ASPECTS ON INTERACTIONS BETWEEN MID- TO HIGH LATITUDE ATMOSPHERIC CIRCULATION AND SOME SURFACE PROCESSES

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Preface

This synthesis and collection of papers constitute a thesis in partial fulfilment of the requirement for the degree philosophiae doctor (PhD) in meteorology at the Geophysical Institute, University of Bergen, Norway.

This thesis focuses on some exchange processes between atmosphere, ocean and sea ice. A new data set of storm tracks over the Northern Hemisphere for the period 1948-2002 has been developed. An atmosphere only general circulation model is used and discussed. Model simulations are performed for the Last Glacial Maximum and for modern day. A model simulation with a tenfold increase in the vertical resolution near the surface has been performed to highlight the need for a sufficient representation of the Arctic planetary boundary layer in climate simulations.

Acknowledgement

I want to thank my supervisor, Nils Gunnar Kvamstø, for all help and support of my PhD work. I want to thank him for taking time and having patience with me and for giving me motivation for continued efforts.

During this work I have spent in total 4 months at International Arctic Research Centre (IARC), University of Alaska, Fairbanks. I wish to acknowledge IARC for their warm reception during my 3 visits there. In particular I give my gratitude to Vladimir Alexeev and John Walsh for arranging these visits and for good collaboration and interesting discussions during my stays there.

I also acknowledge IARC for the unique experiences with the summer schools arranged in 2004 and 2005. This included a visit to the Alaskan North Slope and a research cruise to the Laptev Sea with the Russian icebreaker "Kapitan Dranitsyn". The IARC summer schools have arranged good review lectures in current climate research by excellent professors and researchers. This has increased my understanding, motivation and interest for climate research. During the summer schools my personal network has expanded, new collaborations have been initiated and new friends have been found.

I wish to thank staff, students and colleagues at the Geophysical Institute and at the Bjerknes Centre for a good work environment.

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Abstract

The Arctic is a hot topic in Climate Research. A large number of signs of a warming Arctic Climate have been identified the latest years. This is of major concern in light of the increasing atmospheric content of greenhouse gases. The climate research community projects future warming of the climate in the high latitudes as a response to increased amounts of anthropogenic release of greenhouse gases since the pre-industrial era.

The overall objectives of this work has been to study the mid- and high latitude climate and climate variability, and to evaluate how well some climate processes that contribute to determine the Arctic climate and variability are represented and simulated in climate models.

A new data set of storm tracks trajectories and statistics over the Northern Hemisphere for the period 1948-2002 has been developed. The variability of the cyclones extending to the Nordic Seas is studied in particular, and it is found that both the number of storms and their intensity exhibits a strong decadal and interannual variability. The ocean volume transports into and out of the Nordic Seas shows a relatively close relation to the wintertime cyclone intensity and cyclone count.

To have confidence in future projections of climate, it is necessary to evaluate how the model behaves in a climate regime different from modern day. To do this two model simulations of the last glacial maximum (LGM) was performed. The reconstructions of sea surface temperatures in the Nordic Seas in LGM differ from perennial sea ice cover to having open ocean during the summer. The large scale atmospheric circulation patterns of the two different climate reconstructions are studied. It is found that the perennial sea ice cover produces a circulation pattern which may be too zonal to support the existence of the large north Eurasian ice sheets. In the case with seasonally open ocean the air masses carries larger amounts of heat and moisture towards the ice sheets and represents a larger degree of meridional circulation.

The current general circulation models, including several of those used by the IPCC, show considerable disagreement in simulating present day high latitude climate. This is of major concern and reduces the confidence in future model projections of high latitude climate. To investigate how turbulent vertical exchange processes in the Arctic boundary layer is represented by the climate models a simulation with high vertical resolution in the lower part of the atmosphere is performed. This reveals that the coarse vertical resolution commonly employed in the climate models are unable to reproduce important exchange processes in the Arctic boundary layer. In the case of our model this results in a warm bias over the Arctic Ocean. By increasing the vertical resolution we achieve a better representation of vertical turbulent exchange processes with the result of reproducing more realistic surface fluxes and surface air temperatures.

List of Papers

Paper I Sorteberg A., Kvamstø N. G. and Byrkjedal Ø. (2005). Wintertime Nordic Seas Cyclone Variability and its Impact on Oceanic Volume Transports Into the Nordic Seas. In: The Nordic Seas, An Integrated Perspective. *H. Drange, T. Dokken, T. Furevik, R. Gerdes and W. Berger*. Washington DC, American Geophysical Union. **158:** 137-156

Paper II Byrkjedal Ø., Kvamstø N. G., Meland M. and Jansen E. (2006). Sensitivity of Last Glacial Maximum Climate to sea-ice condition in the Nordic Seas. Climate Dynamics **26**(5): 473-487

Paper III Byrkjedal Ø., Esau I. and Kvamstø N. G. (2006). Sensitivity of simulated wintertime Arctic atmosphere to vertical resolution in the ARPEGE/IFS model. *Submitted to Climate Dynamics*

Chapter 1

INTRODUCTION

1.1 Motivation

The Arctic plays an important part of the Earth's climate system. Even though it is a relatively small geographical area it is a hot topic in climate research. The Arctic Ocean is covered by sea ice and surrounded by land, leaving only one deep passage, the Fram Strait west of Spitsbergen, to the other world oceans. The Arctic sea ice experience large annual variations, with a minimum extent in September and maximum extent in February (Figure 1.1). The area which experience an annual retreat and growth of sea ice is often referred to as the seasonal ice zone, and it plays a crucial role determining the atmosphere-ocean exchange processes in the Arctic. This is the region where the cold and dry polar air masses meets warm and moist marine environments and it is characterized by large instabilities which favour growth of high latitude cyclones. The ocean circulation patterns are a crucial factor in determining the position, extent and variability of the seasonal ice zone both today and in the past.

Climate records have shown large variations over geological times. The Last Glacial Maximum (LGM) was characterized by ice sheets covering large areas on the Northern Hemisphere (Figure 1.2). In the LGM large scale atmosphere and ocean circulation patterns were rather different from modern day. As the extent of the seasonal ice zone has the potential to alter the atmospheric circulation it represents a main challenge for our understanding of climate at mid and high latitudes. To make progress it is thus important to evaluate how well the processes that determine the seasonal ice zone variability and changes are represented and simulated in climate models.

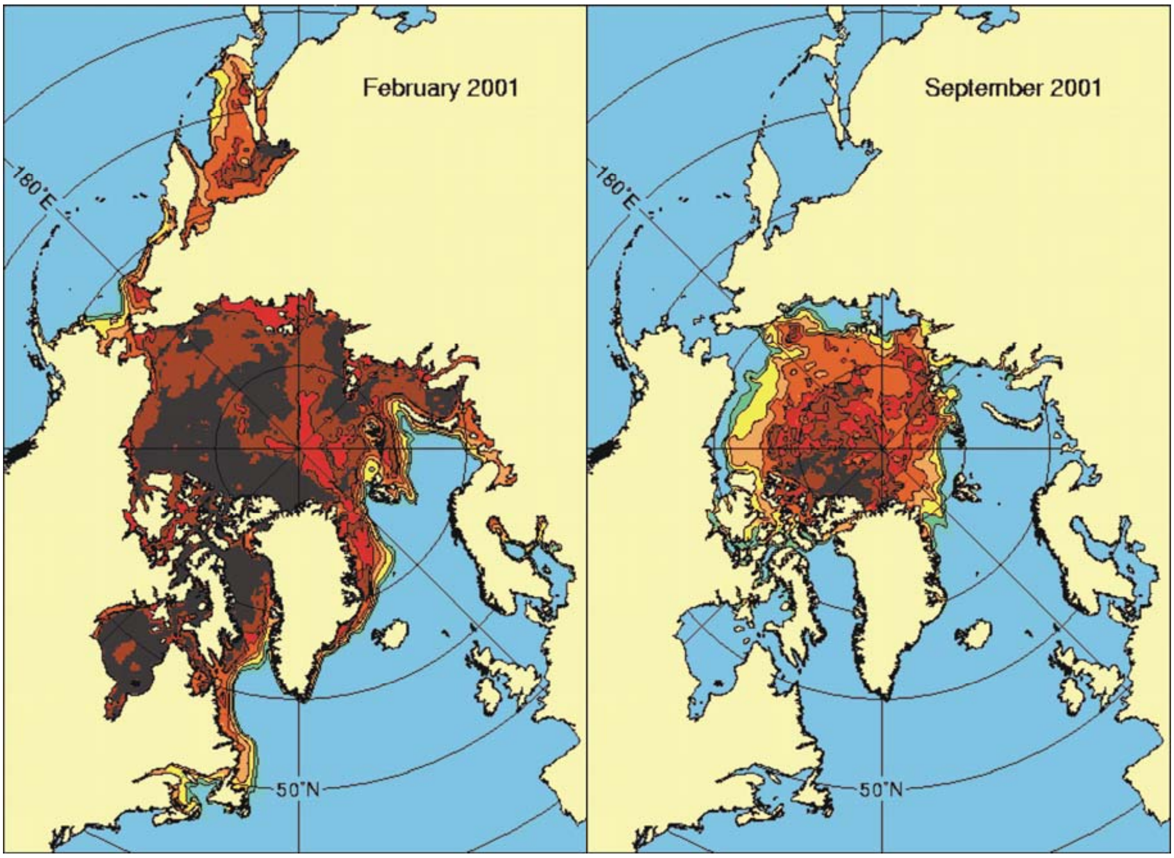

Figure 1.1: *Sea ice concentration in February and September 2001. Source: Bundesamt für Seeschifffahrt und Hydrographie: www.bsh.de*

The Arctic is ecologically very sensitive to climate changes. The seasonal ice zone in the shallow Barents Sea is an important area for biological production. This is to a large extent governed by long term variability and trends in the Atlantic inflow (Ottersen et al. 2000), and on the maximum winter sea ice extent (Loeng 1991). Simulations show that the annual biological production during a warm year can be about 30 % higher than during a cold year for the Barents Sea basin (Wassmann and Slagstad 1993).

In the latest years it has become evident that climate in the Arctic is changing. Large efforts have been made to better monitor and understand these processes. The instrumental record from the Arctic is relatively short and the spatial coverage is sparse. The historical records of Arctic climate are mainly from log books of Nordic fishers and whalers that reported the extent of the sea ice edge. Early efforts of studying Arctic climate were undertaken in the International Polar Years of 1882-1883 and 1932-1933 when a number of Arctic meteorological stations were established. The earliest systematic recording of climate in the Arctic Ocean is by the Russian North Pole drifting stations, the first launched in 1937.

The Arctic is a relatively small area on the Earth and changes in the Arctic mean temperature will not directly influence the global temperature. Based on the empirical models Budyko and Izrael (1991) the high latitudes show, however, a very high sensitivity to climate changes. According to that, the Arctic is the location where the first signals of a global warming can be identified. It is, however, important to bear in mind the uncertainties regarding this issue. Not only is the Arctic sensitive to global climate changes, the high latitudes also experiences large decadal variations which can be attributed to natural variability on regional scales (Hurrell and VanLoon 1997). The signal to noise ratio regarding climate change in the Arctic is thus relatively small. The global warming signal can be large, but the natural variability noise can be even larger. Efforts to study climate mechanisms in the Arctic are therefore of great importance and climate modelling represents a very useful tool in this respect.

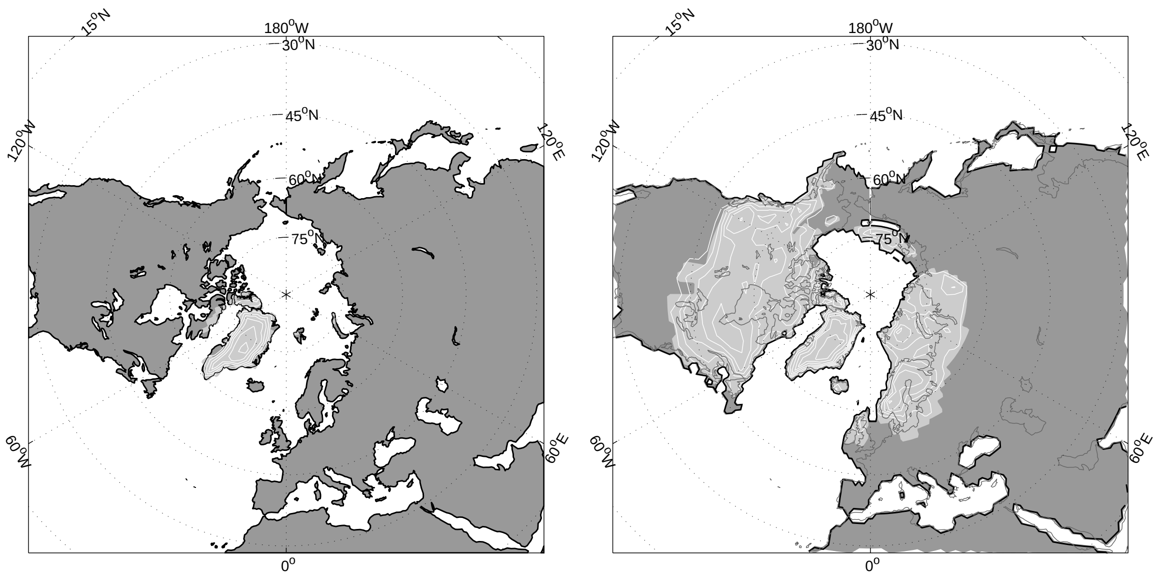

Figure 1.2: *Left figure shows modern day land-ocean distribution and extent and elevation of ice sheets. Right figure shows land-ocean distribution for LGM. Also shown are the ice sheets extent and elevation. Modern day coast line is added for comparison. Elevation of the ice sheets are drawn as white contours with an equidistance of 500m.*

As the Arctic sea ice plays an important role in the global energy budget, the positive ice-albedo feedback is of major concern. As the bright reflective sea ice area diminishes due to increased warming, the nearly black sea surface will absorb far larger amount of solar radiation, and thus contribute to an additional warming of the Arctic. Although a reduced sea ice cover in the Arctic may result in moister atmosphere with increased amounts of low clouds, further consequences of such a response are poorly understood. Increased cloud cover in the Arctic in winter represents a positive feedback as the clouds will insulate the Arctic and reduce the long wave heat loss from the surface. The clouds during summertime in the Arctic represent both negative and positive forcing on the surface temperature depending on the cloud properties (e.g. Shupe and Intrieri 2004).

As mentioned, the high latitudes is the region that is expected to experience the greatest climate change in a global warming scenario. Large polar sensitivity to climate changes has also been found in paleo records by e.g. Hoffert and Covey (1992). Polar amplification of climate change is found in climate models as a response to greenhouse warming (e.g. Holland and Bitz 2003; ACIA 2004; Sorteberg et al. 2005). The polar amplification is partly caused by the ice albedo feedback, as the polar ice cap is melting. But the amplification is also found in idealized model experiments with no sea ice by Alexeev (2003) and Alexeev et al. (2005). They explained the polar amplification in terms of changes in the large scale atmospheric dynamics.

Changes to the Arctic environment may have large impacts on a global scale. The evidences of climate change in the Arctic are numerous. Thinning and decreasing sea ice have been reported by Johannessen et al. (1999), Parkinson et al. (1999) and Serreze et al. (2003). The length of the growing season in the high latitudes has in some areas been extended by 3.6 days per decade over the last 50 years (Walther et al. 2002). Alaska is one of the regions that has shown large increase in the length of the growing season and also shows a large number of other signs of climate change (Hinzman et al. 2005). Instrumental records of temperature in the high latitudes show a positive trend. The strongest trend is found in winter with 2.3°C increase in surface temperature north of 60°N over the last 120 years according to Lugina et al. (2005). This increase is more than twice that found as an average for the whole northern hemisphere. Romanovsky et al. (2002) shows evidence of increasing permafrost temperatures from several locations in the Arctic, both on the American and the Eurasian side. These are all crucial signs of climate change which potentially have enormous ecological implications both regionally and globally. The prevalence of thermokarsts (Figure 1.3) in North America has increased since the late 1800s (Osterkamp et al. 2000).

Figure 1.3: *Collapsed Arctic Tundra (Thermokarst) in the North Slope of Alaska. The pictures are taken on August 13 2004. The lower picture shows a melting ice wedge.*

All climate models predict a polar amplification in a warming climate scenario and in greenhouse gas forcing experiments as those suggested by the Intergovernmental Panel of Climate Change (IPCC), but the spread among the models' responses is large (IPCC 2001; Holland and Bitz 2003). Regarding the mean global temperature increase to be expected in a warming climate, the spread among the models are modest. However the models' projections of the Arctic climate show a large spread which is of the magnitude of the warming signal itself, resulting in a small signal to noise ratio. There is also a large spread among the IPCC models with respect to simulation of present high latitude climate (IPCC 2001).

The large uncertainty and variation among the models in simulating the observed Arctic climate reduce our confidence in their future projections of the high latitude climate. Important small scale processes in climate can not be fully described in the model (limited by horizontal and vertical resolution), but rather by the use of parameterizations (e.g convective clouds, turbulence, vertical heat exchange). Also, several of the climate models in use (e.g. the ARPEGE/IFS model) have been developed to forecast weather on short timescales rather than for projecting climate.

Many of the processes that are parameterized in the climate models are not fully understood. The accuracy of the parameterizations differs among the models and validation of the model performance may sometimes give the 'right answer for the wrong reason'. The validation of the models is mostly focused on mid and low latitude climate for several reasons. These are the regions where most people live and therefore require most detailed weather forecasting and these are also the regions where most of the weather observations are done. The parameterization chosen for certain variables or processes may not be valid for the high latitude climate. In addition to large natural variability, he large spread among the different climate models are mainly a result of different parameterizations. Studies of parameterized exchange processes in the Arctic and evaluation of their implementation in the climate models are thus of great interest to Arctic climatologists.

1.2 Objectives and outline of the thesis

The overall objectives of this work has been to study the high latitude climate and climate variability, and to evaluate how well some climate processes that contribute to determine the

Arctic climate and variability are represented and simulated in climate models. The approach to this overall objective has been the following:

Firstly, to study the interannual and decadal variability of the synoptic activity in the Nordic Seas, and further to determine to which degree the atmospheric processes influences the oceanic volume transports into the Nordic seas. This is one of the climate processes that determine the seasonal ice zone variability and long term change in the Nordic Sea region as the Nordic and Barents Seas are the main gates for atmospheric energy transports into the Arctic Ocean. The results are reported in Paper I.

Secondly, to evaluate a general circulation model. The model has been used as a tool to study past climate. To have confidence in future projections of climate, it is necessary to evaluate how the model behaves in a climate regime different from modern day. This work has involved studies of the atmospheric response to changes in the Last Glacial sea ice cover, and to explain whether changed synoptic activity due to a changed reconstruction of sea ice may clarify how the large Barents and Kara sea ice sheets were formed. The results are reported in Paper II.

Thirdly, a focus has been to investigate how processes in the Arctic atmospheric boundary layer are represented in the general circulation model. Several studies suggest that the vertical heat exchange in the stable boundary layer is poorly represented in climate models. Efforts have been made to describe these discrepancies and to suggest where future modelling efforts should be put. The work has employed increased vertical resolution of the model. The underlying hypothesis is that the coarse vertical resolution is a major contributor to the biases in the stable boundary layer. The results are reported in Paper III.

The thesis consists of two parts:

Part I presents a background and overview for the thesis. This introduction is the first chapter in Part 1. Chapter 2 presents an overview of the geographical regions in this study. Chapter 3 presents a general description of climate modelling and describes its advances and uncertainties. An overview of climatology is given in Chapter 4. This includes a description of the drivers of climate on long timescales, a description of our current understanding of the LGM climate. The climatology of the Arctic and North Atlantic is also presented here. Chapter 5 gives a description of the tools used in this study. Chapters 2-5 thus describe the background theory for the 3 specific studies summarized in Chapter 6. The observation and reanalysis data used are presented in Chapter 7.

Part II of the thesis consists of the three papers that document the main scientific work.

Chapter 2

THE AREA OF INTEREST

2.1 The Nordic Seas

The Nordic Seas (Figure 2.1) includes the Norwegian Sea, Greenland Sea, Iceland Sea and western Barents Sea (Hurdle 1986). North of the Nordic seas the main ocean gate to the Arctic is found. This is the Fram Strait west of Spitsbergen. In the south the Nordic Seas are limited by the Greenland-Scotland Ridge.

The Nordic Seas region is very much warmer than other locations at the same latitude. The surface air temperature in the Nordic Seas region is on average 10-20K higher than the zonal mean at this latitude. This is be partly caused by the warm Gulf Stream or the North Atlantic Current system that brings warm and saline water from the Gulf of Mexico and into the Nordic Seas. The second reason for the high temperatures in this region is the prevailing moist westerly and southwesterly winds that bring heat into the area. The upper mixed layer of the ocean is also warmed during the summertime. This heat is released during winter and keeps the winter temperature in the Nordic Seas high relative to its latitude.

The southern and eastern part of the Nordic Seas contains warm and saline Atlantic water transported by the Gulf Stream. Cold and fresh water from the Arctic Ocean flows into the Nordic Seas through the Fram Strait. This makes the Nordic Seas an area of frontal mixing and thus large variability.

Sea ice is found in the northern and western part of the Greenland Sea during winter. In summertime the ice edge pulls back into the Arctic Ocean (Figure 1.1). The area that is defined by the summer minimum and winter maximum sea ice extent is called the seasonal ice zone. This is a nutrient rich area and spawning ground for Arctic biology.

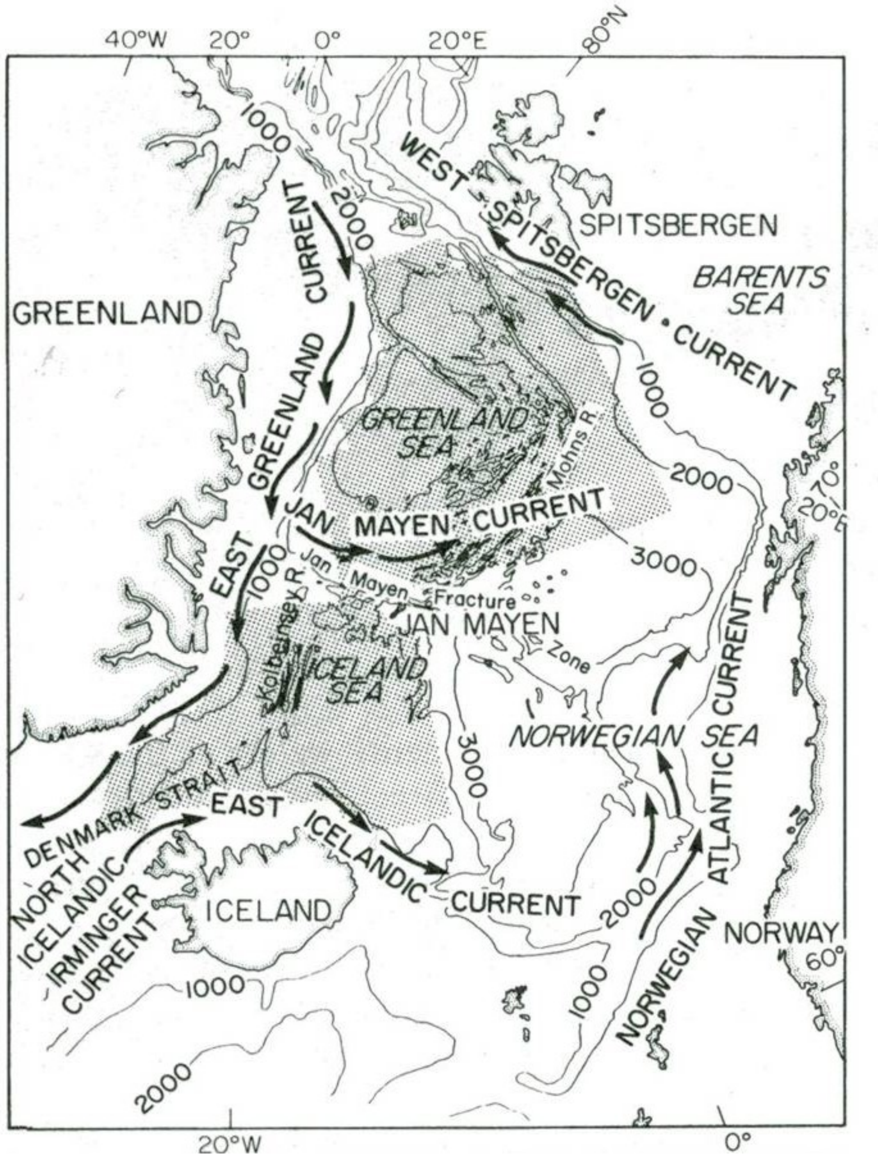

Figure 2.1: *The Nordic Seas with outline of land and sea areas, currents and names (from Swift 1986).*

The Nordic Seas is a sink for atmospheric CO_2. The ocean uptake of CO_2 in the Nordic Seas ranges from 20-85$gCm^{-2}y^{-1}$. This is among the highest carbon fluxes among the world oceans (Anderson et al. 2000; Takahashi et al. 2002; Skjelvan et al. 2005).

Furevik and Nilsen (2005) summarize 4 reasons for the large interest in the Nordic Seas over the latest years. **Firstly**, a strengthening of the westerlies has been observed since the 1960s. The strength of the westerlies has a profound impact on the winter climate in Europe. In the years with strong westerlies southern Europe has experienced cold and dry winters, while Scandinavia and northern Europe have had mild and wet winters (Hurrell 1995). **Secondly**, the Greenland Sea is an important location for deep water formation. This deep water formation has been strongly reduced in the period of the strengthened westerlies (Dickson et al. 1996; Hansen et al. 2001). **Thirdly**, a reduction of the northern hemisphere sea ice extent has been observed over the latest years (Johannessen et al. 1999), in addition to a reduction of the thickness of the sea ice (Rothrock et al. 1999). A corresponding warming of the Arctic Ocean has also been observed (Carmack et al. 1995; Grotefendt et al. 1998). **Fourthly**, there has been observed freshening of surface and intermediate water of the Nordic Seas, and also freshening of the overflow water and deep water.

Because of the warm Atlantic water entering this region, and because of the strong winds in the area, the Nordic Seas experience substantial fluxes of latent and sensible heat from the ocean to the atmosphere. Along the seasonal ice zone, the atmospheric temperatures during winter may be very cold and dry. In the Nordic Seas, a large energy exchange from the ocean to the atmosphere takes place, especially along the ice edge. The average seasonal fluxes from the surface to the atmosphere in the Nordic Seas ranges from 160 Wm^{-2} to 180 Wm^{-2} during winter with maximum fluxes along the ice edge exceeding 300 Wm^{-2}. Simonsen and Haugan (1996) estimated the energy loss from the ocean to the atmosphere to be 242 TW in the Nordic Seas including the Barents Sea. The Barents Sea alone looses 136 TW. In comparison, the Arctic Ocean heat loss is 86 TW.

The large energy exchange makes the Nordic Seas an area with high baroclinicity and thus a synoptically active area. Depending on the strength of the westerlies, storms are steered into the Nordic Seas from the south. The major storm track in the Nordic seas originates in the south around Iceland and the Icelandic low. A large amount of the storms moves northeast and into the Nordic Seas, and also to some degree further north into the Arctic. In the years with strong westerlies (high NAO) more storms are steered into the Nordic Seas (Paper I). The storms from the Nordic Seas and the advection of air from the Nordic Seas and into the Arctic Ocean are major sources of heat and moisture to the Arctic atmosphere.

The Nordic Seas is the main gate of the meridional energy transport from lower latitudes into the Arctic. The meridional energy transport takes two pathways, either as ocean transport or atmospheric transport. Vonder Haar and Oort (1973) estimated that the atmosphere and ocean contributes equally to the meridional energy transport. Estimates by Trenberth and Caron (2001), however, suggest that the atmospheric component contributes by a far larger portion of the poleward heat transport than the ocean. The influx of energy from the Ocean into the Arctic is mainly trough the Nordic and Barents Seas. The warm North Atlantic current flows trough the Nordic Seas and into the Arctic Ocean as two branches. One branch is the West Spitsbergen Current, the other is the Norwegian Atlantic Current which flows via the Barents Sea and into the Arctic Ocean. We know that the atmosphere forcing is the main driver of variability in this current system. However, the relative importance of different atmospheric phenomena is not known. To increase our knowledge on this issue is therefore of great interest.

2.2 The Arctic Ocean

The northernmost Ocean is the Arctic Ocean (Figure 2.2). The Arctic Ocean is covered by a seasonally varying cover of sea ice. This has a minimum extent in late summer and extends to the surrounding coasts in late winter (Figure 1.1). The land masses that surrounds the Arctic Ocean is characterized with Arctic tundra as in Northern Siberia and Northern Alaska (North Slope) and Canada (the Yukon and North West Territories). The island of Greenland is covered by a great ice sheet with an elevation exceeding 3000m. The Arctic Islands of Svalbard, Novaja Semlja, Franz Josef Land and the islands of the Canadian Archipelago are typically mountainous. The Arctic Climate is harsh. Only few people live in the Arctic. This is primarily native people that through generations have adapted to the Arctic climate and live by existence farming, hunting or fishing.

The Arctic Ocean is connected to the other worlds oceans through three straits. The Bering Strait connects the Arctic Ocean to the Pacific. This is a shallow strait with an average depth of only 40m. Water is flowing from the Pacific into the Arctic Ocean trough this Strait. The second gate is through the Canadian Archipelago and Davis Strait to the North Atlantic. Water from the Arctic Ocean is exported through this strait with an amount estimated in the range 1-2 Sv (1 Sv = $10^6 m^3/s$) (Simonsen and Haugan 1996). The third gate is trough the Nordic Seas, either through the Fram Strait, where warm Atlantic Water flows into the Arctic as the west Spitsbergen current, or via the Barents Sea and the warm Norwegian Atlantic Current. The main export of cold and fresh polar water from the Arctic Ocean is through the Fram Strait, estimated to be 3 Sv (e.g. Worthington 1970).

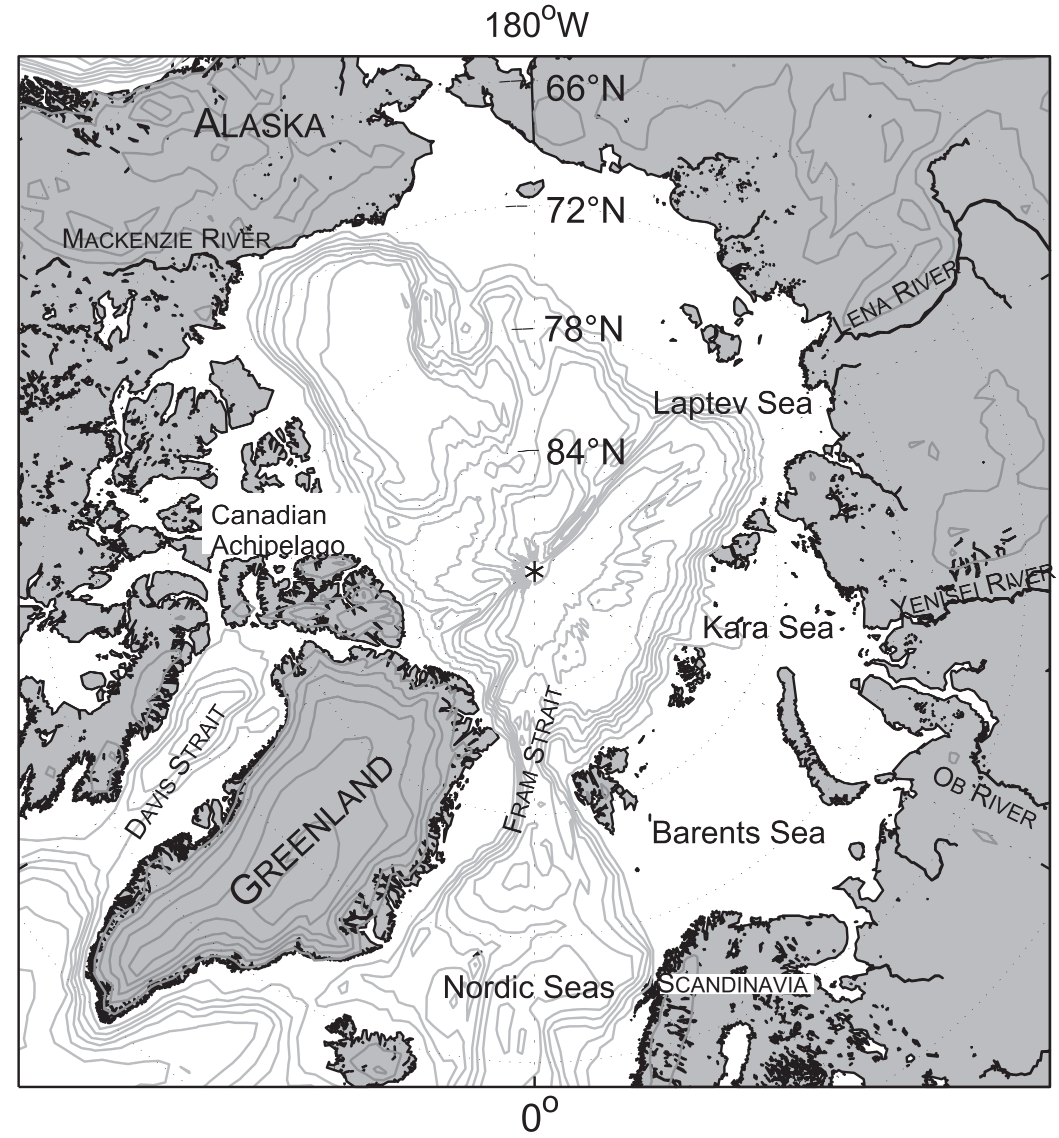

Figure 2.2 *The Arctic Ocean. Bathymetry and topography contours are drawn with an equidistance of 500m*

The Arctic Ocean contains approximately 1% of the total ocean volume on Earth, while it receives 10% of the world river input of fresh water. This makes the surface waters of the Arctic Ocean relatively fresh with an average salinity of less than 30psu in the summer. The large Siberian rivers, Lena, Yenisei, Ob drains the Asian continent north of the Tibetian Plateau into the Arctic Ocean. The Canadian Mackenzie River drains 1.8 million km^2 of the North American continent into the Arctic Ocean. Changes in the river runoff will change the salinity of the Arctic Ocean and can have the potential to modify the Meridional Overturning Circulation (Aagaard and Carmack 1989).

Chapter 3.

CLIMATE MODELLING

3.1 General description

An important tool in studying physical climatology is the climate models. The general circulation model (GCM) is a climate model that can describe atmosphere processes in 3 dimensions and the development of climate by time. The GCM is a simplified mathematical representation of the Earth's climate system (e.g. McGuffie and Henderson-Sellers 1987).

The driver of the climate model is, as in the real world, solar radiation. The surface and atmosphere respond to solar radiation, which is all the time changing, by adjusting surface temperature or moisture content in the air. The tropics will absorb the largest amounts of energy. Temperature differences that arise generate movements in the atmosphere. The climate system must redistribute this energy to achieve climate equilibrium. This sets up an energy transport from the tropics toward the poles. The atmospheric motions are described by the Navier-Stokes equations (e.g. Kundu and Cohen 2004).

The governing equations (the Navier-Stokes equations) for the atmosphere can be expressed analytically in its traditional form. However, digital computers are unable to read and calculate the continuous equations. In mathematics, discretization describes the process of transferring continuous equations into discrete counterparts. This process is carried out as a first step toward making them suitable for numerical evaluation and implementation on digital computers. On a computer it is necessary to divide the space and time into discrete pieces. Space is therefore represented in the climate models as grid cells. The Earth is divided into a finite number of grid cells in latitude and longitude direction. Calculations are performed on a plane of grid cells at the surface and also on discrete number of other planes of grid cells representing different levels in the atmosphere. Typically the Earth can be represented by

5,000-20,000 grid cells horizontally, and in 15-60 levels vertically in a climate model. This represents the spatial resolution of the model. The complexity and computational costs increase as the number of grid cells resolved by the model increases. The time dimension must also be discertizised, time is therefore broken up in timesteps. The simplest example of a discretization is the way a time derivative of a variable X can be expressed:

$$\frac{\partial X}{\partial t} \approx \frac{X^{t} - X^{t-1}}{\delta t} \tag{3.1}$$

Here X^t represents the value of X at time t, whereas X^{t-1} represents the value of X one timestep back in time. The timestep is given as δt, and expresses the time resolution of the model. Following this equation the tendency in the variable ($\partial X / \partial t$) can be computed. This tendency will then be used to project forward in time to achieve the value for the next timestep.

The model computes the climate by starting at an initial state of the atmosphere. At the initial state all possible variables are given at each grid point and in all levels. This is a frozen image of the atmosphere. The techniques of discretization are then used to calculate tendencies or derivatives in all these variables. This is projected one timestep ahead in time (e.g. 900 seconds or 1800 seconds). This gives a new snapshot of the atmosphere which again can be used for the next projection. From this we get a series of snapshots that gives the time evolution of the climate or weather. This can be compared to movie pictures which is also a series of snapshots put together to form a motion picture.

Many climate models are based on models developed for forecasting. This is the case for the model used in this study, the ARPEGE/IFS model (Deque et al. 1994). A closer description of this tool is given in Section 5.1. There are differences between a model used for climate studies and those used for weather forecasting. They are developed to fulfil two different needs and therefore they have different strengths and weaknesses. A forecast model demands high resolution as the forecasters are interested in the weather on a regional scale. Timing and geographical location of specific weather events are important. Forecast models are run for short timescales, typically 10 day forecasts are simulated. The skill or accuracy of the forecasts decreases with forecast length and is usually low beyond 3-4 days. To the forecasters it is therefore not important whether energy is conserved or not in the model. It never is on such short timescales. However, for climatological studies, energy conservation is the key factor. If the energy is not conserved over long time, this leads to a drift in the climate model. A drift is when the average value of a parameter globally or regionally decrease or increase slowly by time. Drift in the model will pollute a long term model projection. To investigate possible drift in the models, long control simulations (several hundred years) with fixed (annual varying) boundary conditions are performed. To avoid drift in the model parameterizations are tuned differently for climate models and forecast models. This will be described more thoroughly in Section 3.3.

Small scale processes are important in climate. Such processes are often referred to as sub-grid processes in climate modelling. They happen on such a fine scale that they can not be resolved by the model grids. They are nevertheless important to climate and must be accounted for in the model. Turbulence and clouds are examples of such processes which must be parameterized in the models. A discussion of common assumptions behind some parameterizations is given in Section 3.3.

The first climate model projection was performed by Arrhenius (1896). Arrhenius constructed a radiation and energy balance model. By hand he calculated the change in global temperature that could be expected by changing the atmospheric content of CO_2. His model results predicted about 5°C increase in temperature by doubling the CO_2 content of the atmosphere. The predicted increase in the global temperature found by Arrhenius is larger than the range (1.1-3.1°C) of the results from doubling CO_2 experiments performed with full complexity GCMs of today (e.g. IPCC 2001). Extending into circulation models, one of the first climate model experiments was performed in the 1950s by Phillips (1956). It had a two-dimensional grid in the atmosphere above a uniform surface. With this model he was able to simulate the basic character of the general circulation of the atmosphere. The models have since developed further based on the increasing knowledge of the atmosphere physics and dynamics, and of the interactions within the climate system. The current state of the art GCMs are able to simulate atmosphere, land and ocean, and interactions between them. They include atmospheric aerosols and also carbon cycles. Vegetation models and atmospheric chemistry is also beginning to be included in the models. The advance of faster computer systems allows for increased complexity in the models and also for increasing the resolution of the models. High resolution simulation on global scales has been simulated on the Japanese Earth Simulator. The Earth Simulator community simulates climate with a global resolution as high as 10km (Mizuta et al. 2006).

3.2 Climate Feedbacks

The climate of a region is a result of long term atmospheric equilibrium. Whereas one winter to the next may be relatively similar as will the winters in one decade to the next, there can be large day to day variations. The atmospheric state on short timescales (days to weeks) is generally not in equilibrium, but has a rather chaotic nature. This is the weather which is characterized by the synoptic timescale of days. The term **climate** incorporates all timescales from the high day to day synoptic variability that characterizes a region through the long term variability to the average mean state of the region.

The surface temperature equilibrium from an atmospheric point of view can be expressed through the surface budget in the following way:

$$\frac{\partial T_s}{\partial t} \propto \sum F = F_{SW} + F_{LW} + F_{SH} + F_{LH} + F_C \tag{3.2}$$

The change in temperature at one location on the Earth's surface is proportional to the sum of energy fluxes at the surface. F_{SW} is the net short wave energy absorbed by the surface, this is the incoming solar radiation subtracted the reflected part of the short wave radiation. The net flux of short wave energy will always be positive. F_{LW} is the net flux of long wave radiation, which is the long wave radiation received at the surface from the atmosphere subtracted the long wave radiation emitted from the surface. F_{LW} is generally negative. F_{SH} is the flux of sensible heat received by the surface from the atmosphere. The sensible heat flux can be either positive or negative depending on whether the surface is colder or warmer that the overlying air. F_{LH} is the flux of latent heat received by the surface. The latent heat flux is generally negative and represents evaporation of moisture at the surface to the atmosphere. F_C is the conductive heat flux from below the surface. This represents conduction of heat through the sea ice in the Arctic. It represents storage of heat in the ground over land and it represents the heat exchange processes between the ocean water masses and the ocean surface. The conductive flux can be both negative and positive. Over land it represents storage of heat in the ground and is thus typically negative in summer and positive in winter. The sum of all the terms represents the equilibrium state of the surface temperature.

If the sum of all contributions is different from zero averaged over a long period of time, this will result in a trend in the climate, and we will experience a change in surface climate ($\overline{\partial T_s / \partial t} \neq 0$). If a change to the equilibrium state is introduced as a change in the climate forcings, the response can be separated in a direct effect and an indirect effect. Aerosols, small particles in the air, exemplify this: An increase of aerosols in the atmosphere will have a direct effect on solar radiation either by increased absorption or increased scattering or reflection of the solar radiation. The indirect effect from increased aerosols relates to their function as cloud condensation nuclei, on which atmospheric moisture can condensate and form clouds. The indirect effect of aerosols on solar radiation is thus a reduction of solar radiation due to increased cloud cover.

The direct effect on the surface heat budget will typically be that the surface temperature responds to changes in one of the surface fluxes, as expressed in Equation 3.2. The indirect effect may be just as important to the climate. The indirect effect can have similar, smaller or even larger magnitude than the direct effect, and it can be either positive or negative. The indirect effect can also consist of several different mechanisms that finally lead to a modification of e.g. the responding surface temperature. The indirect effect is commonly referred to as the **climate feedback**. The feedback is the system's ability to modify the response to the initial forcing. A climate feedback can be either positive or negative (Kellogg 1983).

A positive feedback is when the indirect effect from the change in the forcing represents an additional amplifying response to the climate equilibrium. For a strong positive feedback this can represent a feedback loop that strengthens itself and imposes a large modification to the climate equilibrium. While for a weaker positive feedback the system will more easily be modified to a different equilibrium state.

The negative feedbacks work opposite of the positive. The negative feedback represents an indirect effect to a forcing that is opposite to the direct effect. The negative feedbacks thus act as stabilizers of the climate equilibrium, and thus damp the response to the change in the forcing.

Figure 3.1: *The picture illustrates the difference in albedo between sea ice and open water. The picture is taken during the summerschool hosted by International Arctic Research Center (IARC) onboard the Russian Icebreaker Kapitan Draitsyn September 2005. The picture is taken in the Laptev Sea at around 80°N 120°E.*

The advance of climate models has allowed us to investigate feedback mechanisms in climate in more detail. The following summarizes the most important feedback mechanisms associated to increased radiative forcing by e.g. greenhouse warming which is currently threatening the global climate equilibrium. The atmospheric feedbacks that are described here are all incorporated in climate models. Their magnitude in the models has been studied by Colman (2003).

In a warming climate a positive feedback with the potential to amplify the global warming is the ice-albedo feedback. When the global climate is warming this will melt the high latitude ice sheets and in particular melt the Arctic sea ice which has a wide aerial extent, but is relatively thin (on average 2-3m). Snow and ice generally reflect large amounts of the incoming solar radiation. Snow and ice have high albedo compared to the open ocean (Figure 3.1). When the sea ice melts and its aerial extent becomes smaller, the ocean area increases. Thus a larger amount of the solar radiation is absorbed by the surface and more energy is introduced to the system. The net flux of short wave radiation increases. The surface energy budget becomes positive, which corresponds to an additional increase in the surface air temperature. This describes a positive feedback loop which has the potential to accelerate with time if not balanced by other effects. Melting of Arctic sea ice can have a relatively short timescale since the sea ice cover is relatively thin.

A warmer atmosphere can contain a larger amount of water vapour. The water vapour represents the largest greenhouse effect of all the greenhouse gases, but it has a far shorter recycling time in the atmosphere. In warmer climate the average vapour content in the atmosphere has the potential to become larger, and thus to contribute to increased warming of the planet. From a synthesis of model experiments, Colman (2003) found the water vapour feedback to be 3-4 times larger than the ice-albedo feedback on a global scale.

A moister atmosphere can also promote the formation of more clouds. Clouds and especially low clouds effectively blocks incoming solar radiation. This can have the potential of reducing the initial warming of the climate, and thus represent a negative feedback. However, high clouds may constitute a positive forcing, since they are relative transparent to solar radiation while they are good absorbers of terrestrial long wave radiation. The effect of the cloud feedback on the global climate is highly uncertain according to IPCC (2001). The feedback can be both positive and negative. In the climate models Colman (2003) found generally a positive feedback effect from clouds.

In a warming climate, a large part of the energy added to the surface is compensated by increasing the evaporation rather than increasing the surface temperature. This is generally the case in the lower latitudes. The latent energy in the excess water vapour is released higher up in the troposphere. The mid troposphere thus warms more than the surface. This represents a reduction of the average atmospheric lapse rate (how the temperature changes with height). In model experiments with 2 % increase in the solar radiation Hansen et al. (1984) found a 4°C increase in the surface temperatures at low latitudes, while temperature in the higher troposphere increased by up to 8°C. The reduced lapse rate makes the long wave opacity of the water vapour less effective in absorbing the terrestrial radiation. The increased water vapour in the troposphere also lifts the effective radiating height to a higher level with correspondingly lower temperatures which will reduce the long wave emission from the atmosphere (Colman 2001). The net contribution from the decreased lapse rate is to increase the amounts of terrestrial radiation that is emitted from the Earth. As a global average, this represents a negative climate (Colman 2001). The lapse rate feedback is generally found to be negative in the low latitudes and positive at high latitudes (e.g. Pollack et al. 1993; Colman 2001). Colman (2003) found the lapse rate feedback to be generally negative in the climate models, but relatively small compared to the sum of the other feedback mechanisms. The sum of all these feedback mechanisms in the climate models is positive. All models will therefore show a net positive feedback and thus additional increase in surface air temperature in response to the increased greenhouse gas forcing.

Compared to the complexity of the climate mechanisms it is clear that most models simulate modern climate fairly well. The models' climate equilibrium is very sensitive to the representation of the climate forcings. The models have generally been adjusted, or tuned, to fit the observational record. However, to study climate changes with these models one must be certain that all the feedback processes are represented correctly. If not, a bias or an erroneous representation of a feedback process will lead to a bias in the equilibrium state for the new climate. The models show a relatively large spread in the magnitude of each of the feedback processes (Colman 2003). This is also manifested in the spread in global surface air temperature projected in global warming scenarios by the different models.

3.3 Parameterization of sub-grid processes

Parameterization is a technique used to describe atmospheric processes with simplified equations that give a reasonable representation of the processes. There are several reasons why parameterizations are employed in the models. The first and most important reason to implement a parameterization is the atmospheric processes in question happen on finer time and spatial scales than those resolved. Turbulence and cloud formation are such processes, often referred to as sub-grid processes.

Another reason, related to the previous one, to parameterize processes is that they are too complicated for the computers to calculate in reasonable time. Turbulence is an example of such a mechanism. The choice of turbulence closure often is based on the consequential simulation speed of the model. Radiation physics is another process which is simplified on this basis. The radiative properties are dependent upon the wavelength of the light. This is in climate models commonly only represented by two distinct wavelengths, long wave, representing the terrestrial radiation, and short wave representing the solar radiation. To include several channels or wavelength bands would heavily increase the simulation time of the model.

Thirdly, many processes observed in the nature are not fully understood, and can not be expressed in a single unconditionally accepted equation. There are clearly uncertainties connected to our current understanding of atmospheric processes as well.

The simplifications represented by the parameterizations thus add a degree of uncertainty to the simulations. The main differences between the climate models, and their representation of climate, relates back to different choices of parameterizations employed in the models.

A commonly used example of a parameterization is the so-called K-theory concerning atmospheric turbulence and exchange processes. The theory assumes that the vertical energy fluxes associated with turbulence can be estimated as a constant or function, K, multiplied by the resolved vertical gradient of the variable in question (Paper III). The analytical formulation of the vertical turbulent fluxes becomes increasingly more complicated the more accurate you wish to express them (Stull 1989). In order to reach an estimate of the flux, it is necessary to define the degree of complexity to your calculations and assume the remainder can be expressed by the K-theory. This is the called the closure of the equation. The number of terms involved in calculating the turbulent fluxes increases exponentially as the calculations are made more accurately. Therefore a first order or 1.5 order closure is often employed in climate models. The value of K is often expressed as a function depending on different variables in the model.

A wide range of atmospheric processes are parameterized in the models. To express a parameterization it is always necessary to make a number of assumptions. The parameterizations often use a large number of constants with no physical relationship to the process to be parameterized. This gives a large number of constants without any justified linkages to the atmospheric physics. These constants are then tuned to minimize the biases in the models compared to the observed climate. The parameterizations are a necessity to the models but also constitute a substantial degree of uncertainty.

Chapter 4

CLIMATOLOGY

4.1 Long term variations of past climate

The Earth has experienced large climate variations during the last million years. Earth has experienced periods of warmer and colder climate than today, caused by changes in the main driving mechanism of Earth's climate, the solar insolation. The cycles of past climate on Earth was studied by Milutin Milankovich in the beginning of the 20^{th} century. Milankovich (1920) formulated the changes in the Earth orbit around the sun mathematically and defined tree orbital parameters that cause changes in the Earth's climate on long time scales: the axial tilt, the precession and the eccentricity of the Earth's orbit. The different orbital parameters follow different cycles. The cycles described by Milankovich (1920) has later been validated by the paleo records (Hayes et al. 1976).

The first orbital parameter is the tilt of the Earth axis and varies between 22.1° to 24.5°. At present the tilt is 23.5° relative to the normal to the plane of the ecliptic. The tilt of the Earth completes a full cycle during 41,000 years. A large tilt corresponds to larger differences between summer and winter climate on mid and high latitudes. The change in the axial tilt is often referred to as the Earth's obliquity.

The second orbital parameter describes the position of aphelion and perihelion, the precession of the equinoxes. The Earth's orbit around the Sun is elliptic and not circular. The Earth will at one position along its orbit be at a minimum distance to the sun, perihelion. The opposite position, when the distance is at a maximum, is defined as aphelion. The position of the apsis (perihelion and aphelion) is important in describing the climate in combination with the axis tilt. If Earth is at perihelion at summer solstice on one hemisphere, this hemisphere will experience relatively warm summers. The other hemisphere, which will be at aphelion at its summer solstice, will experience relatively colder summers. The position of the apsis completes a full cycle in 26,000 years. In addition the elliptical plane rotates with a cylicity of 70,000 years. This is called the inclination of Earth's orbit, and was not included by Milankovich's studies. The combined effect of the inclination and the precession results in a 23,000 year orbital cycle.

The third orbital parameter is the eccentricity. The eccentricity can be calculated from the length of the two axes of an ellipse. For a circle the eccentricity is zero. The eccentricity is for an ellipse between 0 and 1. At present the difference in the Earths distance from the sun at aphelion and perihelion is approximately 3.4%. This represents an eccentricity of the orbit of 0.0167. Trough time the eccentricity has varied from nearly 0 to 0.05 with cycles of 100,000 and 400,000 years. Changes in the eccentricity acts to increase or decrease the effect of the precession parameter.

The key factor that determines whether an ice cover is formed is the summer temperatures. The process of melting an ice sheet (ablation) is much faster than the building (accumulation) of an ice sheet. To create an ice sheet, accumulation of snow is important but it is even more important that it does not melt away during the subsequent summer. If the accumulated snow survives the summers, and snow keeps accumulating over several hundreds of years, an ice sheet is formed. However, if the winters become colder, this will tend to produce less amounts of snow which is more easily melted away during summer. Referring to the tilting parameter, colder winters will often correspond to warmer summers.

4.2. The Last Glacial Maximum

About 20,000 years ago Earth experienced the last glacial period (Figure 1.2). This period is known as the last glacial, which had its maximum extent, the Last Glacial Maximum (LGM) 18,000-21,000 years ago. The global temperature at the LGM is believed to be approximately 5°C colder than modern day climate. The largest differences from modern day climate were found at the mid and high northern latitudes, where large ice sheets covered the landmasses. In the tropics the temperature was only 1-2°C colder than modern day. However, the solar radiation at the LGM was not very different from present. The annual insolation at the top of the atmosphere was approximately 2% weaker at the poles, while 0.2% stronger at the equator compared to present day solar radiation. The accumulation of the large ice sheets and the reduction of the global temperature is the result of reduced solar radiation over the 100,000 years prior to the LGM (Andersen and Borns 1994; Ruddiman 2000).

North America was covered by two large ice sheets during the LGM. The Laurentide ice sheet lay east of the Rocky Mountains and covered North America north of 40°N. The Laurentide Ice sheet is believed to be up to 3000m thick (Peltier 1994). The Cordilleran ice sheet was situated west of the Rocky Mountains. These two ice sheets covered all of North America north of 45°N except the inland of the modern day Alaska where the climate was too dry to form any large ice sheet. Scandinavia was also covered by a large ice sheet. The Fennoscandian ice sheet had an assumed elevation exceeding 2500m (Peltier 1994). The Barents and Kara Seas are also believed to be covered by ice sheets during the Last Glacial. Large amounts of water were bound up in these ice sheets, estimated to be 55-64 million km^3 (Denton and Hughes 1981). The excessive amounts of ice on land combined with lower ocean temperature reduced the ocean level by approximately 110-130m. This made the land sea distribution during the LGM rather different from today. Land that is now covered by ocean was revealed: The British Isles was connected to continental Europe, The Bering Strait was closed by a bridge of land between America and Siberia (Figure 1.2), and Indonesia was connected to Australia. The reduced temperatures at high latitudes during LGM also made the sea ice more extensive in the North Atlantic. The ice edge was situated further south and is believed to be found as far south as 45°N during winter (CLIMAP 1981; Pflaumann et al. 1996; Meland et al. 2005).

While the earliest reconstructions of sea surface temperature and sea ice cover showed a perennial ice cover in the Nordic Seas and northern North Atlantic as far south as 45°N (CLIMAP 1981), later studies generally agree on a less extensive ice cover (Pflaumann et al. 1996; Meland et al. 2005). Later studies generally agree on open water in the Nordic seas in summer, although there are differences between the reconstructions based on the core-material and reconstruction methods. A seasonally open ocean in the Nordic Seas might be an important factor to explain the existence of the large ice sheets over the Kara and Barents Sea. A climate with perennial frozen Nordic Seas associated with more zonal synoptic energy transport could not supply enough moisture to create these ice sheets at this high latitude when only surrounded with land or ice.

4.3 Arctic Climate

The Arctic atmosphere is characterized by the Arctic inversion. At plains such as the sea ice over the Arctic Ocean the temperature at the surface and in the surface layer of the atmosphere is very cold. -20-50°C is the typical range of surface air temperatures during winter. Cold air is heavier than warm air so when the winds are calm and the surface looses heat by long wave emittance, the surface and lower atmosphere is cooled and an inversion with increasing temperatures with height will form. Typically during winter the highest temperatures in the Arctic atmosphere are found at 800-1600 meters height. The stability of the inversion below reduces the turbulent heat exchange, and further calms the winds and increases the stability. A temperature difference across the inversion of 20°C is not unusual. During summer solar radiation heats the ground. This typically breaks down the inversion from below and convective processes may occur. This allows for more effective mixing and vertical temperature exchange than during winter. The surface temperature will usually be around the freezing point during the summer.

The Arctic is unique in the sense that it receives solar radiation 24 hours each day during large parts of the summer season. The North Pole receives 6 months of continuous solar radiation from March 21 to September 22. This is followed by 6 months of the polar night, i.e. 6 months with the sun below the horizon. Further south this variation is less extreme. 80°N receives no sunlight in the period from around October 21 to around February 20. The lack of sunlight during the polar night makes the Arctic a net sink of energy. The Arctic radiates more energy out to space than it receives from solar radiation. To keep the temperature from decreasing to a radiative balance, this energy loss is compensated by meridional energy transport in the atmosphere and ocean.

From September until February the sea ice in the Arctic Ocean grows (Figure 1.1). The ice is more reflective of solar radiation than open ocean, which has an albedo of around 0.05 (Allison et al. 1993). The albedo is a measure of the reflectivity of a surface, and ranges from 0 for black bodies to 1 for surfaces which reflects all incoming solar light. Depending on the age and composition of the sea ice the albedo ranges from 0.25-0.64 for first year ice to 0.7 for second year ice (Perovich 1996). The highest albedo (0.7-0.9) values are found when the sea ice is covered by fresh snow. Figure 3.1 illustrates the different albedo between sea ice and open water. The energy input to the Arctic atmosphere during the winter season is by advection of heat from the south and by thermal conduction through the sea ice. The conduction of heat through the ice from the underlying Ocean is relatively small compared to that of open ocean and becomes less as the sea ice grows thicker. This makes the surface of the sea ice very cold in winter. The sea ice insulates the Arctic Ocean and reduces its heat loss. In the spring the Arctic again receives solar radiation and the extent of the sea ice diminishes. Although large amounts of the solar radiation are reflected from the bright sea ice surface, the absorbed radiation will contribute to melt the sea ice and snow cover. When melting starts the radiative properties of the sea ice changes dramatically. Melt ponds are formed on top of the sea ice, these have a lower albedo and will more efficiently absorb solar radiation that contribute to further melting the ice.

The sea ice is in constant movement. The upper layers in the Arctic Ocean are mainly wind driven and consists of an anticyclonic gyre, the Beaufort gyre. The sea ice follows this circulation. Young ice is generally formed in the eastern Arctic Ocean in the Kara and Laptev Seas, it ages and thickens by the years as it either follows the Beaufort Gyre where it completes a circuit in 7-10 years, or it collected in the Transpolar Drift Stream and transported across the Pole and towards the Fram Strait during 2-3 years. The oldest and thickest sea ice is found in the Canadian Archipelago and North of the Fram Strait. The thickness of the sea ice is on average around 2-3 meters. The constant movement of the sea ice leads to stress in the sea ice cover with the result of convergence at some locations and divergence at other locations. Thicker ice, up to 20 meters, is the result of convergence, or ridging, of sea ice. The areas of sea ice divergence are called polynyas. These are small scale phenomena which is important on in climatology on larger scales. When the Ocean is covered by sea ice, this forms a lid which effectively reduces the Atmosphere Ocean heat exchange. The only heat exchange possible is by conductivity through the sea ice. When a polynya opens up, it allows for a sudden and large release of energy from the ocean to the atmosphere. The atmosphere will typically be cold and dry. The temperature difference between the ocean surface and the atmosphere above may be in the range 10-40°C. This gives potential for a large and efficient heat release. Winds will play an important role to supply new cold and dry air, which will make the heat exchange even more effective.

Clouds in the Arctic have been observed on manned Russian drifting stations on a regular basis since 1937. Data from costal stations are gathered from the early 1950s and later.

Clouds are observed by estimating how many tenths that are covered by clouds. By analysing cloud climatology by using the monthly mean cloud cover one must be aware that the mean cloud cover in general is different from the modal value (Zavyalova 2000). Typically, the monthly mean of total cloud cover in the Arctic is within the range 40-90%. But less than 20% of the observations actually report partly cloudy weather (3-7 tenths cloud cover). Typically clear skies (less than 3 tenths) or overcast (more than 7 tenths) are reported. Mean total cloud cover in the Arctic should be interpreted as frequency of occurrence of cloudy weather (Zavyalova 2000).

Typically more clouds occur in the summer season than in the winter season. In the summer clouds occur in approximately 70-80 % of the time while 50-60% in the winter. The variation in the cloud cover seems to follow the annual cycle of sea ice, and is most pronounced in the low cloud cover field. This indicates that the cloud cover is dependent on a local moisture source and vertical mixing. The typical cloud type in the Arctic is stratus (Figure 4.1), which is a flat homogenous cloud. Multiple inversions are often formed in the Arctic when the inversion is broken down and rebuilt e.g. by diurnal variability in solar radiation during spring and autumn. Consequently, multiple layers of stratus clouds are commonly found in the Arctic.

Figure 4.1: *Typical stratus cloud over the Arctic Ocean. The picture is taken in the Laptev Sea during IARC summerschool onboard Russian Icebreaker Kapitan Draitsyn September 2005*

Over the Arctic, most of the year the clouds constitute a positive climate forcing, the presence of clouds has a warming effect on the surface. In the winter, when the there are no solar radiation in the Arctic, the clouds absorbs the long wave radiation from the ground and reemits parts of it back to the surface. The surface is thus warmer than if no clouds were present. This is a positive cloud radiative forcing. In the summer it is not clear whether the clouds represent a positive or negative forcing (e.g. Minnett 1999; Shupe and Intrieri 2004). It represents long wave positive forcing by the absorption and reemitting of long wave radiation. But it also represents negative short wave forcing by reflecting solar radiation. Whether the clouds in summer represent a net positive or negative forcing depends on the height of the clouds and its optical thickness.

Clear evidence of a **changing climate** in the Arctic has been reported. The Sea Ice extent and thickness is decreasing (Johannessen et al. 1999; Parkinson et al. 1999; Serreze et al. 2003). The extent of the Arctic sea ice has been reported since 1978, when the first satellites started monitor the Arctic. Arctic sea ice extent has been found to decrease by 3% per decade in the period from 1978 to 1996. It is mainly the summer extent of sea ice that is diminishing. The ice covered area during summer has been reduced by 20% over the last 25 years. This decline in sea ice area amounts to approximately 1.3 million km^2. A new record minimum in sea ice extent was recorded on September 21 2005 according to a NSIDC (National Snow and Ice Data Center) press release. The freezing of new sea ice was less during the previous year and also the melting season started earlier than previous years according to their findings.

Evidence of a warming Arctic is also found over land. The Arctic tundra is becoming greener. Typical low tundra vegetation is replaced by more shrubs (Sturm et al. 2001). The North American and European tree-line is advancing northwards and reaching higher elevations (Kullman 2001; Lloyd and Fastie 2002). The growing seasons are getting longer and the frost season is getting shorter (Walther et al. 2002).

The Arctic tundra is frozen and permafrost is found in the soil, in some areas down to depths of more than 1000 meters (e.g. northern Lena and Yana river basins). Permafrost exists also under the Arctic Ocean seafloor. Large amounts of water are bound up in the permafrost. The latest years, the number of locations where melting permafrost has been discovered has increased. Romanovsky et al. (2002) have found increasing temperatures in the permafrost in the high northern latitudes in Alaska, Canada and Siberia over the last decade. The depth of the active layer (the layer which experience annual freezing and thawing) has become deeper. Melting will typically start immediately under the surface. The water from the melting ice finds its ways away from the melting area. After some melting the ground will collapse, and the ice wedges will be displayed (Figure 1.3). The Arctic permafrost contains frozen methane (hydrates). The frozen soil and ice wedges contains up to 0.6 volume percent methane (Brouchkov and Fukuda 2002). Warming and melting of the permafrost releases methane to the atmosphere, a greenhouse gas that will contribute to warm the planet even more. This represents a positive feedback mechanism.

4.4 Storm tracks

The term storm tracks can be defined as preferred regions of storm (cyclone) activity. While cyclones or storms can be regarded as disturbances of the mean flow on the synoptic scale (1000km).

Wintertime storm tracks in the North Atlantic are a key climate indicator for the bordering land masses. Figure 4.2 shows the density of the North Atlantic winter storm track for the period 1948-2002, as derived from NCEP reanalysis (Kalnay et al. 1996) by using the storm track algorithm by Hodges (1994; 1995; 1996; 1999). The algorithm will be described in Section 5.2. The probability density function (PDF) gives the number of storms expected to be found within a given area each month during winter. As the mean flow in the atmosphere is eastward, this corresponds to the cyclones being generated in the western parts of the storm track, or upstream of the storm track maxima.

In the northern hemisphere two storm track maxima can easily be identified. The Pacific storm track reaches across the north Pacific. The storms move eastward from Asia, where they are generated near the coast of Japan, across the North Pacific, toward the Aleutians and the Gulf of Alaska. The Aleutians are within the maximum of the north Pacific storm track, and the long term persistence of low pressure systems near the Aleutian Islands is known as the Aleutian Low.

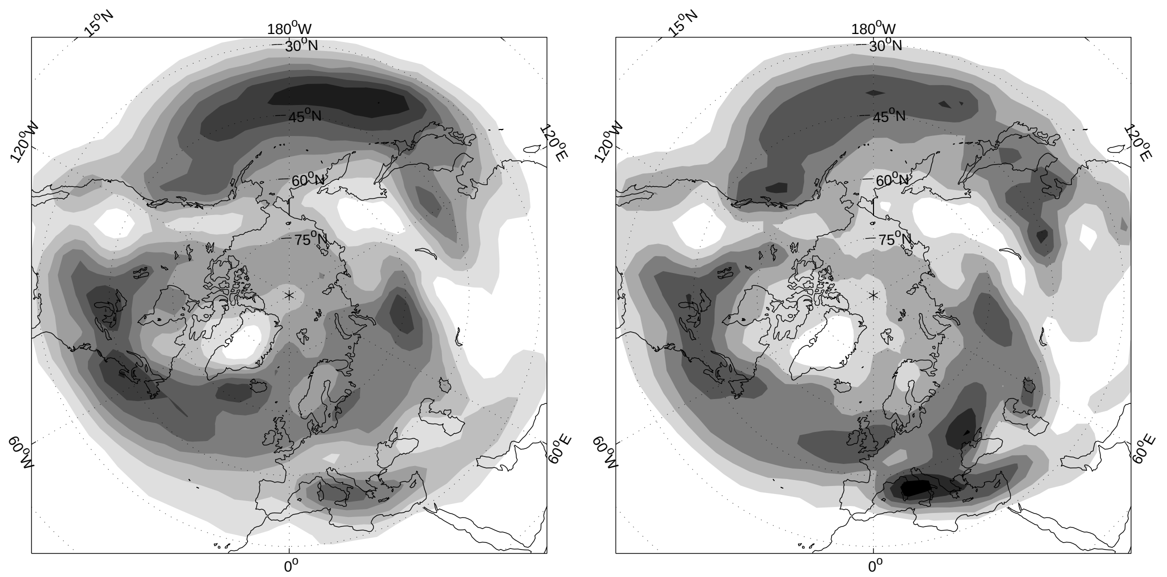

Figure 4.2: *Density of winter storm tracks based on relative vorticity at 850hPa. Left figure is storm track density for DJF from NCEP/NCAR reanalysis for the period December 1948-February 2002. Right figure is storm track density for a 20 year model simulation with the ARPEGE model. Dark areas correspond to a high density of storms. Note the figures are not quantitatively comparable due to different output frequency for the source data.*

The second major storm track maximum in the northern hemisphere is the North Atlantic storm track. The east coast of North America is a major area for creating synoptic lows. This is an area of high baroclinicity, where dry and often cold continental air meets the relatively warm and moist marine air. The coastline from Cape Hatteras to Newfoundland is known as a major cyclogenesis region. From this region the storms move northeastward toward Iceland. The persisting long term low around Iceland is known as the **Icelandic Low**. This is also a major location for cyclogenesis in the North Atlantic. The North Atlantic storm track is relatively broad. A large number of storms move from Iceland into the Nordic Seas (Paper I). These storms either move further toward the Arctic or toward Scandinavia and in particular the western coast of Norway. The centre of the Icelandic Low is located further north than the Pacific counterpart, the Aleutian Low. This is consistent with the Nordic Seas and Barents Sea being the main gate for import of heat and moisture to the Arctic atmosphere. Whereas a large number of the North Atlantic cyclones move into the Arctic Ocean, only few cyclones enter the basin from the Pacific side.

The storms are carriers of energy, and they are part of the meridional heat pump. They carry heat and moisture northward across the Nordic Seas and into the Arctic. A change in the storm count or their intensity may have implications on local climate and hydrology. A change in cyclone intensity, cyclone counts or geographical shifts in the storm tracks may therefore have large implications for people living in the inflicted areas. This can lead to substantial changes in precipitation. Farming is very sensitive to such changes, and especially areas where people are living from subsistence farming are heavily stressed by such changes. An increase in intensity and number of storms may cause floods and strong winds that ruin the land or crops. A decline in number of storms or their intensity may cause droughts. Examples on such changes are the El Niño events, when the Pacific storm track extends much farther downstream than it does during 'normal' winters.

The latitudes south of the North Atlantic and North Pacific storm tracks are dominated by persisting high pressure. This is the subsidence part of the Hadley circulation. These latitudes (20-30°) are characterized by warm and dry climate. The persisting high in the North Atlantic is known as the Azores High.

The position of the North Atlantic storm track varies quite much interannually. While it one year may have the tendency of choosing a northern path, it may choose a southerly path the next year. The tendency of the storms to choose either to go north toward Northern Europe, Scandinavia and into the Nordic Seas or to go across the European continent and to Southern Europe and the Mediterranean and the variability between those two phases is known as the **North Atlantic Oscillation** (NAO). Jones et al. (1997) showed the described link between the North Atlantic storm track and the North Atlantic Oscillation and found also that the position of the North Atlantic storm track seems to follow a decadal cyclicity. Hurrell (1995) reported that the NAO showed an increasing trend from the late 1970s to the mid 1990s.

The effect of the NAO was first noted over the years 1770-1778 by Hans Egede Saaby, a Danish priest who travelled to Greenland. He noted that there was a tendency of mild winters to occur in Denmark during the same years as the winters in Greenland was especially harsh. He also noted that when the winters in Demark were cold, there was a tendency of mild winters in Greenland. By this he suggested an anti-correlation in the climate between the two locations. This has been widely studied from the early 20th century. The first work that uses the term the North Atlantic Oscillation is by Walker and Bliss (1932). They noted that during winters when the Icelandic Low is stronger than normal, there was a tendency of the Azores High also to be stronger than normal. Walker and Bliss (1932) constructed an index of the

NAO which was based on linear regression of pressure, temperature and precipitation at 9 meteorological stations in Europe.

Later the NAO index has generally been calculated in two different ways. It is either calculated as a standardized pressure difference between a station in Iceland and a station in Portugal or the Azores (Hurrell 1995; Jones et al. 1997). The NAO can also be calculated as principal components of the North Atlantic pressure field by using singular value decomposition methods (SVD). The NAO pattern is the first principal component of the SVD analysis. It typical explains ~60% of the wintertime variance in the North Atlantic. By extending the SVD method to include the whole northern hemisphere north of 10°N Thompson et al. (2000) defined the Arctic Oscillation (AO). The AO pattern generally describes ~35% of the variance in the wintertime mean sea level pressure (MSLP). This suggests covariability between the North Atlantic storm track and the Pacific storm track.

When the NAO index is positive this corresponds to a larger pressure gradient between Iceland and the Azores, thus also a strong Icelandic Low. In positive NAO winters the storm trajectories are directed north eastwards. This results in wet and mild winters in Scandinavia and northern Europe. In these years southern Europe and the Mediterranean experience cold and dry winters. The seesaw pattern between winter climate in Denmark and Greenland is also explained by the NAO variability. When the NAO is positive and the Scandinavian winter is mild and wet, western Greenland tend to experience cold and dry climate as the cyclonic circulation of the stronger than normal Icelandic Low brings larger amounts of cold air from the Canadian Archipelago southward to Baffin Bay and the Labrador Ocean. The south eastern coast of North America tends to experience warmer winters these years. In negative NAO winters the Icelandic low and Azores high are both weaker than normal. The storms are directed toward continental and southern Europe. In these winters Scandinavia experience cold and dry weather (Hurrell 1995; 1996). Figure 4.3 shows the difference in pressure and temperature between positive NAO years and negative NAO years for the period 1948-2002.

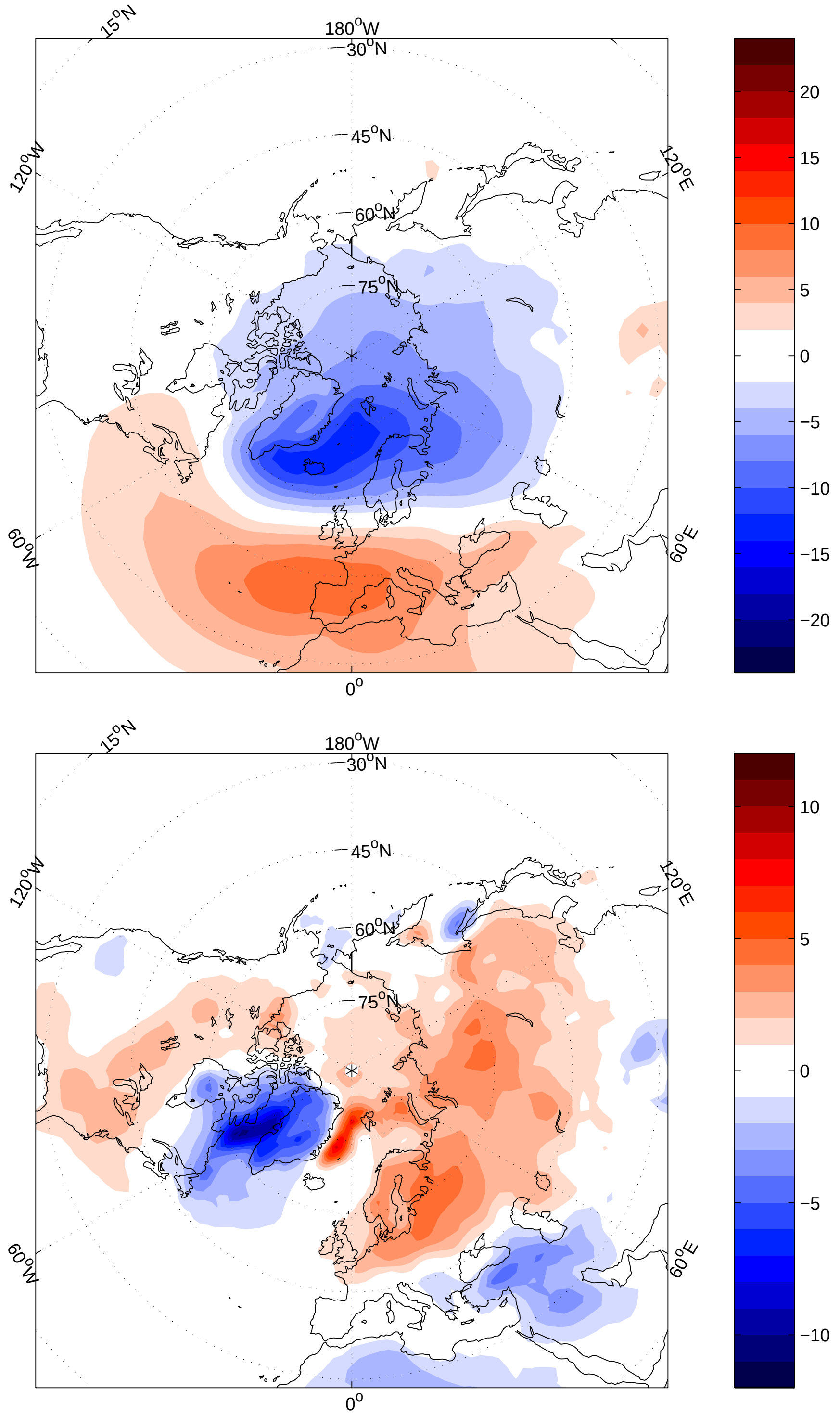

Figure 4.3: *Composites of DJF sea level pressure anomalies (upper figure) and DJF surface air temperature (lower figure) with NAO index greater than 1 STD or less than -1 STD. The unit is hPa and °C.*

The topography and land-ocean distribution on the northern hemisphere (Figure 2.1 and 2.2) plays a fundamental role in describing the storm track patterns. In particular the land-ocean contrasts are clearly important in describing the areas of high baroclinicity and thus the regions of cyclogenesis and energy transfers. The topography also plays an important role as cyclones commonly are generated on the lee side of large topographical features such as the Rocky Mountains and Greenland (e.g. Kristjánsson and McInnes 1999). Contrary to the northern hemisphere storm tracks, the storm track at similar latitudes on the southern hemisphere is primarily zonal. The Southern Ocean is a circumpolar ocean, with no dominating topographic features or land-ocean contrasts. The general circulation models (GCMs) generally have difficulties in simulating the storm tracks correctly on the northern hemisphere. The modelled northern hemisphere storm tracks typically show a too large zonal component. The models are not able to properly capture the northwestward component of the storm tracks in the North Atlantic. This suggests that the role of the topographical features (including gravity wave drag) on the northern hemisphere is not represented correctly in the models. Figure 4.2 compares the storm tracks from NCEP reanalysis with the corresponding from the ARPEGE model.

Too zonal representation of the storm tracks in the GCMs is indicative of the climate in the models being typical to that of negative NAO years. Although the position of the Icelandic Low is correctly located, and the strength of the Iceland-Azores pressure gradient is generally stronger in the model than in the observation, the trajectories of the storms are still too zonal, typical of a negative NAO index winters. Even though the models systematically predict too few storms into the Nordic Seas, they have a variability pattern in the North Atlantic that resembles the observed NAO.

The NAO can be an important factor regarding seasonal forecasting. If the NAO could be predicted, this could be used as a proxy for European winter climate. Bjerknes (1962) found a link between sea surface temperatures (SSTs) and the strength of the westerlies in the North Atlantic. This leads to a discussion whether the NAO forces the SSTs or the opposite. Kvamstø et al. (2004) performed a model experiment which focused on the influence of the sea ice on the North Atlantic storm track. They found that an extensive sea ice cover in the Labrador Sea would force the storm track further south than when the ice cover was reduced. However, the change in wind patterns when the NAO is changed may well force the ice and thus create more extensive or less extensive ice cover. Kvingedal and Sorteberg (2006) found that they could predict the sea ice extent in the Barents Sea by analysing the storm tracks of the previous years. They found a correlation of 0.97 between the decadal sea ice extent, with a

1 year lag, and decadal storm track density. This indicates the important role of the synoptic storms on the distribution of sea ice in the Barents Sea region.

Chapter 5

TOOLS

5.1 ARPEGE

The ARPEGE/IFS (Action de Recherche Petit Echelle Grande Echelle / Integrated Forecast System) model was developed as a forecast model in a joint effort between Meteo-France and European Centre for Medium range Weather Forecasts (ECMWF). Later the model was also expanded and adjusted to the needs for a community climate model for research purposes. The model is described in more detail in Deque et al. (1994).

The ARPEGE model is a spectral model. The horizontal resolution for the dynamical fields is given on a triangular grid. In the studies presented here a spectral resolution T63 has been used. The dynamical fields are expressed as Fourier waves with a cutoff at wavenumber 63. The triangular truncation represents a uniform resolution on the sphere. The surface resolution that corresponds to this resolution is 64 by 128 grid points. As the distance between the meridians decrease toward the poles, the grid has been reduced, so that the internal spacing between neighbouring grid cells are approximately the same at all latitudes. This reduces the computational costs and gives in total 6232 grid cells at the surface. This corresponds to a 2.8° resolution along the equator or 320km in latitudinal and longitudinal direction.

The vertical resolution is given as a hybrid vertical coordinate. The model is divided into a fixed number of layers. In the default configuration of ARPEGE the number of layers is 31. In the troposphere the model levels follow the topography, while they become parallel to pressure surfaces higher up. The spacing between the levels is most dense close to the ground with a vertical spacing of a few 100 meters in the default setup near the ground. This spacing increases exponentially with height. The levels are expressed as follows:

$$p(j) = \alpha(j) \cdot p_s + \beta(j) \tag{5.1}$$

Here p(j) is the pressure at level j at any given grid point. α and β are the coefficients that defines the vertical levels. p_s is the surface pressure at a given grid cell.

The ARPEGE model is an atmospheric model. The model requires a set of boundary conditions to run. Boundary conditions include sea surface temperatures (SSTs), sea ice cover, vegetation, land types, topography and land/sea distributions. These parameters will impact the simulated climate. The distribution of the sea surface temperatures are one of the drivers of the climate variability. Observed SSTs are thus often used as boundary conditions. In the real world the atmosphere feeds back on the sea surface temperatures. In the atmosphere-only models such feedback process is not possible.

Land surfaces often respond quickly to atmospheric forcing. The diurnal solar cycle drives the temperature of the land surface through the day. These processes are parameterized in the ARPEGE model by the ISBA (Interactions between Soil Biosphere and Atmosphere) scheme described by Douville et al. (2000). The land consists of several thermodynamical layers and responds thermodynamically to the atmospheric forcing, and also acts as a reservoir for moisture. The sea ice in ARPEGE is also treated by the ISBA sheme much in the same manner as the land surface. The sea ice thus responds thermodynamically to the atmosphere. However the extent of sea ice is a prescribed boundary condition in the model, i.e. the sea ice can not be melted away in the ARPEGE model.

5.2 The Storm Track Algorithm

The feature track algorithm developed by Hodges (1994; 1995; 1996; 1999) is used for the study of storm tracks in this thesis. As opposed to the often used band pass filtering methods to analyse the areas of high synoptic scale variability, this method localize each cyclone and anticyclone in every time step and link them together to form trajectories.

The data used to construct storm trajectories is relative vorticity at 850hPa calculated from 6 hourly (or 12 hourly) wind components. This field has been chosen rather than sea level pressure since the local maxima and minima are more easily identified in the vorticity field. The relative vorticity field can be very noisy with frontal systems nearly resolved. The processing is therefore done at a spectral resolution of T42. The reduction in resolution associated with the T42 truncation provides some smoothing of the higher frequencies. In addition planetary scales for total wave numbers $n \leq 7$ are removed from the field using a filtering method discussed by Anderson et al. (2003). This spectral cut off does not make any significant change to the results since the planetary scales are relatively weak in the vorticity fields relative to the anomalies. Local extremes are identified in the related fields, and these are interpreted as the centres for negative and positive vorticity anomalies.

The analyses are concentrated on the positive vorticity anomalies (cyclones), the feature points associated with low-pressure systems. The feature points for two consecutive time steps are linked together using a nearest neighbour search. A cost-function based on track smoothness in terms of changes in direction and speed (Hodges 1995; 1996) is calculated for each possible ensemble of trajectories. The ensemble minimizing the cost-function is chosen to be the best fit and thereby restricts the feature to physical movement. Semi-stationary storms (less than 10º total displacement) and short-lived storms (lifetime less than 2 days) are removed from the analysis.

Intensity is calculated each timestep for each cyclone. The intensity is calculated as an integral over depth and aerial extent of the cyclone. The identification of each single storm and its trajectory, and the storm intensity at each time step allows for close investigation of each cyclone's lifecycle.

The trajectory data is used to construct probability density functions (PDFs). This is the probability of a storm passing trough a given area. When this is multiplied by the total number of storm trajectories, the total number of storms that passes through a given region for each season is found. From the ensemble of trajectories several storm statistics can be calculated. Track density is the number of storms passing trough a region each season (or each month). Feature density is the number of feature points in a region per season. A feature point is the localized maxima or minima from the vorticity field. This is different from the track density field since a feature point may be counted more than once in the same area if the storm is moving slowly. Genesis density is the density of cyclogenesis. That is the location where the storm first occurs. Lysis density is the density of cyclolysis. This corresponds to the end position of the trajectory. Other variables that can be analysed from the data set is the cyclone speed, lifetime of cyclones and cyclone intensity.

The data used to construct storm tracks in this thesis is from NCEP/NCAR reanalysis (Kalnay et al. 1996) for the period 1948 to 2002. Also used are model data from model simulations with the ARPEGE model. For the reanalysis data 6 hourly data has been available. For the model data, the analysis is based on either 6 hourly or 12 hourly fields.

Chapter 6

SUMMARY OF SCIENTIFIC RESULTS

The main results from the research included in this thesis are presented in Part II. In this section the summaries of the papers are presented.

Paper I: Wintertime Nordic Seas Cyclone Variability and its Impact on Oceanic Volume Transports Into the Nordic Seas

In Paper I the variability in the cyclones in the Nordic Seas is investigated along with its impact on ocean volume transports. Data from the NCEP reanalysis (Kalnay et al. 1996) has been used as a basis for constructing a dataset of northern hemisphere storm trajectories for the period 1948 to 2002. Statistics of the storm tracks has been calculated to derive a gridded dataset of storm counts each season for each year.

It is shown that the number of storms and the average intensity of storms in the Nordic Seas are correlated with NAO and AO, and thus inhibit a strong interannual and decadal variability. The number of storms in the Nordic Seas region is found to have an interannual variability of 2.5 years and a decadal variability of 7-12 years. The correlation between the number of storms in the Nordic Seas and the Arctic Oscillation seems in general to be imcreasing during the last decades since the 1970s. The correlation between consecutive 10 year periods of the Arctic Oscillation and the number of storms in the region increases from approximately 0.5 in the first half of the period to the latter half to a to a value of about 0.7-0.8. It is noted a total breakdown of a relationship between cyclone activity in the Nordic Seas and the NAO-index in the period 1960-1970. From this it is evident that the station based NAO index can not be interpreted as proxy for cyclone activity in the Nordic Seas. An increase in the variability in the Nordic Seas coincides with an eastward shift in the NAO centre of activity in this period.

The cyclone statistics is also analysed in terms of their impact on volume transports into the Nordic Seas. By comparing the storm track analysis to results from the MICOM (Bleck et al. 1992) model for the period 1949-1999 the cyclone activity shows a positive and significant correlation with inflow through the Faroe-Shetland Channel and with outflow in the Denmark Strait. Negative correlation is found with inflow from the Demark Strait and outflow through the Faroe-Shetland Channel. This is explained by the difference wind stress on the surface waters in years with large cyclone activity compared to years with small cyclone activity.

Paper II: Sensitivity of Last Glacial Maximum Climate to sea-ice condition in the Nordic Seas

The aim of Paper II has been twofold. Firstly, to set up the model and run experiments simulating the last glacial maximum (LGM). The second aim was to do a sensitivity study of the LGM, and see how a modified sea ice extent in the Nordic Seas affects the LGM climate. New reconstructions of sea ice cover in the Nordic Seas have been published by Meland et al. (2005). This reconstruction shows seasonally open waters in the Nordic Seas in LGM.

The global temperature in the model in LGM was 5.7°C colder than for modern day. The temperature reduction is a response to reduced radiative forcing due to a smaller atmospheric content of CO_2 and aerosols, reduced SSTs in the boundary conditions and increase in albedo and elevation due to the large northern hemisphere ice sheets in LGM. The simulated LGM climate is wetter than modern climate due to a stronger Hadley circulation set up by larger fluxes of heat and moisture at the surface. Following the CLIMAP reconstruction the Nordic Seas is perennially covered with sea ice down to 45°N. It is found that the cyclones trajectories in LGM are strongly determined by the ice edge as this is an area of strong baroclinicity. The Icelandic Low, and the variability connected to the Icelandic Low, is inexistent in LGM. Cyclones thus follow the prescribed ice edge closely towards the Fennoscandian.

The sensitivity experiment with seasonally open water in the LGM shows increased temperature and precipitation locally. The surface pressure over the Nordic Seas is also reduced compared to the simulation with the full ice cover. Variability in the area increases, storms carries longer over the Fennoscandian ice sheet, and they are in less degree determined or steered by the ice edge. The simulation of the new reconstruction thus brings more heat and moisture toward the ice sheet and can thus help explain their existence (Boulton 1979; Elverhoi et al. 1995; Mangerud et al. 2002). Possible teleconnections tied to the new boundary conditions are also studied in terms of upper tropospheric wave patterns.

Paper III: Sensitivity of simulated wintertime Arctic atmosphere to vertical resolution in the ARPEGE/IFS model

Current GCMs typically have a positive bias in surface temperature in the Arctic (e.g. Tao et al. 1996; Beesley et al. 2000; Kiehl and Gent 2004). This is often attributed poorly represented vertical mixing in the Arctic stable boundary layer (SBL). In Paper III the focus is on how the model represents the Arctic boundary layer and exchange processes.

The underlying hypothesis is that the parameterization of the stable boundary layer employed in the model gives a relatively good representation of the boundary layer, but it suffers from bad vertical resolution in the model. To test this hypothesis two simulations are performed: a control integration with the default vertical resolution of 31 layers (31L), and a simulation where the vertical resolution in the lower part of the atmosphere has been greatly improved (in total 90 layers, 90L). The results show a rather poor representation of the Arctic inversion in the low resolution run, but this is clearly improved in 90L. The model data are validated against data from ERA-40 reanalysis (Uppala et al. 2005) and data from the SHEBA project (Uttal et al. 2002). In 90L strong inversions and low surface temperatures occurs more frequently than in the control simulation. The surface air temperature (SAT) bias over the Arctic Ocean is thus reduced.

Zilitinkevich and Esau (2003; 2005) demonstrated that properties of the Arctic PBL are different in weak wind and strong wind regimes as well as in weak and strong and atmospheric inversions. In all 4 regimes the coarse resolution run shows considerable discrepancies in representing the vertical temperature profile. These discrepancies are clearly reduced in the high resolution run. The dependence of the SAT on surface winds, surface energy fluxes, inversion stability and boundary layer height is investigated. The coarse resolution run reveals considerable biases in these parameters, and in their physical relations to SAT. In the simulation with fine vertical resolution these biases is clearly reduced. The physical relation between SAT and surface winds also becomes more realistic.

Cuxart et al. (2006) reported that the heat and momentum fluxes in the models that employ first order closure schemes were greatly overestimated in the lower 500m of the atmosphere. The vertical fluxes of heat and momentum is clearly reduced in the high resolution run.

Chapter 7

DATA SETS

Data used in the thesis comes from different sources. The following gives a description of the data that has been used.

NCEP/NCAR realalysis

A detailed description of the NCEP/NCAR reanalysis is given in Kalnay et al. (1996) and Kistler et al. (2001). NCEP/NCAR is the National Centers for Environmental Prediction and the National Center for Atmospheric Research. The reanalysis data are generated by a GCM, but relaxed to existing available observation data. By this analysis method the observation data is incorporated into a gridded global field, generally available at a 2.5° resolution in longitude and latitude.

The confidence of the reanalysis data is generally dependent on two factors. The first factor is the coverage of observation data available. The confidence in the reanalysis fields is therefore largest where the observational coverage is best. While in areas of more sparse observations there might be some biases in the reanalysis fields (e.g. the Arctic Ocean). The number of available observations has also changed over the reanalysis period. This can also to some extent introduce biases in certain periods of the reanalysis. The different fields that are produced by the reanalysis are all grouped into tree different levels of confidence (Kistler et al. 2001). The first confidence level includes the fields where the incorporated observations have a strong influence on the reanalysis data. Such fields are wind components and temperature on pressure levels and surface pressure. The second confidence level includes the fields where the observed values have a large influence, but where the model also have a large degree of influence. Example of such a field is the humidity on pressure levels. The confidence of the fields of the two first levels in the reanalysis is relatively good. The third level includes fields such as precipitation, evaporation and river runoff. Observation data for these fields are not included into the reanalysis, they are thus heavily dependent on the model and the parameterizations employed in the model. The large scale patterns of these fields is found to compare well to observed values, while substantial differences are evident on regional scales (Janowiak et al. 1998). The reanalysis are available as 6-hourly data for the period from 1948 until present and is available from ftp.cdc.noaa.gov. Data from the NCEP/NCAR reanalysis is used in Paper I.

ERA-40

The ERA-40 is a dataset of model generated reanalysis for the period September 1957 to August 2002 by ECMWF (European Centre for Medium-Range Weather Forecasts). A detailed description can be found in Uppala et al (2005). The ERA-40 incorporates satellite observations, observations from aircrafts, ocean-buoys, surface observations and radiosonde observations. The quality and accuracy of the ERA-40 data is in many respects found to be an improvement compared to earlier reanalysis by NCEP/NCAR and ERA-15. Improvements are in particular found in the simulation of the hydrological budgets (Bromwich et al. 2002; Hagemann et al. 2005). The quality of the reanalysis is to some degree dependent of the amount of available observation data. Inclusion of data from the satellite borne instruments since the 1970s improves the quality of the reanalysis in the latter part of the period.

The observation data is incorporated into the ECMWF model which is run with a T159 horizontal resolution with 60 layers vertically. This is also an improvement compared to earlier reanalysis products. Data from the ERA-40 reanalysis is used in Paper III

Environmental Working Group Climate Atlas of the Arctic

The cloud cover data from the NCEP/NCAR reanalysis is generally not reliable and are highly influenced by the model parameterization (Walsh and Chapman 1998) especially in the Arctic. To analyse cloud cover in the Arctic the compilation of observation data by the Environmental Working Group (Arctic Climatology Project, 2000) has been used. The Environmental Working Group is a joint effort between the National Snow and Ice Data Centre (NSIDC) in Boulder, Colorado, and the Arctic and Antarctic Research Institute (AARI) in St. Petersburg, Russia. The data is a compilation of station observations, observations from ice patrol ships and drifting stations in the Arctic Ocean from both the American and Russian side of the Arctic. The Environmental Working Group has compiled monthly means of the available observation data, this includes data at specific locations and also data interpolated to gridded fields. The gridded fields supplied are for decadal averages arranged by month of the year. The gridded data available from this compilation is surface air temperature at 2m height, sea level pressure, total and low cloud cover, and solar radiation. Gridded fields are available for the period 1950-1990. Data at different locations (meteorological stations, drifting stations, ice patrol ships) are available generally for the same period, but some earlier data is also available. The earliest data from floating stations included are those from the Fram expedition in 1893-1897. The earliest station data are from Upernavik in western Greenland which dates back to 1873. The data is generally sparse before 1950. The data supplied in the Atlas have been passed through a series of quality control routines. Details are given by Arctic Climatology Project (2000). Cloud data from the Arctic Climatology Atlas is used in Paper III.

Surface Heat Budget of the Arctic Ocean – SHEBA

With the objective to study the heat budget of the Arctic Ocean, the SHEBA project was launched in 1997. The project is described in more detail in Uttal et al. (2002). The focus of the project was to investigate and make measurements of vertical exchange in the atmosphere and ocean boundary layers. Measurements of atmosphere, ocean, sea ice and snow cover were taken to study how these elements interact. The SHEBA station was an icebreaker frozen in the perennial sea ice for the period from mid October 1997 to mid October 1998. In this period the ship drifted 1400 km with the ice. Measurements of the vertical atmospheric structure were taken from measurement towers. Radiosondes measuring temperature and moisture profiles were also released 2-4 times daily. The radiosonde data from SHEBA is used in Paper III.

Chapter 8

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