of fish, birds and marine mammals with major consequences for the structure, productivity and exploitation potential of the Greenland marine ecosystem.

4.2.2. Important economic species

Greenland waters are important fishing grounds, especially for the Northern shrimp (*Pandalus borealis*) and Greenland halibut (*Reinhardtius hippoglossoides*) fisheries that are essential to the economy of Greenland (Buch et al., 2004). Northern shrimp, currently the most important marine resource in Greenland, accounts for more than 70% of the total fisheries revenue. Catches have gradually increased to around 150 000 t/y (Kingsley, 2007), with 90-95% harvested off West Greenland. Northern shrimp in West Greenland and East Greenland waters are considered separate stocks.

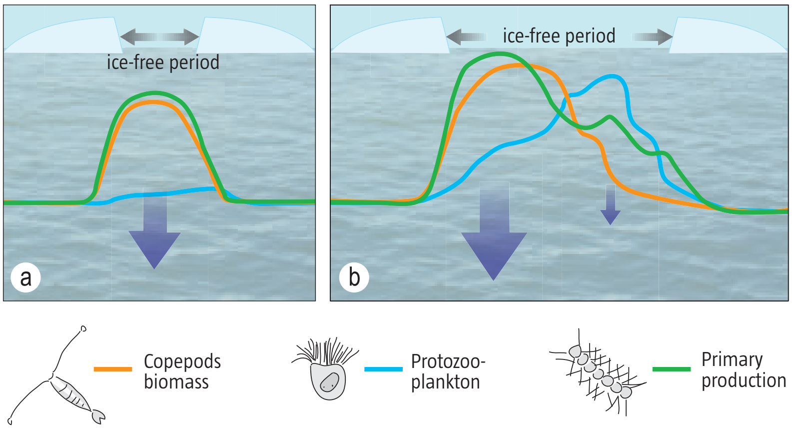


Figure 4.11 *Conceptual model of (a) present and (b) future conditions in sea ice cover, primary production, copepods biomass and protozoo-plankton, as well as sedimentation to the sea floor (arrows). The same figure also illustrates the change from north to south in a present-day situation. Source: Pers. comm. Torkel Gissel Nielsen, University of Aarhus.*

Northern shrimp is a cold-water species and stocks generally thrive when temperature declines. The relationship with temperature change is not well understood and may be complex; including improved survival of the larvae in cold periods but also ecosystem effects such as reduced predation by the concurrent reduction in boreal predators such as Atlantic cod (*Gadus morhua*) (Parson, 2005; Hvingel, 2006). Some of the important shrimp fishing grounds are periodically inaccessible due to ice coverage, notably the Disko Bay area (winter-spring), the central and northern areas of Davis Strait (*Vestis* during spring-summer) and East Greenland (*Storis* during spring-summer).

Commercially, Greenland halibut is currently the most important fish species. In the Ilulissat area, an inshore halibut fishery dates back to the late 19th century and the artisan fishery was further developed in the early 20th century (Smidt, 1969). An offshore fishery developed from around 1970 off both East and West Greenland. The present (2007) catch level (c. 66 000 t/y) is about equally divided between offshore West Greenland (Baffin Bay and Davis Strait), inshore West Greenland (Ilulissat, Uummannaq and Upernavik) and offshore East Greenland (Jørgensen, 2008; Lyberth and Boje 2008; ICES, 2008). Greenland halibut in West and East Greenland are considered separate stocks (Smidt, 1969; Boje, 2002).

The offshore fishery is constrained by ice coverage in both East and West Greenland. In inshore areas, a traditional winter fishery, based on dog sleds for transport, is still of local importance in the artisan fisheries (Lyberth and Boje, 2008). During mild winters, insufficient sea-ice extent and thickness jeopardize the safety of fishermen and dogs on the ice, and this type of artisan fishery becomes very limited. Nevertheless, the sea ice often remains too thick for small boats to break; the local fishery may, therefore, be closed for weeks.

Greenland fish stock composition and stock size is believed to change as an effect of climate change but there are, however, pending research projects to further study and quantify such changes. The effect of increased melting of the Greenland Ice Sheet, however, is believed to be of minor importance compared to the effects of increasing temperature especially on the offshore fishing grounds.

4.2.3. Marine apex predators in Greenland

Currently, there are only a few predictions of the direct effects of the melting of the Greenland Ice Sheet on marine apex predators. However, this large-scale event will have cascading effects on the primary productivity and associated species in the marine ecosystem; thus, many indirect effects can be expected. The two most important impacts will be changes in the annual sea-ice conditions and ecosystem changes associated with shifts in sea temperatures, currents and lower trophic level species due to warming and the influx of Atlantic water in southwest Greenland. These two indirect effects will, inevitably, have an impact on marine apex predators.

A general reduction in sea-ice coverage in Baffin Bay and the Greenland Sea has been predicted from models of the impacts of global warming. Changes in sea ice will directly impact ice-associated seals, walrus (*Odobenus rosmarus*) and polar bears (*Ursus maritimus*) as these species are periodically dependent on sea ice for molting, breeding or feeding (Laidre et al., 2008b).

Ice-breeding phocid seal species are critically dependent on stable pack ice, at least until pups have weaned and completed their post-weaning fast and transition to pelagic feeding (Stirling, 2005). Reduced coverage and stability of sea ice in whelping areas may cause neonatal mortality, changes in food availability for pups, and increased risk of epizootic diseases due to crowding (Lavigne and Schmitz, 1990; Heide-Jørgensen et al., 1992; Johnston et al., 2005). Large numbers of seals use sea ice for molting in both East and West Greenland and it is likely that, during a warm period, these molting areas would retract to more northern latitudes with more stable pack ice.

Polar bears are mainly dependent on sea ice as a platform for hunting. Sea ice also facilitates seasonal movements, mating and, in some areas, maternal denning even though these activities can all take place on land. Seasonally ice dependent polar bears using the southern extent of the ice edge are likely to disappear from the southern parts of their present-day range and to retreat to the northern parts of Greenland and the polar basin.

Walrus feed on shallow banks in west and northeast Greenland during winter and spring, and haul out onto sea ice between feeding activities. In summer, walrus haul-out both on land (in East Greenland) and on ice floes (in West Greenland). Ice provides several advantages including free transportation with the current, a platform for pupping and nursing, and the ability to remain and rest over offshore feeding areas. The proximity of the haul-out to food resources at shallow depths is important for walrus, and loss of sea ice may reduce their feeding options. In addition, the productivity of walrus prey populations is likely to be directly affected by warming of their environment (Grebmeier et al., 2006).

For marine apex predators, such as Brünnich's guillemot (*Uria lomvia*) or bowhead whale (*Balaena mysticetus*), which are not directly dependent on the sea ice, the timing of sea-ice recession is important for the onset and development of the trophic cascade that starts with primary production and ends in apex predator species. West Greenland constitutes a large latitudinal gradient in production, ranging from sub-Arctic areas influenced by Atlantic waters to high Arctic areas influenced by seasonal sea-ice cover and the presence of sunlight.

The largest colonies of Brünnich's guillemot are found in northwestern Greenland and these concentrations are likely to be sustained by the extreme blooms of primary production

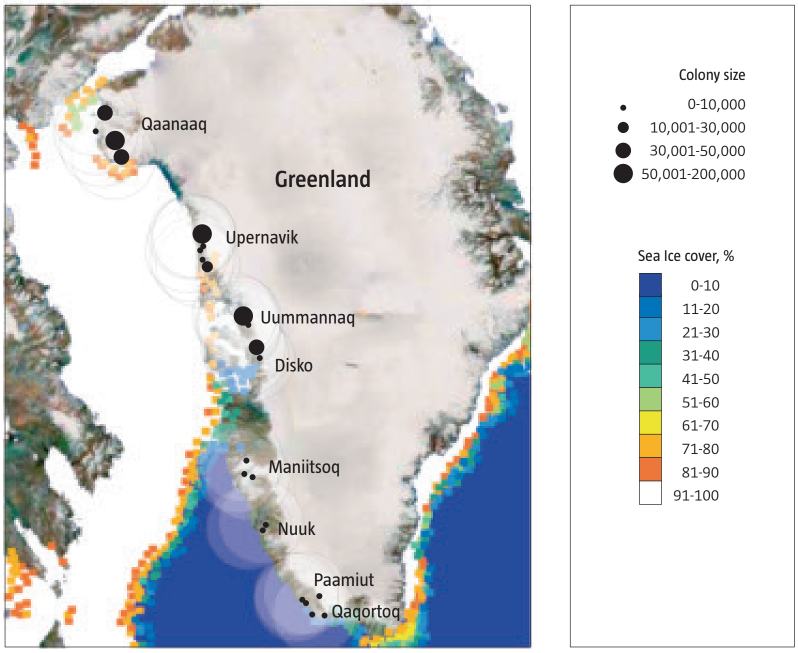


Figure 4.12 *Extant and extinct breeding colonies of Brünnich's Guillemot in West Greenland (maximum historical population size) with sea ice concentration shown for the last week of May (SMMR/SSMI annual mean 1979-2003). A potential maximum foraging range of 150 km around each colony was assumed during the breeding season. Source: Laidre et al. (2008b).*

and the resulting pelagic structuring of prey concentrations during spring when the birds arrive at the breeding grounds (Figure 4.12). This structuring is governed by the timing of sea-ice recession and the associated primary production bloom in leads and cracks where ice-associated crustaceans and fish are accessible to the shallow-diving guillemots (Laidre et al., 2008b). Other seabirds, such a common eider (*Somateria mollissima*) and king eider (*S. spectabilis*), will probably benefit from reduced sea-ice coverage on their winter-feeding grounds in West Greenland.

Cetacean species such as bowhead whale and beluga (*Delphinapterus leucas*) may benefit from reduced sea-ice coverage since it acts as a barrier for access to new feeding areas. For example, bowhead whales in Disko Bay are now feeding during winter in coastal areas that were not accessible during periods with more extensive ice coverage (Heide-Jørgensen et al., 2007; Laidre et al., 2007). Belugas wintering in West Greenland are forced to winter closer to the coast because of the Baffin Bay pack ice formation. In years with less sea ice they are found further west on the productive western edge of the Store Hellefiske Bank.

The narwhal (*Monodon monoceros*) is a specialized ice-associated cetacean that spends six to eight months of the year in very dense pack ice over deep water. They undertake strict annual migrations from summering grounds in coastal areas of northwest Greenland, East Greenland and Canada to wintering grounds in offshore pack-ice habitats. Narwhal feed at great depth on Greenland halibut and on mid-water squid during winter. They have a very specialized prey preference and are sensitive to unpredictable sea-ice changes that could cause ice entrapments (Laidre and Heide-Jørgensen, 2005). Their preferred prey, especially Greenland halibut, is likely to be affected both by changes in oceanographic conditions and in competitive interactions with new invading species or new predators.

The coastal banks of West Greenland are strongly affected by the strength of the warm current that travels northward from South Greenland. The effects of this warm and saline current of Atlantic origin on the ecosystem and the productivity along West Greenland are pronounced. Several apex predators are impacted by fluctuations in production in this area, including Arctic species and sub-Arctic species that use the area to feed, such as humpback whale (*Megaptera novaeangliae*), fin whale (*Balaenoptera physalus*), and minke whale (*B. acutorostrata*). Krill (*Euphausia superba* and *Meganyctiphanes norvegica*) are important prey species for these sub-Arctic whales, which feed in large aggregations throughout the summer and autumn.

Another key species for apex predators in West Greenland is capelin (*Mallotus mallotus*), a smelt fish species that is highly sensitive to sea temperatures and will move north to colder areas if temperatures increase by about 2 °C (Rose, 2005). A possible future increase in Atlantic cod along southwest Greenland would positively impact on several of the fish-eating marine mammals. Currently, several sub-Arctic species move into the Arctic in summer to feed on krill, capelin and Atlantic cod, and some of these may start arriving further north at progressively earlier dates, and may perhaps start competing with species living year-round in the Arctic (Laidre et al., 2008a). For example, humpback whale, fin whale, minke whale, blue whale (*Balaenoptera musculus*), pilot whale (*Globicephala melas*), killer whale (*Orcinus orca*), and harbor porpoise (*Phocoena phocoena*).

4.2.4. Summary

Generally, climate change has immediate effects on the physical environment of the ocean and in Greenland waters these effects are increased by a strong inflow of freshwater from the melting Greenland Ice Sheet. These changes may have large consequences for succession, composition and production at the base of the Arctic marine food web. The changes will propagate to levels higher up the food web and will ultimately affect populations of fish, birds and marine mammals with major consequences for the structure, productivity and exploitation potential of the Greenland marine ecosystem.

To date, there has been almost no research on the effect of the melting Greenland Ice Sheet on marine ecology; therefore much of the present knowledge on the effects of climate change is related to the melting and reduction of sea ice. Research is therefore needed to quantify how the higher trophic levels of the food chain will be affected by changes in the physical marine environment around Greenland as a result of the increased melting of the Greenland Ice Sheet. To be able to make proper and scientifically based plans and decisions for the future, such research is extremely important to the Greenland society since its economy is highly dependent on the surrounding ocean.

5. Socio-economic and cultural aspects of changes in the Greenland Ice Sheet

This chapter evaluates the possibility for projecting socio-economic and cultural impacts on Greenland's society caused directly or indirectly by changes in the Greenland Ice Sheet. There are, as yet, no well-documented direct causative links between the conditions for a society dictated by nature, and the way a given society develops. This chapter describes the development of the modern Greenland society from a historical perspective and introduces a number of specific cases that illustrate the propensity for change in a society that is derived from the Inuit culture. The Inuit culture has survived at the margin of human existence over a millennium of climate change and external cultural influences.

Some of the observations presented on climate change and societal adaptation interrelationships are, however, contentions rather than solid scientific facts. The historical record shows temporal coincidence between observed changes in climatic parameters and changes in society, but these coincidences may not have causal links, and even if causal links did exist under the circumstances prevailing during the historic record, these may have little or no predictive power for future changes in Greenland's society.

Despite the lack of predictive power, societal dynamics from inside Greenland as well as cultural influences from outside Greenland may have played the determinant role for the direction of change in Greenland over the past 300 years. Given the rate of change in response to globalization experienced by most societies, sociologic dynamics are likely to remain the strongest force for change in Greenland's society in the near future. It is not possible to estimate whether climate change may take the leading role for changing living conditions in Greenland on a longer time scale.

5.1. History of change

Human populations in Greenland have a long history of adaptation to changes in environmental and social conditions, whether imposed by changes in climate, socio-economic shifts and advances in technical resource exploitation, or by interactions between them. During the 19th and 20th centuries, three major shifts were experienced.

The first major shift took place during the latter part of the 19th century and continued until around the Second World War. This was the period during which the economy changed from one based on marine mammals to one based on fisheries. This shift coincided with a marked increase in sea temperature and a decrease in marine mammal stocks, combined with a dwindling world market for blubber and sealskin. Atlantic cod became the main commercial species and a major consequence of the shift in people's livelihood towards fisheries was a more sedentary lifestyle. This marked the starting point of the modernization of Greenland.

The second transformation occurred during the 1970s and 1980s and marked the shift from an economy based on Atlantic cod fisheries to one based on Northern shrimp fisheries. The main reason for this shift was probably the observed drop in sea temperature that restricted spawning possibilities for the Atlantic cod stock. However, an expansion in offshore cod fisheries by domestic and international vessels, following the depletion of the inshore cod stocks during the 1960s, may also have contributed to the collapse of the Atlantic cod population in Greenland waters.

The third transformation was the – still ongoing – shift towards a more diversified focus on fisheries. This shift has definitely been influenced by the present changes in sea temperature and is based on shrimp fisheries as the backbone of the economy, together with harvesting of Greenland halibut, snow crab (*Chionoecetes opilio*) and other species providing substantial contributions to the economy. The fishing industry has also become more dependent on fisheries in distant waters, for example by means of cod quotas acquired from Norway and Russia in the Barents Sea. Another characteristic of the present transformation is a general shift in the economy towards major contributions from business activities unrelated to renewable resource extraction. Therefore, besides the importance of the local environment on the resource situation in general, the characteristics of the world market are becoming increasingly important.

A common characteristic of all three major shifts is that both the process and the outcome have been the result of interaction between changes in the physical environment and changes in the socio-economic structures. In none of these major shifts has the result has been determined by a simple one-way causality. The implications of climate change have been the strengthening of ongoing socio-economic changes, and have contributed to outlining possible development paths. On the other hand, socio-economic and cultural changes have been reflecting a more or less pragmatic approach to both the natural forcing and political reality.

The shift from an economy based on marine mammals to one based on fisheries has had profound consequences on society, first of all through the sedentarization process. The shift from a cod- to a shrimp-based economy consolidated an already ongoing process of concentration and centralization, but did not totally restructure the organization of the communities. The present-day diversification of both renewable resource exploitation and the economy in general may contribute to a partial reversal of the former concentration policy by making some of the hamlets located near new resources and new economic opportunities (e.g., tourism) a potentially more active part of the economy.

The adaptation to a changing climate has followed many paths. The case of Kangaatsiaq in West Greenland illustrates how a diversified use of available resources enabled a community to withstand the effects of a sudden change in one of these resources. In Kangaatsiaq, the societal response has shown the following array of changes in the main livelihood over the past 70 years: Sealing and caribou hunting – cod fishing – salmon fishing – seal harvesting with nets – catfish fishing – shrimp trawling – caribou hunting with former fishing vessels – snow crab harvesting – lumpfish catching – cod fishing.

In addition to the main resource, many other natural resources are used throughout the year, thereby contributing to the resilience of the community. In central West Greenland, around 70 different species were traditionally used during a year. Further north, in Uummannaq, around 50 species were harvested. If possible, this high diversity in species utilization was combined with wage jobs. A social tradition of sharing and cooperating, a strategy for survival characteristic of the old Inuit culture, is still part of community life in many areas, and this tradition has been important for maintaining societal coherence through the great transformations. The informal economy and subsistence hunting and fishing is still contributing to local economies in towns as well as in hamlets.

In some areas, the need for small-scale adaptation to changes in climate and resources has been met by different approaches. In Qaanaaq and Ittoqqortoormiit, for instance, changing sea-ice conditions render the traditional activities of hunting and fishing on and from the sea ice difficult, and alternative species that could be exploited with the existing local harvest technology have not been identified. In Uummannaq and Upernavik, similar challenges have been met by changing the focus of hunting and fishing to more open-sea activities by boat. In this way, these communities have managed to respond to a new situation by opening up for new activities, such as tourism, that have contributed a supplementary income.

It is essential to emphasize that impacts of climate change cannot be studied in isolation from other environmental and social and cultural changes. History shows that Arctic societies, like those of Greenland, have always changed and have always demonstrated a great ability to adapt to change.

(Hertz, 1977, 1995; Rasmussen and Hamilton, 2001; Guenette et al., 2001; Hamilton et al., 2003; Rasmussen, 2003a,b, 2005, 2007c; Poppel, 2006a, 2007; Poppel and Kruse, 2009).

5.2. Socio-economic and cultural trends

Notwithstanding attempts to maintain an image of Arctic peoples as traditional hunters and trappers, reality shows modernized circumpolar societies with new technologies, satellite dishes, snowmobiles and welfare societies where concerns about future socio-economic structures have become the focus of politics – and should also become the focus of research. In this respect, Greenland is no exception. Modernization and globalization processes have led to a situation whereby less than a fifth of the population (i.e., 10 000 people) are economically involved in renewable resource harvesting, while by far the largest proportion of the population is involved in or connected to social services, administration, education, and similar activities.

In the traditional hunter society, marine mammals, especially seals, were essential for survival. For many people in the smaller settlements, seal hunting is still important. Every year, around 150 000 seals are killed and the meat is used locally; the main commercial seal exploitation is currently through the sale of skins to the Great Greenland tannery in Qaqortoq. As the milder climate and more fertile soils render South Greenland suitable for sheep farming, this livelihood has been developed since the beginning of the 20th century and now provides around 20 000 lambs for slaughter each year.

However, marine fauna in the very productive waters around Greenland have been the basis of the production economy for the past 50 years. By far the most important of the commercially exploited species is the cold-water Northern shrimp, with an annual catch of around 150 000 tonnes. Several other species are commercially used, and in the future, Atlantic cod fisheries may recover due to increased sea temperature.

Although Greenland has considerable raw material deposits, commercial exploitation has been limited by adverse natural conditions and difficult logistics. Cryolite, coal, copper, marble, zinc, lead and silver have all been mined, and gold, olivine, and rubies are currently extracted. International interest in exploration for minerals and hydrocarbons is increasingly and some deposits may prove of economic interest in the future.

Hydroelectric power is a resource that is increasingly being used in Greenland, not for energy export but for domestic consumption and for energy intensive activities such as aluminum production. It is important to note that hydroelectric power is based entirely on water basins supplied by meltwater from the Greenland Ice Sheet.

5.2.1. Implications of changes

The new expectations for large-scale mineral and hydrocarbon production and energy intensive industries as well as the dependence on fisheries and hunting activities are influenced by changes in climate and associated changes in the Greenland Ice Sheet and sea-ice conditions. Renewable resources are affected through their interactions with the environment, while non-renewable resources are affected by changes in accessibility and transportation.

The smaller settlements will experience the immediate effects of these changes because renewable resources are still of significance to them and because access to resources and to settlements will be altered by changes in accessibility, abundance and geographic distribution of these resources, as well as by changes in their economic importance. Tourism has become an alternative source of income for many local communities, enabling a positive interaction between new economic opportunities and traditional activities that are less affected by change than exploitation.

At a longer perspective, Greenland's economy is shifting towards increased dependency on non-renewable resources and in so doing will be highly affected by the ongoing changes, including those affecting access to resources and transportation. The decisive force will be the population's ability to adapt to the changes and to continue to improve the level of education.

(Hamilton and Seyfrit, 1993; Paldam, 1994; Rasmussen, 1997, 1998, 2002; Winther, 1999, 2001; Zelarney and Ciarlo, 2000; Duhaime et al., 2001; Hansen, 2003; Nellemann and Vistnes, 2003; Poppel, 2006b; Poppel et al., 2007; Kruse et al., 2008).

5.3. Influences

5.3.1. Demographic challenges

When discussing changes in the Arctic, focus is often on the obvious and visible large-scale drivers such as changes in climate and the environment, while less visible, but nevertheless very effective, drivers may be neglected or overlooked. In relation to the inhabited Arctic, ongoing demographic changes may have similar or even more wide-ranging consequences than changes in climate. Two important factors are often neglected: (1) the gender perspectives of development and (2) the generation shift.

The traditional focus on middle-aged males (i.e., the hunters and fishermen, the 'providers', the administrators) as the characteristic elements in the economic and social aspirations of the community excludes more than 75% of the population, even though it is increasingly apparent that the latter may be crucial and decisive for the future development of communities in the Arctic. Demographic indicators comprise a fundamental dimension of human settlements, and socio-economic factors cannot be understood without taking into account population size, its changes, rates of births, deaths and migration, and the resulting age structures and gender ratios. Similarly, the demographic and health indicators are important in monitoring the sustainability of the socio-economic development, the wellbeing of populations and their adaptation to change.

Although resource exploitation is still perceived to be the main economic basis for communities in the north, reality is that the service sector, with wage jobs in administration, education and social services, has become the main source of income for most families. In Greenland, these incomes have, in turn, become necessary for maintaining many of the traditional renewable resource activities, as 24% of hunters and fishermen have supplementary wage jobs. In more than 70% of the households, wives contribute to the family income, typically through wage jobs in the service sector.

Both changes in climate and ongoing demographic and socio-economic changes are individually important in relation to the future of Greenland. Over the past century, Greenland experienced both climate warming and cooling, and the subsequent increase and decline or crash in stocks of commercially important species in coastal waters. Changes in population size have been mediated by altering job and income opportunities in local or regional labor markets.

Adjustment to change has been crucial, and climate change may even have facilitated the modernization process with its positive effects on health, longevity and population growth. But at the same time there are also abandoned and depopulated settlements serving as examples that not all individuals or communities have had the ability for *in situ* adjustments to change. At present, it appears that people in remote rural communities will be more vulnerable to climate change, changes in the environment and shifts in demographic and economic factors.

(Kuznets, 1975; Preston, 1975; Hansen, 1992; Hamilton and Seyfrit, 1994; Kirk, 1996; Bjerregaard and Young, 1998; Hamilton and Otterstad, 1998; Kirmayer et al., 1998; Cutler and McClellan, 2001; Leineweber et al., 2001; Bjerregaard and Curtis, 2002; Fogel, 2004; Bjerregaard et al., 2004; Rasmussen, 2005, 2007a, 2008; Cutler and Miller, 2005; Berner et al., 2005; Björkstén et al., 2005; Curtis et al., 2005; McCarthy et al., 2005; Deaton, 2006; Parkinson, 2007; Hicks and Bjerregaard, 2007; Niclasen and Bjerregaard, 2007; Andreeva et al., 2008; Rauhut et al., 2008; Poppel and Kruse, 2009).

5.3.2. Accessibility and impact on communities

Changing ice conditions, both on land and at sea, are important factors influencing everyday life in most communities in Greenland, mostly by determining the accessibility of the water bodies. Access to the sea, and sea transport to and within Greenland, are determinant factors for hunting, fishing, exchange of goods and tourism, which is becoming increasingly important for many communities.

Hunting terrestrial mammals is contributing to both subsistence and informal economic activities of many Greenlanders. Muskox (*Ovibos moschatus*) and caribou / wild reindeer (*Rangifer tarandus*) are hunted by licensed commercial hunters and by sports hunters in the autumn, when as much as a fifth of the population goes hunting. Changes in the position of the Greenland Ice Sheet margin may affect hunting activities by changing the routes of rivers draining meltwater from the ice, and in conjunction with rising temperatures may affect the migration patterns of the game. In cold weather, the animals move towards the coast, where they become more accessible to hunters, while in warmer weather, the animals move towards cooler conditions closer to the ice margin in order to reduce insect harassment. As a result, the walking distance of the hunters to the animals becomes increasingly difficult to cope with, first while searching for the animals, and especially when the meat is carried by hand back to boats on the coast.

Similarly, subsistence hunting for marine mammals as well as fisheries will be affected by changes in water temperature and ice conditions due to changes in accessibility and changes in species diversity. In many communities, especially in the smaller and more remote settlements where subsistence and informal economic activities may provide as much as 50% of the economic basis, changes in sea-ice conditions and sea temperature will affect the availability of traditional harvest species. Although the overall commercial value of seal hunting is limited, seals still play an important role in the diet of Greenlanders. In East and North Greenland, seal meat is important also as dog food. Dog teams are still necessary for pulling sledges used for hunting and fishing during the long winter season. Dogs and sledges are also key attractions for tourism, which is an increasingly important contributor to local economies.

The potential impacts of climate changes on communities in different regions in Greenland are diverse. While hunters in the northern parts of Greenland are experiencing problems with thin and less reliable sea ice during winter, a new generation with more focus on fisheries sees the changes as an opportunity to expand the fishing activities. In sheep farming on South Greenland, the temperature increases are considered a boon by extending the growing season, ensuring a more stable production of winter fodder for the animals, and rendering improved possibilities for expanding crops such as potatoes and beets. In the same region, a reduction in the density of spring-summer drift ice (*Storis*), which causes severe navigational problems, would be considered another positive effect of the warming process.

(Preston, 1975; Dahl, 2000; Curtis et al., 2005; McCarthy et al., 2005; Rasmussen, 2005, 2007b; Poppel, 2006b; Tommasini and Rasmussen, 2006; Rauhut et al., 2008; Poppel and Kruse, 2009).

5.3.3. Possible consequences for new activities

Reduction of glaciers, ice sheet and sea ice in and around Greenland will increase accessibility to new areas with the potential for commercial exploitation of minerals and energy resources. A significant side effect to the retreat of the ice sheet margin will be the increased volume of meltwater that can be harnessed for producing hydroelectricity. This potential is already included in the new economic strategies for Greenland. Over the past 50 years, the hydropower perspectives in Greenland have been mapped and evaluated as a potential contribution to the development of the national economy. The hydropower exploitation has already emerged as three hydropower plants: A 45 MW plant near Nuuk, a

1.2 MW plant in Tasiilaq, South Greenland, and a 7.2 MW plant near Qaqortoq, South Greenland. A 15 MW hydropower plant near Sisimiut, West Greenland, is under construction and will become operational in 2010.

Several other hydropower plants are in the planning process, now aiming at providing power to new economic activities connected to energy intensive large-scale raw material processing. The first project is the supply of energy to an aluminum smelter currently planned at Maniitsoq, West Greenland. Even though Greenland does not have the ore resources needed for aluminum production, import of ore (i.e., bauxite) from Australia or South America to Greenland could be a viable economic activity given the access to large and continuing amounts of meltwater from the Greenland Ice Sheet and the derived hydroelectricity.

A key question, however, is the stability of the water supply to the new power plants as the ice sheet margin retreats due to melting. Increased amounts of silt and sand in the meltwater may fill dammed lakes and affect the viability of the dams, eventually requiring additional reconstruction in order to maintain the water flow needed. Or worse, the bedrock topography exposed by the retreating ice may cause new meltwater drainage patterns, leading to a partial or complete drying out of the water supply to the hydropower plants.

(Anonymous, 1994; Udvalget om socio-økonomiske virkninger af olie- og gasudvinding samt mineralindustri, 1997; Rasmussen, 1997, 2003a,b, 2005; Keddeman, 1998; Storey and Hamilton, 2003).

5.4. The Ilulissat case

The Ilulissat glacier at the head of the Ilulissat Ice Fjord is one of the world's fastest moving glaciers producing more than 10% of the total discharge of icebergs from the Greenland Ice Sheet. During the culmination of the last Ice Age 21 000 years ago, an ice sheet that reached out onto the margin of the continental shelf covered the present-day ice-free land bordering the Greenland Ice Sheet.

After the end of the Ice Age, the melt-back of the ice started about 11 700 years ago and, during a warm climatic period 8000 years ago, the ice rapidly regressed causing the glacier front to retreat inland to a position more than 50 km east of the present grounding line. With the onset of the Little Ice Age, the glacier front again advanced, and in 1851, a maximum extension was recorded in the middle of the Ilulissat Ice Fjord close to the former Inuit settlement Qajaa. Following this advance, the ice front stabilized and a slow regression prevailed until the front suddenly collapsed in 2002-2003 causing the floating ice tongue to withdraw more than 15 km eastward in a few years. This rapid withdrawal ceased in 2007, when the ice front reached the grounding line.

5.4.1. Inuit settlements, cultural changes and role of the Ice Fjord

The Ilulissat glacier and the Ice Fjord have had an immense impact on the population of the region during the history of human occupation. The glacier produces a large amount of nutrient-rich meltwater that facilitates the primary production in the fjord. The constant input of nutrients thus forms the basis for a rich fauna in the fjord, which has provided excellent hunting and fishing conditions for people inhabiting Disko Bay through the ages. Prehistoric hunting areas and settlements were established and abandoned as the location of the glacier front changed. In this way, the Ilulissat area illustrates the interplay between an ever-changing glacier environment and the development of human occupation.

Many archaeological findings testify to the long history of settlement in the Ilulissat Ice Fjord region, where hunting and fishing have always been the basis for life. The Saqqaq people reached West Greenland from Alaska 4500 years ago and settled in the Ilulissat area where they lived for 1500 years before they disappeared. West Greenland was then uninhabited until a new wave of paleo-eskimo settlers, the Dorset people, arrived 2800 years ago. They occupied the region for 800 years before they disappeared leaving the Ilulissat region and Greenland uninhabited for the next 1000 years. By 1200 AD, people of the Thule culture settled in the region and established two flourishing communities, Sermermiut and Qajaa, along the ice fjord. These settlements are among the largest prehistoric Inuit settlements discovered in Greenland and, together with the dense pattern of smaller settlements in the area, indicate that the living resources around the fjord were rich, stable and plentiful over long periods of time (Grønnow and Meldgaard, 1988).

5.4.2. Ilulissat Ice Fjord and present-day socio-economic effects

In 1741, the Danish colony of Jakobshavn (today Ilulissat) was established and this led to the depopulation of the rich Sermermiut Inuit settlement around 1850. Today, Ilulissat is the third-largest town in Greenland with a population of more than 4000. Half of Greenland's coastal fisheries are now centered here and the town is also home to the country's largest fish-processing factory. Greenland halibut are plentiful throughout the fjord system and are subject to one of the most intense forms of fishing in the world. Long-line fishing for Greenland halibut from the sea ice in the Ice Fjord has been an important part of the fishing tradition through centuries and it is still taking place.

The abundance of resources and the adaptability of the traditional culture to the changing conditions of the glacier and winter sea ice have been crucial to the prosperity of the social and economic life of the region. Although the remarkable present changes of the ice will constitute new challenges to economic and social activities in the region, the conditions may now be different, less predictable and more challenging than before: Icebergs are smaller and the immense discharge by the glacier in combination with a rise in sea temperature, have caused a considerable reduction in sea-ice cover during winter.

This may imply profound changes for the production of nutrients – the basis for biological life in the region – and may lead to smaller stocks of Greenland halibut and Northern shrimp. These species are fundamental, not only to human harvesting, but also to the seals and whales that form a significant part of the daily food supply for humans and dogs. Also, the distribution of seabird species may be affected. Other species may enter the region offering new and potentially prosperous options for economic activities. Such challenges will be met as before.

The adaptations of today are, however, far more complicated. Modern and economically viable harvest of renewable resources is highly industrialized, and to establish new production will be costly, requiring national initiatives as well as international involvement and regulations applied to managing, accessibility and rights. The local capacity for adaptation will be met by implications of today's global interaction.

Changes in sea-ice cover – especially a reduction in winter sea-ice extent and thickness – will inevitably alter marine transport conditions. Reduced sea ice may offer new transportation routes and prolonged navigation seasons, but will also increase marine access to natural resources and, accordingly, the potential for disturbance and for more efficient fisheries. Although the changes in sea ice may lead to significant changes in the socio-economic life of the region, it is still premature to state the nature of these socio-economic changes. The unpredictability of climate change and its local implications make planning, decision-making and management even more challenging. In spite of its richness, Arctic nature is vulnerable and poor management may lead to serious and irretrievable damage.

5.4.3. The Ilulissat Ice Fjord and its impact on science, politics and tourism

Ilulissat Ice Fjord and its glacier is one of the world's best-studied glaciers with the longest observational record. The ice fjord has attracted many scientists through centuries, the first being H.J. Rink who studied and described the glacier and the Greenland Ice Sheet in the mid-19th century. The studies were instrumental in formulating the important theory of the Quaternary glaciations. Today, the Ilulissat Ice Fjord is still the subject of detailed international research because of its exceptional manifestation of changes.

The Ilulissat Ice Fjord was nominated a UNESCO World Heritage site in 2004. The World Heritage List uses the name Ilulissat Ice Fjord for the entire world heritage site, which includes land areas, the glacier front and parts

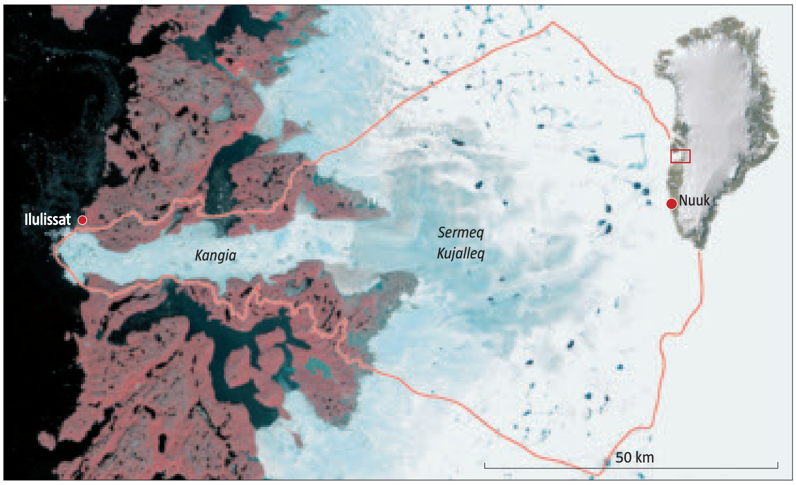


Figure 5.1 *Location of Jakobshavn Isbræ and adjacent region. Place names are: Ilulissat = town of Ilulissat (formerly named Jakobshavn); Kangia = the ice fjord proper (white because of tightly packed floating glacial ice and stranded icebergs); Sermeq Kujalleq = Jakobshavn Isbræ (also known as Ilulissat Glacier). Copyright by Geological Survey of Denmark and Greenland (GEUS).*

of the Greenland Ice Sheet adjacent to the ice fjord, recognizing the uniqueness of the area and the extensive interplay between nature and culture.

By entering this prestigious list, tourism has virtually exploded in the region. This has made Ilulissat Greenland's leading tourist town and the area now hosts more than 50% of the total number of tourists visiting Greenland. Thus, the Ilulissat Ice Fjord is by far the main attraction and cruise ships arrive in increasing numbers during the summer to let their passengers visit the fjord.

It is not just tourists that are congregating in the area. The Ilulissat Ice Fjord has also attracted a large number of politicians from around the world who wish to experience firsthand the global climate change as witnessed here. The Ilulissat Ice Fjord has now entered political discussions in a way that has never been seen for any part of Greenland before (Mikkelsen and Ingerslev, 2002).

5.5. Conclusions

Both local and global effects of the ongoing changes in Earth's climate will inevitably affect Greenland and its population. This chapter has attempted to assess the extent to which changes in the Greenland Ice Sheet will be major factors in shaping the future society of Greenland. As it is impossible to establish direct causal links between natural boundary conditions and the socio-economic development of any society, this chapter has reported the response of Greenland's society to past changes in climate and has compared these to the effects of societal dynamics. The history of the Thule culture, which gradually transformed into the present-day society in Greenland, has been one of constant adaptation to changes in climate and increased interaction with European culture.

At present, global climate changes are being expressed more strongly in the Arctic region than at lower latitudes. It is also clear that Greenland has experienced large fluctuations in climate during pre-historic and historic Inuit habitation. Hence, the Greenland culture is rooted in a tradition of constant adaptation to new natural and cultural conditions. In the globalized modern society there are several important demographic reasons why the Greenland society is undergoing radical changes irrespective of changes in climate.

Until now, the observed effects of changes in the Greenland Ice Sheet due to global warming have caused only minor impacts on society. The most marked change observed by the Greenland population has been the retreat of Jakobshavn Isbræ. This has caused an increase in tourism and has increased international attention on Greenland as an icon of global climate change.

In the coming years, changes in drainage patterns around the Greenland Ice Sheet may affect coastal waters and thereby cause changes in the distribution of marine fish stocks and game species. Future changes in drainage patterns may also be a major concern for large infrastructure investments with decadal to centennial life spans, such as hydroelectric power plants and transmission lines, roads, airports and seaports. This chapter has not considered effects from changes in permafrost, although these may have considerable economic implications; the issue of permafrost is considered as separate from that of the ice sheet.

Based on the historic observations and the strong drive for change in Greenland's society today, it is likely that the consequences of changes in the Greenland Ice Sheet will be minor compared to socio-economic, cultural and demographic influences generated within the Greenland and international societies.

6. Major knowledge gaps and recommendations

6.1. Introduction

New research results from the Greenland Ice Sheet often lead to the identification of new gaps in knowledge, although they may become more specific over time (see Table 6.1). For example, research projects initiated under the International Polar Year 2007-2008 should eventually lead to the identification of new gaps in knowledge or to changes in the existing gaps in knowledge.

Table 6.1 *Major gaps and recommendations.*

Major gaps	Recommendations
Large variability between presently used SMB models. Incomplete understanding of SMB processes. Large spatial variability in SMB at the GRIS margin. Large error in determination of surface annual accumulation.	Targeted *in situ* SMB data with which to calibrate/validate models. Systematic studies focused on improvement in process understanding. Better sub-grid scale quantification of the key processes; refreezing, albedo and blowing snow. Improved determination of snow densification.
Discrepancy in estimating the mass change from satellite gravity measurements. Incomplete estimate of the mass loss from marine terminating outlets. Incomplete determination of the annual cycle in SMB from satellites. Lack of ice discharge determination from ice streams and narrow outlets.	Establish a community wide recognition on which processing method provides the most reliable method for tracing mass changes. More systematic mapping of ice thicknesses from outlet glaciers. Sampling sufficiently frequent to capture the annual cycle in accumulation, melt and iceflow. Development of high resolution sampling methods from satellites to capture basin scale changes.
Large seasonal as well as interannual variability in marginal ice fluxes. Basic understanding of coupling between ocean and marine based fronts. Lack of thickness determination around highly dynamic outlets. Incomplete understanding of flow properties deepest in the ice sheet.	Improved theoretical understanding of the basal hydrology and its impact on ice flow. Theoretical understanding of the calving and basal melting at marine margins where changes occur. Develop methods to retrieve ice thicknesses at heavily crevassed margins. Determination and validation of the thickness of the basal temperate layer as well as deep ice temperatures.
Incomplete cryospheric representation and determination in climate models. Lack of full coupling between surface dynamics and the atmosphere. Major dynamical features like ice streams lack representation in ice sheet scale models. Several-fold changes in velocity variations at marine based outlet glaciers.	Validate climate models more systematic with *in situ* observations of surface climate and mass balance of the GRIS. More accurate incorporation of surface albedo and snow microphysics in climate models. Better model resolution as well as process understanding leading to well parameterized models. More complete determination of process leading to improved parameterizations of basal and marginal (ocean) linked process.

6.1.1. Arctic Climate Impact Assessment

Earlier assessments identified some of the major gaps that are currently being addressed. The Arctic Climate Impact Assessment (ACIA, 2005) focused on improving projections of future changes and identified several major gaps in knowledge that are also valid for the Greenland Ice Sheet: (1) A need to improve understanding of albedo changes and feedback mechanisms; (2) the need for studies of outlet glacier dynamics with an emphasis on their potential for triggering persistent and rapid thinning; (3) a need for improving ice dynamic models for determining the long-term response of the Greenland Ice Sheet to past climate change; (4) a need for improving parameterization and verification of internal accumulation models; and (5) a need to improve understanding of the relationships between climate change, meltwater penetration to the bed, and changes in iceberg production.

Uncertainty in scenarios concerning how glaciers, ice caps and ice sheets contribute to sea level rise in decadal to centennial time scales needs to be reduced. This requires an assessment of the Greenland Ice Sheet and its stability and vulnerability to climate change, including sudden and potentially irreversible climate change. The Climate and Cryosphere Project (CliC), established in March 2000 to stimulate, support, and coordinate research into the processes by which the cryosphere interacts with the rest of the climate system, recommends that an insight is needed into the highly non-linear links between the various components of the cryosphere and their likelihood for producing significant global changes. CliC recommends that the long-term ice sheet monitoring system is improved in order to make realistic representations (modeled as well as observational) of spatial and temporal variability in the surface mass budget in areas sensitive to sea level change (Climate and Cryosphere Project, 2007).

The IPCC's Fourth Assessment Report (IPCC, 2007) stated that realistic scenarios of the Greenland Ice Sheet surface mass balance require a resolution exceeding that of the current AOGCMs used for long-term climate experiments. Present climate models do not generally include a representation of the refreezing of surface meltwater. All models predict an increase in snow accumulation, but there is much uncertainty in its size. In projections for Greenland, an increase in ablation is important but uncertain, being particularly sensitive to temperature change around the margins. The main uncertainty is the degree to which outlet glaciers having a marine terminus respond to climate change. Further accelerations in ice flow, of the kind recently observed in some Greenland outlet glaciers (and in West Antarctic ice streams), could increase the ice sheet loss of mass substantially, but quantitative projections cannot be made with confidence.

6.1.2. Summary of earlier assessments

The research needs presented in the ACIA, CliC and IPCC assessments of the Greenland Ice Sheet vary in their degree of detail, but all focus on the importance of making more precise scenarios and filling gaps in knowledge. There is a common agreement, that improved understanding of processes is needed in order to determine meltwater run-off more precisely. Also, the rapid increase in ice discharge from the outlet glaciers observed over the last decade calls for a better understanding of processes as well as a better formulation of theory before models quantifying the observed increase in ice flux can be developed. The three assessments all call for better monitoring capabilities and a more precise determination of present state of balance in the Greenland Ice Sheet before more reliable scenarios can be made.

6.2. Gaps in knowledge of surface mass balance

Significant differences exist between different estimates of the components of the surface mass balance derived from numerical modeling and downscaling of climate re-analysis data. For example, there are important differences in distribution and total accumulation in the 'wet' southeast Greenland. The differences between model estimates can be as large as the net surface mass balance for a given year and the standard deviation of the differences is of the order of 100 Gt. Although values for total run-off may agree between models, the individual terms included in the calculations of total run-off vary much more widely, particularly when comparing melt and refreezing. Total differences are of the same order of magnitude as the inferred increase in mass loss due to changes in ice dynamics over the last decade (Rignot and Kanagaratnam, 2006). The differences suggest that mass budget calculations are seriously hindered by uncertainties in surface mass balance.

Uncertainties in surface mass balance originate from a range of sources related to both lack of *in situ* data and incomplete process understanding. To reduce uncertainties to a level useful for mass budget calculations and reliable predictions will require concerted effort. Especially, (1) to collect more targeted *in situ* data to calibrate and validate models, (2) to improve process understanding, and (3) to model key processes such as refreezing, surface albedo, blowing snow and sub-grid scale effects. Sub-grid scale effects are particularly significant in coastal regions due to higher surface relief, which makes ablation observation uncertainties difficult to incorporate into surface mass balance model validation. The most important parameters for controlling run-off in surface mass balance modeling, however, are surface albedo and meltwater refreezing.

6.3. Gaps in knowledge of net balance or components thereof

At present, the best estimate of the mass of the Greenland Ice Sheet is probably provided by GRACE (see section 2.4.2.3). However, different processing methods result in different results. One method of averaging the total mass of the Greenland Ice Sheet over 30-day periods and calculating seasonal loss (Velicogna and Wahr, 2006) provides a mass loss of 211 Gt/y, while another uses different processing algorithms for analyzing the distance between the two satellites (Luthcke et al., 2006); this second approach provides a divergent mass loss of 154 Gt for the year 2007 (Witze, 2008). Such discrepancies reflect ongoing discussion on which method is most valid for detecting mass changes in the Greenland Ice Sheet using GRACE.

Measuring volume change by altimetric methods such as laser (IceSat) or by radar (CryoSat) provides another tool for detecting changes in mass balance. However, connecting volume change to mass changes requires insight into densification processes of the surface snow and firn, which also has a strong seasonal dependence when the surface temperature reaches the melting point. Moreover, it remains difficult to produce a measuring accuracy any better than a few tens of centimeters, and this also renders it impossible to capture the seasonality of the parameters.

Satellite radar techniques measuring ice displacements and ice fluxes near calving glacier fronts are unique in providing a means for detecting iceberg production when the ice thickness is known. However, the lack of measured ice thicknesses from all outlet glaciers and the difficulty of measuring ice movement in the interior of the Greenland Ice Sheet constrain this technique.

The various techniques used to estimate the state of balance of the Greenland Ice Sheet differ in their methodology for detecting changes in specific components of the system; which means that they do not return results that are mutually compatible. Closing the gaps between the results for different techniques will make an important contribution towards a more consistent understanding of change in the Greenland Ice Sheet. Filling these gaps in knowledge will require common or compatible temporal and spatial sampling as well as an understanding of the annual cycles in mass balance, including its most important parameters: Accumulation, melt, run-off, and ice flow. Temporal sampling will need to be frequent enough to capture the annual cycle, or designed to mitigate the effect of interannual variations in this cycle on annual estimates.

Spatial sampling capturing variation at a drainage-basin scale would allow for a more direct comparison of changes in discharge and measures of ice sheet mass or volume. Improved sampling of surface height with new satellite missions may provide altimetric determinations that more faithfully reflect the impact of annual cycles on surface height. However, in order to secure significant progress in understanding densification processes, observation and modeling of the surface snow are also required.

6.4. Gaps in knowledge of ice dynamics

A better understanding of basal boundary conditions (i.e., sliding laws) is needed. This involves a more precise understanding of the mechanisms of basal motion in order to derive good parameterization, which accounts for seasonal and longer-term variability. An improved understanding of basal boundary conditions is inherently linked to an understanding of subglacial hydrology, which is assumed to drive the present speed-up of calving outlets. For instance, subglacial hydrology and the complex connection between surface hydrology (supraglacial lake drainage) and the bed are integral to basal flow, which can dominate ice flux near the margin. Modeling the ice sheet hydrology, however, is still in an early stage of development and has not yet been coupled with ice dynamics.

Concerning the ice-ocean interface, gaps in knowledge are mainly related to the assessment of melting rates under floating glacier tongues, where the ocean and heat circulation in combination with freshwater from melting ice drive the circulation and impact on the calving front (Motyka et al., 2003). Generally, iceberg calving and basal melting at marine margins are difficult to measure and to simulate, and these processes are absent or overly simplified in current ice sheet models.

Observational gaps mainly relate to determining ice thickness at heavily crevassed outlet glaciers, due to difficulties retrieving radar signals in these regions. Also, ice temperatures and, in particular, thickness of the basal temperate layer are not well known, which in combination with ice fabric (Lüthi et al., 2002) show a strong dependence on ice deformation rates and hence dynamics of the outlets.

6.5. Gaps in knowledge of ice sheet and climate modeling

Models of meteorological processes, ice sheet mass balance, and ice sheet dynamics are constantly improving, but there are still many uncertainties associated with simulating the Greenland Ice Sheet and its likely response to climate change.

6.5.1. Climate models

Many climate models do not resolve the steep topography of the Greenland Ice Sheet margins very well, limiting model skill in simulating orographic precipitation and ice sheet ablation in this zone. High-resolution meteorological models such as PMM5 (e.g., Box et al., 2006) overcome this problem, but there are still outstanding issues with regard to temperature biases, simulation of mass balance processes such as sublimation and boundary layer winds, and the large-scale boundary conditions for regional climate modeling in future climate-change scenarios.

The dominant precipitation and ablation processes can also be event-driven in Greenland (i.e., resulting from a small number of extreme synoptic weather events), and these processes are well captured in simulations that are forced by historical (i.e., re-analyzed) climatology but do not necessarily arise in climate model simulations strong in predictions of the mean state; and they do not always represent the full range of weather variability.

Probably the most important boundary conditions for climate model simulations of the Greenland Ice Sheet are the surface properties (i.e., snow's physical representation i n the models). The inherent instability of snow crystals on the ground, the ever changing meteorological conditions, and the feedback between snow and atmosphere via snow albedo makes the snow-atmosphere coupling an important and highly dynamic one. New progress in snow physics theory and parameterizations needs to be incorporated into climate models.

6.5.2. Glaciological models

Ice sheet models now operate at resolutions of 5 to 20 km for the Greenland Ice Sheet, representing the overall ice sheet geometry quite well but still not capturing details of ice margin positions or some of the major fjords and outlet glaciers. This means that it is still difficult to compare model simulations with geological reconstructions of recent (i.e., 20th century) ice sheet changes. In addition, it is difficult to compare modeled ice discharge with observations, with the more distributed (i.e., less channelized) ice flux in the model. Major ice dynamical features such as the glaciers at Ilulissat and Storstrømmen (also known as the Northeast Greenland Ice Stream) are known to account for a large percentage of ice discharge from the Greenland Ice Sheet, but these features are generally absent in ice sheet models. This is partly an issue of model resolution and partly an issue of complex dynamical processes that are presently too poorly understood to be well parameterized in models.

Ice sheet models are presently insensitive to ocean forcing, in contrast with recent observations of significant (several-fold) velocity variations in marine-based outlet glaciers, most likely to have been triggered by ice-ocean interactions and thinning at the grounding line. Models still cannot simulate this interannual flow variability, so it is unclear whether they are able to provide a credible estimate of the response time of ice sheet change to atmospheric or oceanic forcing.

6.6. Perspectives

During the 20th century, the mass discharge from the Greenland Ice Sheet was almost equally divided between meltwater run-off and iceberg discharge (Reeh, 1994) – making the two terms equally important. The magnitude of the present-day balance is uncertain due to lack of coherence between the different estimates relying on different methods. Closing the gaps between the results is a prerequisite for reliable estimates of the present balance. Furthermore, a firm knowledge about the present balance is needed in order to generate reliable future scenarios.

Making specific recommendations on research needs to close the gaps in knowledge requires detailed insight into the problems. However, detailed insight is still lacking for many of the major knowledge gaps, such as the processes underlying increased ice discharge and meltwater refreezing. Instead of reducing uncertainty about changes in the mass balance of the Greenland Ice Sheet, the new observational methods from space have revealed to the research community the incomplete understanding of the fundamental dynamics of calving glaciers. This lack of understanding may explain why gaps in earlier assessments are still unresolved in the present-day assessment of the Greenland Ice Sheet. Closing such gaps in knowledge, where there is a lack of fundamental understanding of the processes, demands the development of theory and methodology to validate the theory against field measurements.

Most changes appear in the marginal regions of the Greenland Ice Sheet where data coverage is most scarce due to logistical access constraints. Moreover, thin ice in these regions, combined with a complex terrain, results in highly variable ice dynamics, melt rates, and accumulation patterns. This calls for high spatial resolution modeling in combination with enough *in situ* measurements to capture adequate detail and to gain sufficient understanding of the present thinning process.

New satellites have triggered a quantum leap in surface observations of the Greenland Ice Sheet. However, *in situ* measurements are still progressing at a slow pace due to resource-demanding fieldwork. The development of new, and less labor-intensive, measuring techniques could potentially advance the process understanding over larger areas. A combined effort in advancing process theory, field methods and coverage as well as modeling efforts could facilitate major advances in understanding the present balance of the Greenland Ice Sheet and, hence, help create less uncertain future scenarios.

7. Extended summary

The Greenland Ice Sheet is the largest body of freshwater ice in the Northern Hemisphere. If this ice sheet were to melt, the global eustatic sea level rise would be almost 7 meters. In the coming decades and centuries, the Greenland Ice Sheet will be highly susceptible to the predicted strong warming of this part of the Arctic.

The reaction to climate change by the Greenland Ice Sheet must be seen and understood within the context of climate changes in, around and above Greenland. Long-term temperature measurements from the east and west coasts of Greenland display uncorrelated trends. In general terms, the west coast experienced climate warming from 1873 to1930 followed by cooling until 1985. Since then, West Greenland has experienced a warming of 2 to 4 °C, primarily driven by winter temperature anomalies. The present warm period is not unprecedented in the temperature record and is matched by the record warm decades of the 1930s and 1940s. The few and scattered direct climate records from observations on the Greenland Ice Sheet also reveal a warming trend since 1985; the time span of the data series on the ice sheet.

As a result of increasing air temperatures, the mass balance of the Greenland Ice Sheet has changed: The interior, high altitude parts of the ice sheet have thickened because of accumulation from increased snowfall. The area above 2000 m has gained an average of 5 (±1) cm in altitude each year since 2000; adding 60 (±30) Gt of mass to the ice sheet each year.

The increment of accumulated mass is more than counterbalanced by the increased loss of mass by melting and by ice discharge into the ocean. The area experiencing surface melting is monitored by satellites and has increased significantly in extent since 1979, with a record-breaking surface melt area measured in 2007. In general, half of the total mass loss from the Greenland Ice Sheet is caused by surface melt and run-off. The annual gain in surface mass (i.e., received snowfall minus mass lost by melt), has a 50-year average value of 290 Gt, but has been reduced by 45 Gt over the past 15 years. This trend has been identified as being above the background variability caused by normal fluctuations in climate.

The margin of the Greenland Ice Sheet is characterized by very variable ice dynamics, as areas of slow glacial movement are separated by fast-flowing outlet glaciers and ice streams. Much of the ice from the fast-flowing ice streams and outlet glaciers is discharged into the ocean. Annual mass loss through ice discharge has increased by 30% during the recent decade; from 330 Gt in 1995 to 430 Gt in 2005. This increase was brought about by accelerated flow of outlet glaciers and the loss exceeds that caused by melt processes during the few recent warm years.

The total change in ice sheet mass can be assessed by comparing the surface mass balance (received snowfall minus mass lost by melt) with the loss of mass through ice discharge, by remote sensing of elevation changes, or by mass changes – detected as gravitational changes – measured from space.

A compilation of mass balance estimates reveals that the Greenland Ice Sheet is losing volume and that the mass loss has increased in recent years from 50 (± 50) Gt each year in the period 1995-2000, to 160 (± 50) Gt each year in the period 2003-2006.

Projections of future reactions of the Greenland Ice Sheet to climate warming indicate that the loss of mass will increase. Further, for the high Arctic region the IPCC climate scenarios predict temperature increases around 50% higher than those predicted globally. This will increase the length and intensity of the summer melt season and so will increase the extent of the area experiencing summer melt.

There are important uncertainties in the forecasts of climate and surface mass balance. The evolution of the surface mass balance demands a downscaling of atmospheric models tightly coupled with surface mass balance processes. However, the downscaling is not yet developed to a sufficient level. Current climate models project that Greenland's surface mass balance will become negative in the event of a global warming of 3.1 ± 0.8 °C (i.e., warming over Greenland of 4.5 ± 0.9 °C).

The additional mass loss through ice dynamics will cause Greenland's overall mass balance to become consistently negative. Current projections with coupled ice sheet and climate models indicate an annual average mass loss of the order of 180 Gt for the 21st century (i.e., 0.5 mm of global eustatic sea level rise annually). This is equivalent to a 5 cm sea level rise by 2100, primarily due to increased melting and run-off. Model estimates range from a sea rise of 1 to 11 cm.

Model predictions diverge for the period beyond 2100, as differences in ice dynamics and regional studies to date forecast an accelerated ice sheet decline in response to a 3 to 4 °C regional warming over Greenland, but with about 1000 years required for a 1 m contribution to global eustatic sea level rise from the Greenland Ice Sheet. The prognosis is very susceptible to the degree of sustained warming over Greenland. A temperature increase of 5-6 °C will lead to around 2 m of sea level rise by year 3000, while 10-12 °C or warmer would result in 5-7 m of eustatic sea level rise over the coming millennium.

Understanding of Greenland Ice Sheet dynamics has serious gaps limiting the ability to project the ice sheet's sensitivity to climate warming. The dramatic acceleration of many of the fast-flowing glaciers and ice streams is still poorly understood and, therefore, not fully implemented in ice sheet models. Over the past decades, observations have shown that ice discharge can double within one to two years – and may even be reversed. Investigations point towards the influence of relatively warm ocean currents on the stability of glacier termini and ice flux as well as the increased volume of surface melt water penetrating to the glacier base where water then acts as a lubricating agent enhancing ice flow.

The first attempts to include in the models the increasing ice discharge via the marine outlet glaciers have predicted an additional 4.7 cm of sea level rise by 2100. Doubling the discharge would raise the sea level 9.3 cm. Combined with the losses associated with surface mass balance, 10-14 cm of eustatic sea level rise from the Greenland Ice Sheet can be expected by 2100. It needs to be stressed that these models still suffer from a very incomplete understanding of the atmosphere-ocean-ice interactions.

The Greenland Ice Sheet changes will have both local and global impacts and will affect human societies significantly. Globally, the melting ice sheet will contribute to sea level rise. At present, the annual ice sheet loss of 110-210 Gt corresponds to 10-20% (i.e., 0.3-0.5 mm) of the 3 mm observed global sea level rise each year. Recent projections for global eustatic sea level rise by 2100 are as high as 1.0 ± 0.5 m. Total sea level rise also comprises thermal expansion of the ocean and melting of ice caps and glaciers, as well as the evolution of the Antarctic Ice Sheet.

Evaluating the risks and changes introduced by the projected evolution of the Greenland Ice Sheet is strongly related to how fast the changes, and hence sea level rise, will occur. This is highly unknown as it relates directly to the understanding of the processes accelerating the ice streams and fast-moving glaciers. These processes need to be studied much more intensely.

With increased meltwater run-off and iceberg discharge from the Greenland Ice Sheet, the salinity and density of the receiving marine surface waters are reduced, especially in the Greenland fjords and coastal waters. These changes will result in stronger gradients between the fresh surface water and the more saline water below and strongly impact on the primary production of the affected marine areas. The increased freshwater run-off may boost the hydrological circulation in the fjords, and by so doing significantly increase the transport of nutrients from the ocean into the fjords. As a result, primary production in coastal marine areas is expected to increase due to a combination of increased nutrient import and improved light availability following a reduction in sea-ice cover.

The magnitude of the increase in primary production depends on solar radiation, and the availability of nutrients and carbon dioxide as well as the vertical extent of the mixed layer. Climate-induced changes in primary production will propagate to higher trophic levels and ultimately affect populations of fish, seabirds and marine mammals with significant consequences likely for the structure, diversity, productivity and exploitation potential of the Greenland marine ecosystem. More research is, however, needed in order to quantify how the higher trophic levels will be affected by changes in the physical environment of Greenland waters, resulting from increased melting of the ice sheet. In order to create proper plans and to take science-based decisions for the future, cross-disciplinary research addressing these issues is in high demand and will be extremely important to the Greenland society, so economically dependent on marine resources.

Some of the increased meltwater volume from Greenland will reach the areas of deep water formation in the North Atlantic. However, the amount of freshwater from the Greenland Ice Sheet is an order of magnitude less than for other Arctic freshwater sources (i.e., the Arctic Ocean) and is therefore regarded to be of minor importance to changes in the global thermohaline circulation.

Changes in climate, in the ice sheet and in sea-ice conditions all have impacts on people and society in Greenland, and these changes are strongly interwoven with modernization and globalization processes. While less than a fifth of the Greenland population is economically involved in direct natural resource harvesting, the cultural identity of Greenlanders remains closely connected with the renewable resources. At present, most people in Greenland are involved in or connected to social services, administration, tourism, education, and similar activities. New prospects for large-scale mineral and hydrocarbon production and energy resource production are influenced by changes in accessibility and transportation. Further, hydropower, expected to be a growing energy resource for domestic consumption and new energy intensive industrial activities, will continue to be based entirely on a supply of meltwater from the Greenland Ice Sheet.

Over recent years, it has become very apparent that global warming is causing the Greenland Ice Sheet to lose mass. The Greenland Ice Sheet is the ‘awakening giant’ and increased melt and ice discharge from this ice sheet can seriously increase the predicted global eustatic sea level in the next century and, locally, can strongly impact on the socio-economic development of the Greenland population. Close monitoring of the mass balance of the ice sheet as well as coupled climate and ice sheet predictions are urgently needed and highly recommended as tools to facilitate improved forecasts of how to better adapt to a warmer world.

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List of Acronyms and abbreviations

ACIA	Arctic Climate Impact Assessment
AMAP	Arctic Monitoring and Assessment Programme
AMOC	Atlantic Meridional Overturning Circulation
AO	Arctic Oscillation
AOGCM	Atmosphere-Ocean General Circulation Model
ATM	Airborne Topographic Mapper
CO_2	Carbon dioxide
CliC	Climate and Cryosphere Project
CP	Area-averaged SMB sensitivity to precipitation
CT	Area-averaged SMB sensitivity to temperature
DMI	Danish Meteorological Institute
ECMWF	European Centre for Medium-Range Weather Forecasts
ERA-40	ECMWF re-analysis covering the ~40 year period 1957-2002
GIA	Glacial isostatic adjustment
GRACE	Gravity Recovery and Climate Experiment mission
HIRHAM	A regional climate model
HIRHAM-NAOSIM	A coupled regional atmosphere-ocean-ice model
hPa	Hectopascal (1 hPa = 100 Pa)
ICESat	A satellite laser altimeter
IPCC	UN Intergovernmental Panel on Climate Change
kyr	Thousand years
MAR	Modèle Atmosphérique Régional
NAM	Northern Annular Mode
NAO	North Atlantic Oscillation
PDD	Positive degree day
PDO	Pacific Decadal Oscillation
PMM5	Polar Mesoscale Model version 5
RCM	Regional climate model
RF	Re-freezing
SAT	Surface air temperature
SLP	Sea level pressure
SMB	Surface mass balance
SRALT	Satellite radar altimetry
SSMB	Specific surface mass balance
Sv	Sverdrup (1 Sv = 10^6 m^3/s)
TMT	Turbulent moisture transport
TWV	Total water vapor