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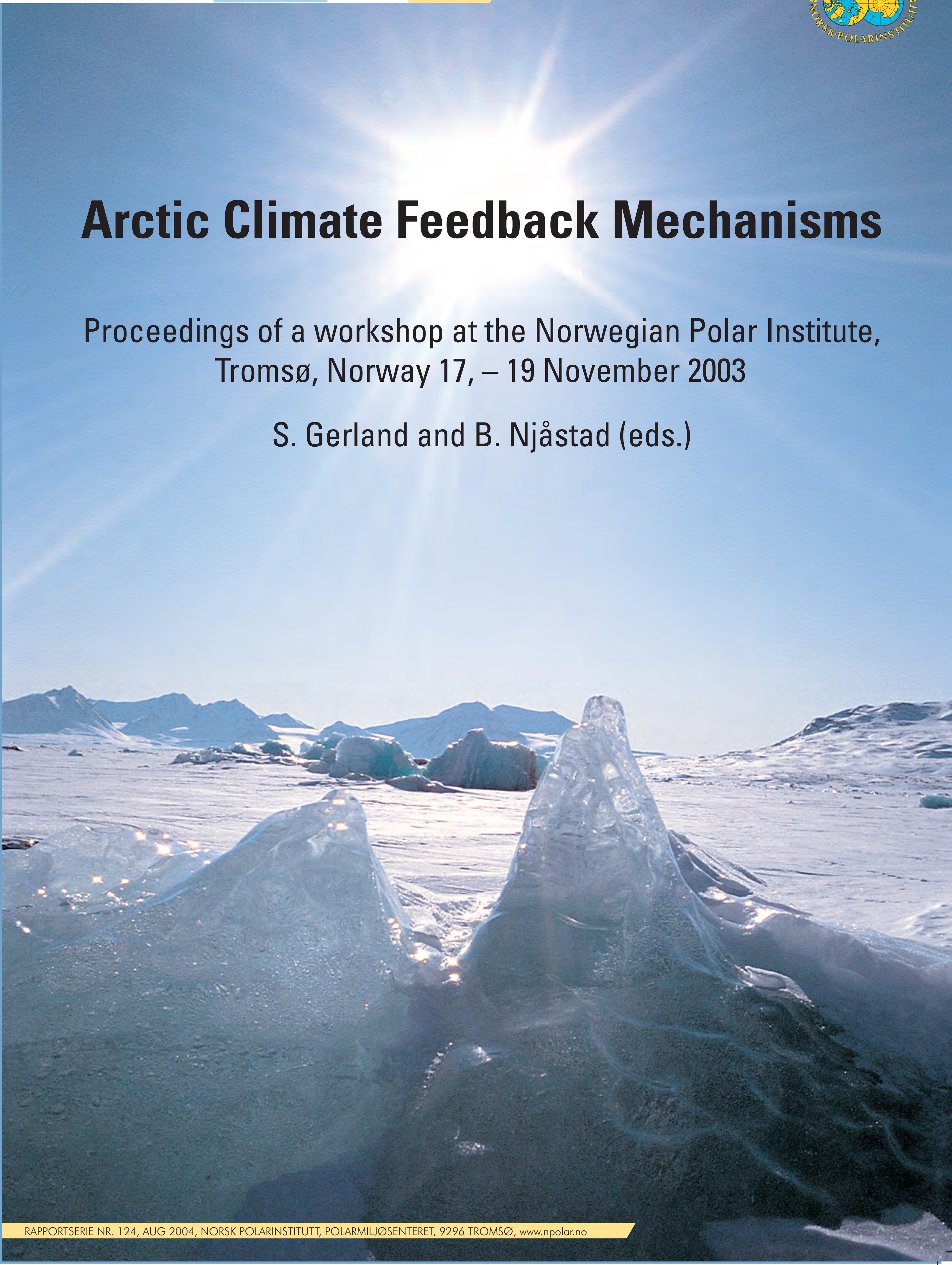

Arctic Climate Feedback Mechanisms

Proceedings of a workshop at the Norwegian Polar Institute, Tromsø, Norway 17, – 19 November 2003

S. Gerland and B. Njåstad (eds.)

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Preface

In November 2003 the workshop "Arctic Climate Feedback Mechanisms" was organized by the Norwegian Polar Institute as a contribution to the Arctic Council's Arctic Climate Impact Assessment (ACIA) process. A total of 60 participants, representing researchers from eight countries, as well as representatives from Norwegian authorities, took part in the workshop. The workshop was held in Tromsø, Norway.

One of the aims of the ACIA-process is to consider issues related to knowledge gaps and uncertainties that need to be taken into account in future research and monitoring work. The workshop "Arctic Climate Feedback Mechanisms" aimed to contribute to the process of identifying and illuminating the challenges and issues one faces in the further work of assessing the consequences of climate change in the Arctic.

This report presents scientific information and identifies knowledge gaps discussed at the workshop. Recommendations for future research and monitoring are also included. The report is based upon presentations given at the workshop as well as summaries from the discussions on the following topics:

- Feedback mechanisms in terrestrial systems;
- Feedback mechanisms in ocean systems;
- Feedback mechanisms in sea-ice systems; and
- Feedback mechanisms in atmospheric systems.

The Organizing Committee would like to thank the invited speakers for their help and expertise. The Organizing Committee would furthermore like to thank the Ministry of the Environment for supporting the ACIA-process and the Steering Committee for ACIA-Norway[1] for financing the workshop.

The Organizing Committee gives special thanks to Anne Kibsgaard, Ellen Berg and Ingrid Storhaug. Their help was vital to the success of the workshop.

Sebastian Gerland and Birgit Njåstad

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[1] ACIA-Norway is used to connote the Norwegian national process established to support the international ACIA-process. The national process is lead by the Ministry of the Environment. The Norwegian Polar Institute holds the secretariat for ACIA-Norway.

Contents

Background

The Arctic Climate Impact Assessment process

The Arctic Climate Impact Assessment (ACIA) is an international process implemented by the Arctic Council.

The goal of the ACIA-process is to:

- evaluate and synthesize knowledge on climate variability, climate change, and increased UV radiation and their consequences; and
- provide useful and reliable information to the governments, organizations and peoples of the Arctic region in order to support policy-making processes and to the Intergovernmental Panel on Climate Change's (IPCC) further work on climate change issues.

Through the ACIA-process it is the aim to collect and collate information about climate variability and climate change in the Arctic, consider what climate situation we may expect in the future and assess potential future consequences, both to the physical, biological and societal structures.

The ACIA-assessment does not itself implement new research, but is primarily based on knowledge gained through national and regional research projects and programmes. Through the ACIA-process it is nevertheless desirable to identify areas where further research and monitoring is needed.

The results of the ACIA-process will be presented through three reports: i) a scientific document, ii) a synthesis document and iii) a policy document. The first phase of the ACIA-process will be concluded in 2004.

ACIA-Norway

The international ACIA process is an important initiative seen in a Norwegian perspective, and Norwegian authorities consider it important to be an active partner and contributor to the extensive international process. A national process lead by the Ministry of the Environment has been initiated, with the following objectives:

- support the work of the Norwegian authors in the international ACIA-process;
- involve affected parties (scientists, decision-makers, industry representatives, etc.) in the ACIA-process, and in this manner increase the focus on the potential consequences of climate change in northern Norway and Svalbard; and
- contribute to the collection of relevant data that can be used in the ACIA-process and through this strengthen the ACIA-process.

Workshop on Arctic Climate Feedback Mechanisms

One of the aims of the ACIA-process is to consider issues related to knowledge gaps and uncertainties that need to be taken into account in future research and monitoring work. Clearly uncertainties relative to the consequences of climate change lie to some extent in the uncertainties related to feedback mechanisms.

As a part of the work of ACIA-Norway a workshop the "Arctic Climate Feedback Mechanisms" was held at the Norwegian Polar Institute in Tromsø, Norway, 17 -19 November 2003. The aim of the workshop was to contribute to the process of identifying and illuminating the challenges and issues one faces in the further work of assessing the consequences of climate change in the Arctic.

About 60 persons from eight countries attended the workshop (a list of participants is included in the report). The workshop consisted of invited talks divided into five sessions (climate programmes, terrestrial systems, oceans, sea-ice, and atmosphere), as well as poster presentations. Working groups discussed issues related to the four scientific session fields, identifying the state of knowledge, challenges, and recommendations.

Extended abstracts from the invited talks, poster abstracts, as well as recommendations from the four working groups, are presented in this report. It is hoped that the findings of this workshop will provide useful insights for the future work connected to climate research in the Arctic.

Summary and Conclusions

ACIA and Climate Feedback Mechanisms

One of the aims of the ACIA process is to consider issues related to knowledge gaps and uncertainties that need to be considered in future research and monitoring work. This includes identifying gaps in basic knowledge and identifying fundamental data that need to be acquired to better understand climate variability and change. Further, uncertainties regarding the consequences of climate change lie to some extent in the uncertainties of feedback mechanisms, e.g., particularly the depiction of the mechanisms in General Circulation Models (GCMs).

Terrestrial systems

The terrestrial system in the Arctic, including snow and the Arctic Ocean's ice-cover, as well as rivers, lakes, and vegetation, plays a crucial role in the global climate system through many complex interactions and feedback mechanisms. Despite numerous studies and our possession of broad knowledge about the subsystems, knowledge gaps exist within the complex interactions among the different systems. In hydrology, the broadest impacts of global warming on the non-glaciated terrestrial Arctic regions will result from changing permafrost structure and extent. The snow cover in the northern hemisphere also has a large influence on the Earth's albedo and on the global radiation balance. The snow further strongly interacts with the vegetation. Snow cover changes will also influence the release of greenhouse gases from soil. However, knowledge about feedbacks on larger spatial and temporal scales is lacking owing to insufficient monitoring systems and insufficient use of holistic approaches. For the ice-covered terrestrial areas in the Arctic, several strong positive feedback mechanisms characterize glacier–climate interactions.

Oceans

Significant progress has been made in the past 10 years both in oceanographic observations, e.g., measurement of transport through the Nordic Seas, and development of high-resolution oceanographic models. Yet, the Arctic Ocean is still very poorly known. In general, present-day GCMs cannot assess sufficiently the observed changes to the thermohaline circulation. Further, changes in stratification associated with temperature and salinity changes, atmospheric processes, and river runoff, are poorly understood. The position and dynamics of fronts and the ice edge of the marginal ice zone, influencing deep-water formation and overflows, need to be better determined. Further knowledge gaps exist around factors controlling greenhouse gas release and carbon sequestration. Future activities will address the dynamics of the Arctic Ocean circulation, considering the freshwater balance, and improved parameterizing Arctic

Ocean and Nordic Seas processes in the GCMs. This will allow model reconstructions of the present state in the Arctic, and improved future simulations. Complete fundamental surveys of the sea-ice and ocean, along with long-term monitoring activities, have to be maintained and expanded to document and understand Arctic climate variability.

Sea-ice

Sea-ice extent, thickness and distribution; snow depth; and energy budget and dynamics are key factors for understanding the role of the ice-cover in climate change. It is well established that the ice extent is significantly decreasing. While there is evidence that the ice thickness is also decreasing, the record is not as comprehensive as that for ice extent. It is imperative that in situ ice thickness monitoring and satellite remote sensing be continued and expanded. Observations indicate an increasing net energy budget of the ice. Limited information indicates a downward trend in albedo, but knowledge of large-scale changes in albedo is incomplete. Because of the ice–albedo feedback, changes in albedo are of major importance. Albedo information is also required for advanced parameterizations in climate modelling. Snow acts as an insulator, retarding ice growth in winter, and as a reflector, reducing ice melt in summer. However, snow-on-sea-ice information is sparse, and improvement of remote sensing techniques capable of monitoring snow properties is necessary. Future work needs to include monitoring studies, as well as efforts directed at understanding the key physical processes. Interdisciplinary approaches will be of particular value.

Atmosphere

Internal atmosphere evaporation processes and extent, knowledge of humidity, temperature and condensation, and the process of polar heat transport in the Arctic are relatively well understood. A high level of understanding prevails regarding the role of the surface albedo in atmospheric processes. Knowledge gaps were identified within feedback processes relating especially to clouds and aerosols in the atmosphere, the interaction between troposphere and stratosphere, eddy transport processes, mechanisms controlling the development of atmospheric frontal positions, and issues regarding the Arctic boundary layer. The investigation of so-called "extreme events" was identified as another important challenge, as even the concept itself may not be well defined. Extreme events may be defined as complex risk related events, threshold-defined/irreversible events, simple statistical outliers, or events connected with the human perception of extreme or changing processes. Many events could fall into several of these categories. Extreme events occur with high interannual variability or seldom, which makes their investigation difficult. However, they also require further attention, as they potentially can result in severe socio-economical and ecological problems.

Overall conclusions

Generally, enhanced monitoring, process studies and modelling were identified as high priority future work. Better spatial and temporal resolution, as well as use of modern technology, are keys to improved investigations of the Arctic, and indispensable for the investigation of non-linear processes, extreme events and rapid changes. Remote sensing studies will need detailed ground-truthing work to be a powerful tool for accurate monitoring. However, also linking studies/integrated and interdisciplinary work, were identified as being necessary and important. Further research on the consequences of climate change in the Arctic will be necessary also after the submission and presentation of the planned ACIA reports in 2004. It is important to initiate a process aiming at identifying and illuminating the challenges and issues one faces in this further work.

Working Group Report: Terrestrial Systems

Chairs:
Elisabeth Isaksson (NPI) and Niels Reeh (ARTEK, BYG •DTU)

Participants:
Oddbjørn Bruland (SINTEF), Jon Ove Hagen (UiO), Larry Hinzman (UAF), Annika Hofgaard (NINA), Jack Kohler (NPI), Oddvar Skre (University of Oulu) and Gunter Weller (UAF).

Introduction

The terrestrial system, including snow and ice-cover, rivers, lakes and vegetation, is a crucial component of the global climate system through a wide range of complex interactions and feedback mechanisms. Some of these systems have been thoroughly studied, and many of the involved processes are well known. Due to the complexity of the systems, however, the interactions between the different systems are particularly difficult to assess. The first order effects of a warming climate in the Arctic regions are already becoming apparent; some second order or subsequent impacts to the ecological and hydrologic systems are also becoming evident. The pressing research need for the coming decade is to quantify the interactions and feedbacks among related processes.

Hydrology

The broadest impacts to the non-glaciated terrestrial Arctic regions will result from effects of changing permafrost structure and extent. As the climate differentially warms in summer and winter, the permafrost will become warmer and the active layer (the layer of soil above the permafrost that annually experiences freeze and thaw) will become thicker. These simple structural changes will affect every aspect of the surface water and energy balances. As the active layer thickens, there is greater storage capacity for soil moisture and greater lags and delays are introduced into the hydrologic response times to precipitation or snowmelt events. When the frozen ground is very close to the surface, the stream and river discharge peaks are higher and the base flow is lower. As the active layer becomes thicker, the moisture storage capacity becomes greater and the lag time of runoff increases. This has significant impacts on large and small scales. The timing of stream runoff will change, reducing the percentage of continental runoff released during the summer and increasing the proportion of winter runoff. This is already becoming evident in Siberian Rivers. As permafrost becomes thinner and is reduced in spatial extent, the proportions of groundwater in stream runoff will increase as the proportion of surface runoff decreases, increasing river alkalinity and electrical conductivity. This could impact mixing of fresh and saline waters, formation of the halocline and seawater chemistry. Additionally, changes in surface soil moisture will affect sensible and latent heat fluxes, impacting local and regional climate.

The snow cover is one of the key parameters that determine the development of the permafrost. A warmer climate will lead to an earlier snowmelt and a later start of the snow accumulation and thereby a longer active layer melt period. Increased winter precipitation will give better insulation of the soil and thereby influence the refreezing of the active layer. A more extreme climate with stronger winds can on the other hand increase the redistribution of the snow. This will increase the snow loss due to sublimation and give larger snow free patches exposed to the low winter temperature and in turn increased permafrost.

The snow cover on the northern hemisphere also largely influences the Earth's albedo. Changes to the Arctic and Sub-Arctic snow cover will therefore directly influence the radiation balance of the earth. In addition, the snow cover strongly interacts with the vegetation. Vegetation traps the snow and reduces redistribution and snow sublimation, leading to a thicker snow pack. This will, again, influence the vegetation. The interaction between snow and vegetation will also strongly influence the Earth's albedo as vegetation reaching above the snow cover strongly reduces the reflection of the bright snow.

Changes to the snow cover will also influence the release of greenhouse gases from the soil. As long as snow covers the soil, this release is limited. During the spring melt period gases stored in the snow are released in a pulse. After the snow has melted, and as the soil surface thaws, the gas exchange at the surface significantly changes.

River discharge is the integrated response of a river catchment. However, processes affecting discharge are so many and so complex that there is no good way to invert the river discharge back to elucidate the basic hydrologic processes. Although we need more gauging stations to understand global water balance, we will not learn much more about hydrologic processes by adding more river gauging stations to the existing network. A coordinated set of research basins, all monitoring snow distribution, snow melt, energy balance, soil moisture dynamics, runoff, evaporation, transpiration, and sublimation with complementary techniques is necessary to develop an understanding of changes in interactions among hydrological, biological and climatological processes. Changes in climate initiate a cascade of changes in interdependent processes. We need to characterize how river discharge levels and sediment yields will vary with permafrost degradation and increased glacier ablation. It is important to understand how these changes link terrestrial and marine processes by impacting estuarine processes and ultimately oceanic circulation.

State of knowledge

- Physical processes regarding permafrost, snow, river runoff and the first order effects of climate change on these processes are relatively well known; and
- There are good models and developments activities, which are ahead of the GCMs, i.e. they are not yet implemented in GCMs.

Challenges

- Increase understanding of 2nd order effects of a changed hydrological regime;
- Increase understanding of interactions with vegetation and atmosphere;
- Acquire more data on winter precipitation in the Arctic;
- Enable quantification of river runoff in Russia (due to lack of data on both discharge and climate); and
- Increase understanding of the consequences of shifts in river break-up timing involving nutrients and freshwater.

Recommendations

- Carry out multidisciplinary studies focusing on the 2nd order effects;
- Carry out snow measurements programmes for winter precipitation; and
- Establish research watersheds where snow pack distribution, ablation, evaporation, soil moisture, ecological dynamics, and climate stations are all monitored in strategic locations throughout the Arctic.

Vegetation
Direct and indirect climate-related changes in vegetation linked to snow distribution, wind, hydrology, permafrost and disturbances (e.g. fire and insects) are known in some detail. However, knowledge on feedbacks relevant to larger spatial and temporal scales is faulty due to fragmented and insufficient monitoring systems and lack of holistic approaches. Effects of climate-related disturbances and non-climate-related disturbances (for example human activities) have to be integrated to understand responses of the system. To better understand and predict dramatic and irreversible shifts in vegetation or ecosystem state there is a need for increased attention to the pronounced non-linearity in system responses to climate variability and change. This view needs to be included in all theoretical and empirical procedures concerning studies of climate change and its consequences for Arctic and Sub-Arctic systems. The complexity and uncertainties of the forest–tundra ecotone dynamics and their consequences for, and feedbacks to the climate system, biodiversity and regional socio-economy call for increased scientific attention to broaden our understanding and to facilitate for better predictions of e.g. movement of tree line and rate of movements. The tree line has generally been predicted to show rapid advance, but there have in fact been recent and regionally large-scale retractions, (due to e.g. paludification, climatic stress, and human causes). Accordingly, rate of movements have to be analysed in the light of episodic causes vs. adaptation/adjustments.

State of knowledge
- Direct and indirect climate-related changes linked to snow distribution, hydrology, and permafrost are known in some detail.

Challenges
- Increased understanding of effects of disturbances such as insects and fire;
- Increased understanding of shifts in vegetation/ecosystem state;
- Increased understanding of movement of tree line; and
- Increased understanding of the rate of movement – episodic causes more vs. adaptation/adjustments.

Recommendations
- Multidisciplinary regional to circumpolar studies focusing on interactions between vegetation, hydrology, permafrost, and human activities.

Glaciers, ice sheets, ice cores
Glaciers and ice sheets interact with climate on a range of time-scales ranging from annual, due to changes of snow cover (albedo-change), through decadal to century or even longer time-scale variations, due to fresh water exchange with the ocean, up to millennial, due to changes in landscape topography by down-wasting of ice masses and isostatic crustal movement. Several strong positive feedback mechanisms characterize glacier–climate interactions, e.g. the melt-rate–albedo feedback, the melt-rate–glacier sliding feed back and the mass-balance–surface-elevation feedback. In accordance with the worldwide trend, most Arctic glaciers and ice caps showed significant mass loss during the second half the 20th century. With regards to the Greenland ice sheet, recent investigations suggest that this ice-mass is to presently losing mass, although it is not yet clear whether this mass loss reflects a long-term trend or whether it simply reflects short-term temporal variability of snow accumulation and melt-rate. The present state of balance of the Antarctic ice sheet is still an open question.

While the world's total glacier and ice sheet area is pretty well known, this is not the case for glacier volume. In particular, the total volume of the small glaciers and ice caps is based on estimates and not on observed ice thickness. The processes related to the surface mass balance of glaciers are pretty well understood and climate-induced changes of surface mass balance can be modelled with reasonable accuracy. An exception lies in our ability to model processes in the percolation zone, where melting/re-freezing processes dominate. Also, our understanding of processes at the glacier bed (primarily sliding), glacier calving, and how these processes are influenced by climate change, is at best fragmentary if not completely lacking.

Ice cores as palaeo-archives have contributed to our knowledge about natural climate variability on time scales from glacial – interglacial to interannual – decadal scale. Ice cores from smaller Arctic ice caps have not been studied to the same extent as Greenland ice cores largely due to uncertainties about the effect of melt water percolation on these lower elevation glaciers. However, results from many recent ice cores studies have shown that with careful site selection, high-resolution sampling and multiple chemical analyses, it is possible to recover ice cores with preserved information about major trends in both climate parameters and pollution history. Therefore an improved spatial coverage of ice cores covering the last several hundred years have the potential to greatly improve our knowledge about the timing and extent of rapid climate change and thus feedback mechanisms. The recent increase of summer melting on these ice caps makes it urgent to recover ice cores before the melt water percolation has destroyed the stratigraphy.

State of knowledge
- Area and, to a lesser extent, volume are known; and
- General pattern of area changes (retreat) is known.

Challenges
- Acquire data on quantitative volume (mass) changes;
- Improve interpretation of observed surface elevation change in terms of mass change (densification, ice lens formation);
- Quantify calving fluxes;
- Understand calving processes;
- Understand glacier dynamic response to basal hydrology changes; and
- Retrieve intermediate deep ice cores (500 years time coverage) before ice caps melt away.

Recommendations
- Maintain glacier mass balance series;
- Continue field studies for validation of remote sensing data;
- Conduct in situ studies of snow–firn–ice transformation in percolation zones;
- Maintain and establish new automatic weather/mass balance stations on and near glaciers and ice sheets; and
- Ensure better spatial coverage of intermediate ice cores.

Overall recommendations on terrestrial systems
- Improve monitoring activities both the quality and spatially;
- Increase the use of remote sensing together with field-ground truthing;
- Use down scaled regional models;
- Initiate multi-disciplinary studies; and
- Improve knowledge about the Russian Arctic.

Activities around The International Polar Year (IPY) may be a good opportunity to fulfil some of these recommendations.

Working Group Report: Ocean Systems

Chairs:
Peter M. Haugan (UiB) and Vladimir Pavlov (NPI)

Participants:
Ken Drinkwater (IMR), Michael Karcher (AWI), Harald Loeng (IMR) and Ernst Maier-Reimer (Max-Planck-Institut für Meteorologie)

State of knowledge

Very significant progress has been made in the past 5-10 years both in observations, e.g. measurement of transport through the Nordic Seas, and development of high-resolution models capable of simulating some pathways and anomalies. Yet the Arctic Ocean is still very poorly known. Observations of the basic state and variability are missing, or are weaker than for most of the world oceans.

Status can usefully be assessed by discussing gaps of knowledge. Relevant knowledge gaps, mainly physical, were considered, using the Marine Systems part of the draft ACIA scientific report as basis for the discussions.

Gaps of knowledge

Present-day Global Circulation Models (GCMs) can not assess the changes to the Thermohaline Circulation (THC). This is partly because of uncertainties concerning the formation, distribution and redistribution of fresh water exported from the Arctic. There is therefore some urgency in improving understanding, measurements, and modelling of fresh water export.

- Changes in vertical stratification associated with changes in temperature and salinity is poorly understood. Forcing, including wind and runoff, influences vertical mixing. Changes in temperature and salinity are influenced by mixing and sometimes density compensation.

- The impact of climate change on ocean currents and transport pathways. Will there be gradual change or mode changes (sense of circulation, shelf vs. open ocean convection)? Are there switchgears on the shelves or in the deep Arctic? Precipitation changes in the large drainage area, enhanced permafrost melt, and river discharges force the large shelf areas where changing transformation processes may influence also deep ocean pathways.

- Fronts in the ocean and their position and strength need to be better understood. Water mass extension (SST) and ice edge location influence the atmosphere and therefore the climate. Front positions may also influence deep-water formation and overflows. Their dynamical oceanographic importance is determined by position in relation to topography, e.g. slope around the Arctic Ocean.

- Processes that control release of greenhouse gases and sequestration of carbon. Uptake related to sea-ice formation and carbon transports associated with ventilation.

Challenges

- Understand the dynamics of the present Arctic circulation, parameterization of the key processes in models with different resolution, and reproduction of the present state of the Arctic in models.

- Understand the different sensitivities of global climate to fresh water export from the Arctic (it seems to differ largely between models).

Recommendations

- Develop GCMs with higher resolution and more focus on ocean processes critical to high latitudes:

  - Use regional high-resolution ice–ocean models (limited area or focused global models) to address regional aspects of global climate change (dynamical downscaling).

  - Model intercomparison focusing on key process representation (e.g. water transformation processes).

- Initiate efforts to complete fundamental surveys of the Arctic's ice and ocean. Monitoring activities have to be maintained and expanded to document and understand climate variability:

  - Simultaneous transport measurements in all important gateways for climate variability.

  - Surveys in the central Arctic in the post submarine era, possibly with new innovative technology supplementing ice breakers.

  - Wintertime (all year) profiling instrumentation (under ice positioned floats and drifters, ice-based profilers) primarily for process understanding.

  - Use of tracers to document net effects of transport and mixing for testing of models.

  - Utilization of altimetry data for estimating temporal variability of sea surface height and circulation. Also other satellite data including SST and ocean colour, in addition to the sea-ice characterization.

  - Reconstruction of forcing fields on decadal to centennial time scales.

  - Inclusion of biogeochemical parameters (carbon and tracers) in particular on arrays for transport estimates in gateways and in conjunction with physical process studies.

- Increase efforts on marine process studies in order to achieve better understanding of climate change and variability and to help improving parameterization of numerical climate models.

- Increase efforts on studies on oceanfronts as areas sensitive to physical mixing processes.

- Increase efforts to estimate carbon reserves in Arctic marine sediments and to understand mechanisms of release.

- Increase efforts to study oceanic sequestration of carbon in the Arctic marine water column and increase understanding of air–ice–ocean exchange.

Working Group Report: Sea-ice Systems

Chairs:
Don Perovich (ERDC-CRREL) and Peter Wadhams (University of Cambridge)

Participants:
Sebastian Gerland (NPI), Jari Haapala (FIMR), Ingo Harms (University of Hamburg), Boris Ivanov (AARI), Christina A. Pedersen (NPI) and Gabrielle Østern (Norwegian Ministry of the Environment).

Overview

The group examined sea-ice from the perspective of climate change and feedbacks. Within this framework the discussion focused on three questions: what is known, what is not known, and what are the most important things to learn. Several important sea-ice parameters were identified, followed by a discussion of whether the parameters were changing, the certainty of any observed change, the importance of the parameter, and what future work is needed. Results from this discussion are summarized in Table 1.

Key factors for sea-ice in climate change

Sea-ice extent, thickness distribution, and energy budget were identified as key factors in understanding the role of the ice-cover in climate change. It is well established that the ice extent is decreasing and that this observed decrease is of great significance. While there is evidence that the ice thickness is also decreasing, the record is not as comprehensive as that for ice extent. It is imperative that monitoring of ice thickness using submarine surveys, ship observations, and upper looking sonars be continued and expanded. The ongoing development of techniques to remotely sense ice thickness offers the future promise of large-scale ice thickness monitoring.

Satellite observations along with a few field observations indicate the net energy budget of the ice is increasing. This is demonstrated by an apparent decrease in the ice volume and an increase in the duration of the open water period along the periphery of the ice pack. There is also limited information indicating a downward trend in albedo, but our knowledge of large-scale changes in albedo is incomplete. Because of the ice–albedo feedback, changes in albedo are of major importance and need additional study.

Snow cover on sea-ice

Snow acts as an insulator retarding ice growth in winter. In contrast, it also acts as a reflector reducing ice melt in summer. Both the total snow depth, and the timing of the snow accumulation, have strong influence on the energy budget. However, information on the snow cover of Arctic sea-ice is sparse, limited to results from a few field campaigns. More field studies are required, as well as the development of remote sensing techniques capable of monitoring snow cover properties over the entire Arctic ice-cover.

Outlook

Future work needs to include monitoring studies as well as efforts directed at understanding the key physical processes. The group recommended that continued monitoring of the ice-cover using satellites, field stations and autonomous buoys was essential and that these efforts be closely coordinated. For example, information on ice conditions and physical processes obtained from field campaigns can also provide the surface characterization data needed to validate remote sensing algorithms. Summer was identified as a period of great interest that merits additional study. It is a time of continual change in ice conditions that has a major impact on the mass balance of the ice-cover. Interdisciplinary studies investigating the interactions between the atmosphere, ice, and ocean will be of particular value.

Table 1. Key sea-ice parameters for climate change and feedbacks. Also listed are changes in the parameters, whether the changes were detected by observation or theory, our confidence in the results, the importance of the parameter, what future work needs to be done, and comments.

Parameter	Change	Observed or Theory	Confidence	Importance	What to do	Comment
Ice extent	Decreasing	0	High	High	Continue time series, improve spatial resolution	
Thickness distribution	Shifting			High	ULS-time series, sub- spatial surveys, ship obs	Key climate parameter
Ice type	Decreasing MY	0	High	High	Continue time series, improve spatial resolution	
Mean thickness	Decreasing	0	High	High	Increase satellite cal-val	From 1950s to 1990s
Ridges	Decreasing	0	Medium		Campaigns of subs (or auv) and aircraft	
Rafting	Increasing	T	Low		Process studies, thickness surveys, models	Difficult to determine what is rafted
Thin ice	Unknown		Low		Process studies	Important for ice production, inferred from RGPS
Leads and polynyas	Unknown		Low		Process studies	Important for ice production
Energy balance	Increasing	0	Medium			
Ice mass balance	Decreasing	0	Medium		Field campaigns, autonomous buoys. Particular ice types	Required to delineate impact of dynamics and thermodynamics on thickness decrease
Albedo	Decreasing	T	Low	High	Process studies (FY and special surfaces), satellite input, model algorithms	Key to summer melt
Length of ice free period	Increasing	0	Medium	High	Satellite analysis	Important for ice mass balance, biology, shipping
Melt pond fraction	Increasing	T	Low	High	Field campaigns	Major impact on albedo
Ice motion	Different regimes	0	High		High time resolution (hourly), coastal radars	Driven by atmosphere: large scale known, small-scale uncertainty. Influences thickness distribution
Pancake ice in Odden	Hasn't appeared in past few years	0		Medium	Ice-ocean process study	Important in Greenland Sea. Salt flux causes deep convection
Snow depth and properties	Unknown	?	Low	High	Satellite snow depth maps, field campaigns	Snow ice formation, superposed ice
Surface contaminants	Unknown	?	Low	Medium	Field studies, aerial photos	Includes sediments, pollutants, radionuclides

Working Group Report: Atmospheric Systems

Chairs: Jens Christensen (DMI) and Kim Holmén (NILU)

Participants: Rasmus Benestad (met.no), Sigbjørn Grønås (UiB), Chris Lunder (NILU) and Gudrun Magnusdottir (University of California Irvine)

The group produced the list in Table 1 that is intended to stimulate discussions about priorities in future research.

Interactions across the entire system
It is important to consider secondary feedbacks in the form of backfire from impacts in other parts of the climate system due to atmospheric changes.

Model data/experiments
The performance of models, both good and poor, should assist in defining measurement campaigns in order to target process studies to obtain intellectual knowledge and to improve model ability.

Extreme events
To start to get to grips with 'climatic events' and changes in extremes there seems to be a need for a clearer definition of what we mean by these terms. The following list is not exhaustive, but is meant to illustrate that there is a wide set of possibilities. The instruments to cope with these various categories of 'climatic events' will differ, and so will the need for data:

- Statistical outliers
- Physical definitions
- Phenomenal – complex events/risk related
- Rare events/relevance to society
- Thresholds
  - Hysteresis
  - Irreversible changes
- Quality of life
  - Human perception of extremes
  - Human perception of changes

It should be noted that many events could fall into several of these categories, but the way they would/should be analysed presumably differs. Avalanches could be used as an example. These are the consequence of accumulated atmospheric events. The impact in terms of damage is much different in most of Greenland than it is for example in Norway near Tromsø. It is obvious that climatic shifts can influence events in a multitude of ways and our current level of knowledge gives poor predictive abilities regarding the final outcome of the interactions in complex events.

Table 1. Feedbacks, current understanding of processes involved, and time-scales that need to be considered

Feedbacks	Understanding of processes involved (Modelling realism)	Time-scales
Internal atmosphere		
Clouds	Poor	Process climatic
Aerosols	Poor	
Type	Some	
Radiation	Fair	
Ozone		
Humidity	Fair	
Temperature	Fair	
Condensation/evaporation processes/extent	Fair	
Troposphere/stratosphere interaction	Poor (some)	Seasonal climatic?
Dynamical/dynamical	Poor (good)	
Polar heat transport	Fair	Seasonal climatic
Eddy transport/meso-scale activity	Poor	Daily seasonal climatic?
Blocking developments	Poor	Seasonal climatic
Terrestrial/Oceanic/Sea-ice		
Surface albedo	Fair (good)	Daily climate
Sea-ice	Fair	
Snow	Fair	
Ocean	Fair	
Land surface	Fair	
Arctic boundary layer	Poor (poor)	Daily Climate
Momentum flux	Some	
Surface roughness	Some	
Strong stratifications	Some	
Gravity waves	Poor	
Heat flux	Some	
Moisture flux	Some	
Cold air out break/marginal ice zone	Some	

Abstracts from Oral Presentations

Extended abstracts

Short abstracts

Does Climate Change in the Arctic Hold the Key to Understand Global Change?

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Introduction

Assessments of the impact of climate change must be based on estimates of possible changes in physical climate parameters. A variety of options are available to make such estimates, although it appears that there is a great deal of uncertainty associated with their use. The main emphasis in this paper is on physically based models of the climate system and the relation between global climate change and regional effects in the Arctic area. It is concluded that any estimate of Arctic climate change must be based on a global model simulation, and regional detail can be investigated using higher resolution models (particularly, regional climate models) or empirically based statistical downscaling techniques. There remains in any case great uncertainty in the quantitative estimates.

Global projections of climate change are based on a predefined emission scenario of greenhouse gases and other atmospheric constituents that can affect climate. Such emission scenarios are developed using estimates of population growth, technical developments, and other societal changes. In the ACIA assessment it has been decided to use the A2 and B2 scenarios defined in Nakićenović et al. (2000) and IPCC (2001). The model projections give a warming in the range 3-9ºC over the time period year 2000-2100 for the Arctic area. This is about twice the simulated globally averaged warming. The enhanced Arctic warming is associated with a large model-to-model scatter and a considerable interannual variability. Computing signal to noise ratios over the Arctic and globally it has been found that this ratio is actually smaller in the Arctic area than over most other areas, despite the large warming over the Arctic region. In conjunction with a temperature increase models find increased precipitation and a decreased sea-ice-cover.

Global climate models

All the global models used for climate predictions are coupled atmosphere–ocean general circulation models with a spatial resolution on the order of a few hundred kilometres and a parametric representation of processes that occur on unresolved scales. Examples of such processes are clouds, turbulent exchange at the surface, radiation and land surface heat and moisture storage. Many of these processes are poorly simulated due to the limited horizontal and vertical resolution as well as limits in our knowledge of how to treat some processes in the Arctic area. As a large part of the Arctic warming is associated with a decreased ice-cover and a subsequent increase in the ocean–atmosphere heat exchange, the surface exchange processes are a vital part of the model dynamics. Several weaknesses in surface process descriptions in many of these models are known to exist, and it is generally believed to be one of the most serious shortcomings in present day Arctic climate modelling.

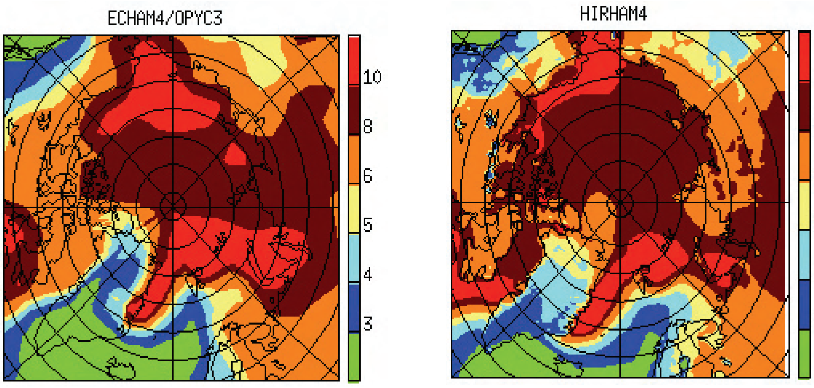

Figure 1. Simulated temperature change in winter (DJF) in (a) one of the ACIA coupled AOGCMs (ECHAM4/OPYC) and (b) downscaled using HIRHAM4. Units in ºC (from Katsov et al. 2004).

Regional climate models

On a regional scale two different options for making higher resolution climate change scenarios exist. One is based on dynamical modelling, while the other relies on statistically derived relations between large-scale flow variables and local climate parameters. The dynamical technique is exemplified primarily with simulations from the north European and central Arctic areas. It is found that in particular orographical enhancement of precipitation is handled in more detail and more accurately with a regional model than with a

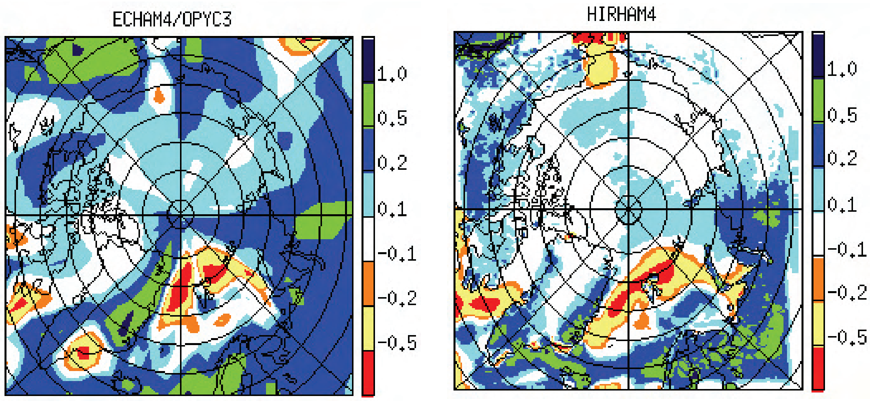

Figure 2. As Figure 1, but for local water balance (precipitation minus evaporation). Units in mm/day (from Katsov et al. 2004).

global model, but the geographical distribution of the detailed information is governed by the large-scale flow. This is illustrated in Figures 1 and 2. The driving model – in this case ECHAM4/OPYC of the Max Planck Institute of Meteorology in Hamburg – has a horizontal resolution of about 300 km, while the regional model – HIRHAM4 – has a resolution of about 50 km. As the large-scale flow is given from global model simulations, some uncertainties are directly inherited from the global to the regional simulations. This also applies to the statistically based techniques, but has not been shown in detail in this paper. For specific application purposes climate change effects can be estimated based on results from downscaling techniques, but care must be taken in choosing the actual downscaling technique to be applied.

Future challenges

To improve the reliability of Arctic climate change simulations we need to develop a better understanding of physical process descriptions in climate models. As much of the simulated Arctic surface warming is coupled to surface heat exchange and a decreased ice-cover, the Arctic boundary layer description is a particularly important area as well as the handling of low level clouds and cloud-radiation interaction. Also the Arctic Ocean deserves an increased attention from a climate modelling perspective; changes in the thermohaline circulation as well as topographic steering of dense bottom currents need to be better described in models and better understood from observations. Despite an increased knowledge of physical processes and a better representation in higher resolution models there will always remain an uncertainty in climate change simulations due to the natural variability of the climate system. This is particularly true for the Arctic where observed cycles such as the North Atlantic Oscillation and the Arctic Oscillation contribute to a large fraction of the total climate variability. Changes in greenhouse gas concentrations may not only give rise to a general warming trend, but can also trigger changes in the natural modes of variability. To better understand such changes we need to explore ensemble climate simulations in a more systematic and comprehensive manner. This requires an increase in computing capacity as well as a further development of climate models, both ocean and atmospheric components. Finally, there may still be surprises ahead of us. The climate change simulations available today rely on present knowledge of the fundamental mechanisms that governs Arctic climate. There are clear gaps in our knowledge (as highlighted by the present workshop), and there may still be some important processes that are not yet fully understood and that may prove to be essential for simulating and understanding Arctic climate change.

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Some Aspects of Climate Variations in the Arctic

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Introduction

Climate changes in the Arctic during the last century, generally characterized by amplification of anomalies from mid-latitudes, are documented from observations (Hanssen-Bauer & Førland 1998; Førland & Hanssen-Bauer 2000; Jones et al. 1999; Serreze et al. 2000, Moritz et al. 2002; Bengtsson et al. 2003; Johannessen et al. 2003). Significant decadal climate variations took place, expressed in many climate parameters such as surface air temperature (SAT), precipitation, sea-ice extent, snow cover, permafrost temperature and vegetation. Ice and possibly snow albedo feedback are regarded as the main amplifying mechanisms. Some decadal variations are considered to be natural internal variation of the climate system, some are due to external radiative forcing. Further back in Holocene, an increasing amount of proxidata indicates considerable variations on decadal, multi-decadal and millennium scales. Variation in solar activity is believed to have been the main external forcing (e.g. Bond et al. 2001).

In this paper three problem areas are reviewed briefly: the amplification of decadal variations towards the Arctic, the response of an external forcing on the Arctic Oscillation (Thompson & Wallace 1998) (AO, here virtually synonymous with the North Atlantic Oscillation), and the role of air–sea interaction in the North Atlantic for multi-decadal variations.

Amplification of decadal variations

The main findings of the investigations, referenced in the beginning, might be illustrated by SAT anomalies as a mean for the area north of 60°N, compared to mean anomalies for the northern hemisphere (Figure 1 left) and zonal means of SAT anomalies (Figure 1 right). The two well-known multi-decadal warming trends, and the cooling trend between them, are clearly seen (the early last century warming after 1910, the recent warming during the latest decades and the cooling from the warm forties to the cold sixties). The amplification of the variations in the Arctic is evident in all three periods, but seemingly smaller for the recent than the early warming. The early warming was not connected to the AO at all, while the recent warming is strongly projected on a trend in the AO-index during the last decades. While IPCC (2001) indicates that the early warming and the cooling might be explained by a combination of natural and anthropogenic external forcing, Bengtsson et al. (2003) and Johannessen et al. (2003) suggest that the event might be a random internal variation. A suggestion for this is nearly similar variations found in a control climate simulation without external forcing. Another indication is given by Delworth & Knutzon (2001), who in an ensemble of runs with a global climate model through the last century, forced by anthropogenic greenhouse forcing only, found one member that gave a reasonably accurate simulation of the early warming.

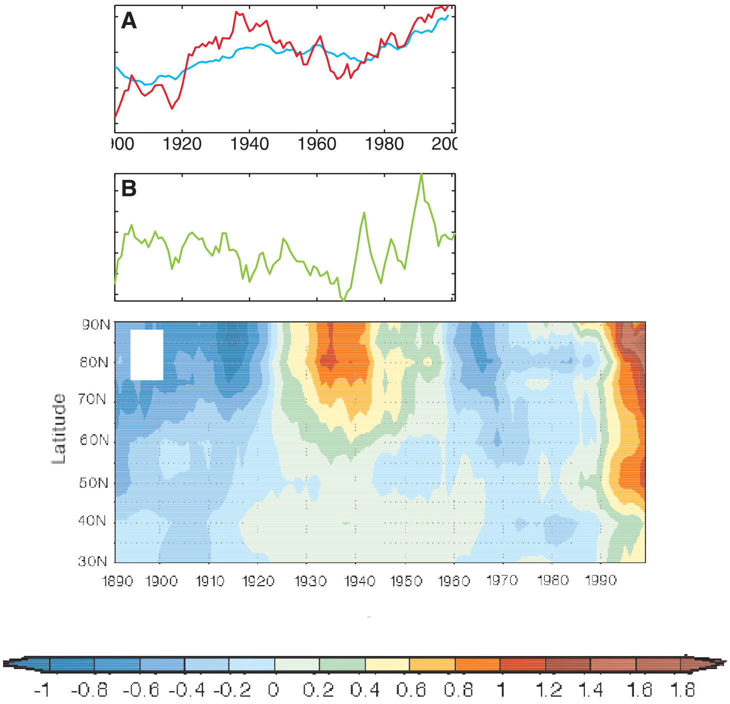

Figure 1. a) SAT anomaly variations north of 60 °N (red) and for the northern hemisphere (blue). b) The AO-index. (Moritz et al. 2002).
Below: Zonal mean anomalies of SAT (Johannessen et al. 2003).

CMIP2 runs, where the CO2 greenhouse forcing is increased by 1% per year to 80 years (doubling of present day concentrations after 70 years), show largest impact in the Arctic. However, such runs also show a particularly large spread between the models north of 60°N. This was demonstrated by Räisänen (2001) for zonal means of SAT anomalies in 19 models around doubling of CO2 (results averaged over a time slice of 20 years between 60 to 80 years in the simulations). One interpretation of this has been that the models have special problems in simulating variations in the Arctic, e.g. problems connected to the ice-module of the models. A recent ensemble of six CMIP2 runs made by the Bergen Climate Model (Asgeir Sorteberg, unpublished), where the initial conditions in the ocean are taken at different levels of AMOC (Atlantic Meridional Overturning Circulation) in a long control run, show large spread in trends over 30 years (Figure 2). When the trends are averaged for a period more than 40 years, the spread does not increase towards the Arctic. This is demonstrated in Figure 2, where trends from 1 to 80 years show a small, but similar spread all the way from mid-latitudes to the pole. This result indicates more internal decadal and multi-decadal climate variations in the Arctic than at lower latitudes. In all CMIP2 runs, including those from Bergen, the response of greenhouse warming shows a maximum in the Arctic. However, since natural decadal

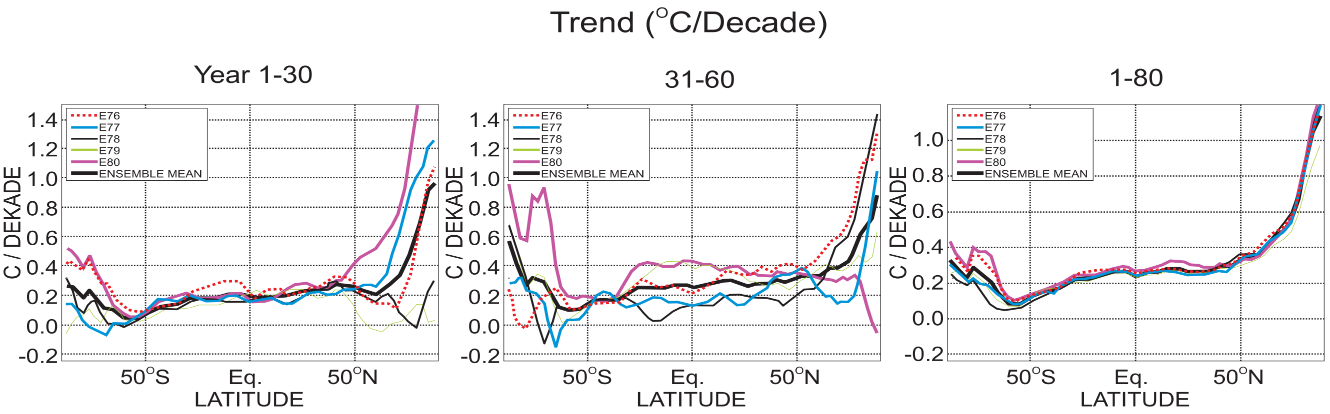

Figure 2. Results from CMIP2 experiments with the Bergen Climate model (Furevik et al. 2003; resolution T63). SAT trends for each member of an ensemble of six runs: trends 1-30, 31-60 and 1-80 years (Asgeir Sorteberg, Bjerknes Centre for Climate Research).

variations are larger than elsewhere, the Arctic might not necessarily be the best place to look for signals of anthropogenic global warming. Further studies should be done to understand decadal variations in the Arctic, both from observations, proxy data from the Holocene and control runs and other experiments with high-resolution climate models. Focus should be put on the mechanisms that regulate ice and snow albedo feedback, first of all feedback from the clouds.

Response of radiative forcing on the AO

It has been shown that the recent warming in the Arctic is connected to a positive trend in the AO-index during the last decades (Moritz et al. 2002; Bengtsson et al. 2003; Hanssen-Bauer 2003) (Figure 1). This means that the warming shows a different geographical structure than the early warming. The Arctic amplification is less, and the warming signal stronger over northern continents (Siberia and the northern parts of North America). The variation includes three poles of SST anomalies in the North Atlantic (e.g. Visbeck et al. 1998). If the trend in the AO-index can be explained, much of the recent warming is also explained. The trend in NAO has been simulated in ensemble runs with global models of the atmosphere forced with observed SST (Rodwell et al. 1999; Hoerling et al. 2001). Hoerling et al. indicate that the trend is caused by warming and stronger convection in the tropics, assumed to be a consequence of anthropogenic global warming. However, the simulated trend is only half of the observed trend. It has also recently been shown that variations in the AO are connected to dynamical interaction between the stratosphere and the troposphere (Shindell et al. 1999; 2001). In this process, Shindell et al. (1999; 2001) seem to demonstrate that interactive treatment of ozone is important in the models, at least when solar activity represents the main radiative forcing. Few climate models treat ozone interactively. A hypothesis for changes in NAO has been established from recent studies (e.g Shindell et al 2001). For a positive forcing, it starts with higher SST and stronger convection in the tropics. This gives stronger latitudinal temperature gradients and zonal winds in the lower stratosphere. This in turn affects the wave propagation in the troposphere both towards the tropics and towards the stratosphere in such a way that the transport of zonal momentum in the troposphere increases towards the north. In this way the AO is strengthened in its positive phase. Similarly, a negative external forcing results in a low AO by virtually opposite effects. Interesting interaction mechanisms between the troposphere and stratosphere have been proposed by Ambaum et al. (2001).

Simulated variations in the Maunder Minimum

Opposite to the warming periods during the last century, the Maunder Minimum (MM; 1675-1710 AD) gives an example of extensive cooling in the Arctic, probably caused by a minimum in solar activity and frequent eruptions of sulphur gasses from volcanoes. It is interesting to compare simulations for the MM made by Shindell et al. (2001), where the ocean was represented by a mixed layer (one ocean slab), and simulations by Fischer-Bruns et al. (2002) made with a fully coupled climate model. Shindell et al. made equilibrium runs with estimated forcing conditions after and under the MM, while Fischer-Bruns et al. ran the model from 1550 to 1800 and investigated the anomalies under the MM. Shindell et al. included interactive ozone and used only solar forcing, while Fischer-Bruns et al. had no active ozone, but included forcing from known volcano eruptions in their runs.

Both experiments gave large responses in the Artic: higher surface pressure most of the year and colder climate over the continents. In this respect the AO was weak with a circulation dominated by outflow from the Arctic, with prevailing north and northeasterly winds, e.g. east of Greenland. The results from Shindell gave a cold land–warm sea SAT pattern with positive SST anomalies over the northern Atlantic, with a maximum southwest of Iceland. However, Fischer-Bruns et al. got a negative anomaly in the same area, with a minimum in the Labrador Sea, close to the southern tip of Greenland. There are few observations to validate the results, but seemingly, the coupled simulations are in better agreement with historical evidence. For instance, the sea-ice was probably in an extreme southerly position, e.g. reported from Icelandic chronicles (e.g. Bergthorsson 1969; Lamb 1982). Another indication is the same sign of SST anomalies over the ocean basin between North America and Europe in agreement with proxydata for cold periods in the Holocene (Bond et al. 1999; 2001). It is believed that climate deterioration played a significant role when the Nordic settlement on Greenland died out around 1350 (Barlow et al. 1997). The change could have been caused by a similar event to that in the MM. However, rather little is known about anomalies in solar forcing at that time.

Air–sea feedback close to the ice edge

Hall & Stouffer (2001) examined variations in SAT in the southern Greenland/Iceland area in a long control run with a coupled climate model. They found one event with much colder climate (anomalies up to 4 °C) for a period of 40 years. The event was characterized by excessive northerly winds east of Greenland. It is known that the heat loss from the ocean is tremendous in outbreaks of cold air (e.g. Grønås & Skeie 1999), but normally the loss is distributed to large depths. In this way change in SST is modest. The density of the surface water in the area is mainly determined by the salinity. Hall & Stouffer found that the surface water became much fresher (a huge salt anomaly) during the mentioned event. In this way the stratification in the upper layers became much more stable. Consequently, the heat loss was distributed in a shallow layer only, resulting in a large reduction in SST. It might have been active and played an important role in the formation of the basin-wide negative SAT in the simulations by Fischer-Bruns et al.

The feedback mechanism above probably affects the Atlantic Meridional Overturning Circulation (AMOC) in such a way that it becomes weaker. The feedback of variations in AMOC on SAT has been investigated by several, e.g. from control runs in climate models. For instance, composites of differences in SAT, between years with variations more than one standard deviation in AMOC and years with less than minus one standard deviations, have been made in the Bergen Climate Model (Bentsen et al. 2003). They show variations with the same sign in the North Atlantic basin, but no signal of the feedback mechanism in the Greenland/Iceland area. Further research is certainly needed.

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Evaluation of the Sea-Ice Components of the ACIA AOGCMs

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Introduction

Global climate models predict a maximum warming in the high latitudes of the Northern Hemisphere. Polar amplification is mainly due to the positive feedback mechanisms related to the insulation and albedo effects of the snow and sea-ice. True magnitude of the feedback effect of the ice/snow surface is still unknown because of the incomplete physical description of snow and ice physics and several other unresolved processes of the Arctic climate. A clear indicator of this is that the projected changes exhibit a large range of warming scenarios in the Arctic (Räisanen 2002; Walsh & Timlin 2003). Holland & Bitz (2003) have found that the modelled mean sea-ice state of the present climate has a considerable influence on the polar warming in the climate models. In the following the effect of the physical description of the sea-ice on the modelled mean state of the sea-ice conditions is analysed in detail, and the influence of simplifications of the sea-ice physics on the climate predictions is discussed.

Physical description of sea-ice in climate models

Horizontal variability of the sea-ice thickness in a continuum scale is described by the ice thickness distribution function g(h) (Thorndike et al. 1975). The most common approximation in the climate models is that only one thickness and concentration category is resolved and redistribution terms are neglected, commonly called two-level models (Hibler 1979). More advanced sea-ice models resolve evolution of the sea-ice thickness of several categories in the eulerian (Hibler 1980; Flato & Hibler 1995) or langrarian ice thickness space (Haapala 2000; Bitz et al. 2001).

The first order approximation of the momentum balance is that acceleration, sea surface tilt and internal stress terms are assumed to be negligible, which leads to the free drift model. The cavitating fluid model assumes that the internal stress depends only on the bulk stress of the ice. In the viscous-plastic model (Hibler 1979), the internal stress is determined by the nonlinear bulk and shear stresses and the ice strength. In two-level models, ice strength is proportional to ice mass and concentration, while in the multi-category model ice strength is directly related to the energy consumed during deformation (Rothrock et al. 1975).

The simplifications of the sea-ice physics made in the climate models differ considerably. None of the models selected for the Arctic Climate Impact Assessment (ACIA) resolve sea-ice physics based on the fully primitive equations (Table 1).

Numerical experiments

In order to determine how sensitive the modelled mean ice state is for the simplifications of ice physics, we used the multi-category sea-ice model with the prescribed atmospheric and oceanic conditions. The model has a global coverage. However, the focus is on the Arctic Ocean because it employs orthogonal curvilinear coordinates. The co-ordinate system is equivalent to the Max Planck Institute global ocean–ice general circulation model MPI-OM1 where the poles are located over the Canada and western Siberia (Marsland et al. 2002).

Atmospheric forcing was obtained from the German ocean model inter-comparison project OMIP (Röske 2001). The OMIP data provide the mean annual cycles of daily surface fields, superimposed with a synoptic scale variability. The data is derived from the European Centre for Medium-Range Weather Forecast (ECMWF) 15 years re-analysis (Gibson et al. 1997). The ocean surface temperature and currents were obtained from the previous MPI-OM1 simulations.

The evolution of the ice pack was simulated with the thermodynamical only model (TDM), the free drift model (FDM), the viscous-plastic (VPM), the viscous-plastic model with an island on the North Pole (VP-NPM) and with a multicategory model (MCM). In the multicategory model five underformed and two deformed ice categories were used. All simulations begin from the same initial conditions, and stationary conditions were obtained after ten years integration. In order to analyse the uncertainties of the modelled future mean ice conditions of the climate predictions, models' sensitivity to change of thermal forcing was tested. In these simulations air temperature was increased by 4 K, all other forcing factors were the same as in the control simulations.

Results and conclusions

Comparison of different ice model results shows that the physical description of ice dynamics does not have significant effects on the modelled maximum ice extent. Some differences are noticeable in the coastal regions, in the ice edge and, in particular, in the Barents Sea. Major differences are in the modelled ice thickness fields. TDM produces a uniform slab of 3.5 m thick ice in the central Arctic. FDM generates ice thicker than 20 metres to the convergent regions. VPM and VP-NPM produce an ice thickness pattern that has about similar horizontal distribution as that deduced from the observation (Bourke & Garrett 1987), except that the ice is thinner than expected. An artificial North Pole has some blocking effect, especially in the Siberian side of the island. In this resolution, the ice can still drift from the central Arctic to the Fram Strait and hence heavy accumulation of ice to the Arctic is impossible.

In September, the models produce rather different results both with regard to the ice extent and thickness. It is remarkable that the ice extent of FDM is much larger compared to the other model results. In the FDM simulation, the Siberian coastal region is completely covered with two-metre thick ice, while the other models simulate almost ice-free conditions in these regions. The bias of the FDM is certainly due to the overestimation of the dynamical thickening of the ice.

Modelled annual maximum ice extent is rather insensitive to the ice dynamics or ice thickness distribution used. This indicates that the maximum annual ice extent is determined by the thermodynamical growth of new ice, and that correct modelling of sea surface temperature is the most important factor. In addition to the inaccuracies in the surface heat balance, an overestimation of the modelled ice extent in climate models can be the result of an underestimation of the Atlantic heat transport or vertical mixing.

Minimum ice extent is sensitive to the modelled ice thickness, which in turn, is highly dependent on the ice dynamics or thickness distribution used. Overestimation/underestimation of the dynamical growth of sea-ice leads to overestimation/underestimation of ice mass and to a situation where ice does not melt/survive during the summer stage. This clearly shows that a mass–momentum coupling is essential for sea-ice and only the plastics models are physically realistic in climate simulations. All other models generate highly unrealistic ice thicknesses that are commonly hidden, causing a numerical diffusion or an artificial upper limit for ice thickness.

The response of the sea-ice model to the changes of the thermal forcing depends on the modelled mean state of the control climate and thickness distribution used. In particular, beyond a certain degree of warming, two-level models may predict total disappearance of sea-ice in the Arctic, but multi-category ice models may predict existence of thick-ridged ice also during the summer season.

In coarse resolution models, an artificial North Pole island may cause accumulation of ice, but its effect on the modelled ice extent is minor. Utilising regional climate or MPI-OM1 type models may overcome the problem. These models will also provide more detailed information on the future changes in the Arctic.

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Table 1. Sea-ice physics of the climate models used in the ACIA

model	ocean component	ice thermodynamics	ice dynamics
CCCma	MOM-1	energy balance	modified free drift
GFDL	MOM-1	?	free drift
ECHAM	OPYC3	simple thermodynamics	viscous-plastic
HadCM3	Cox-Semtner	0-layer Semnter	modified free drift
NCAR/CSM-1	MOM-1	3-layer Semtner	cavitating fluid

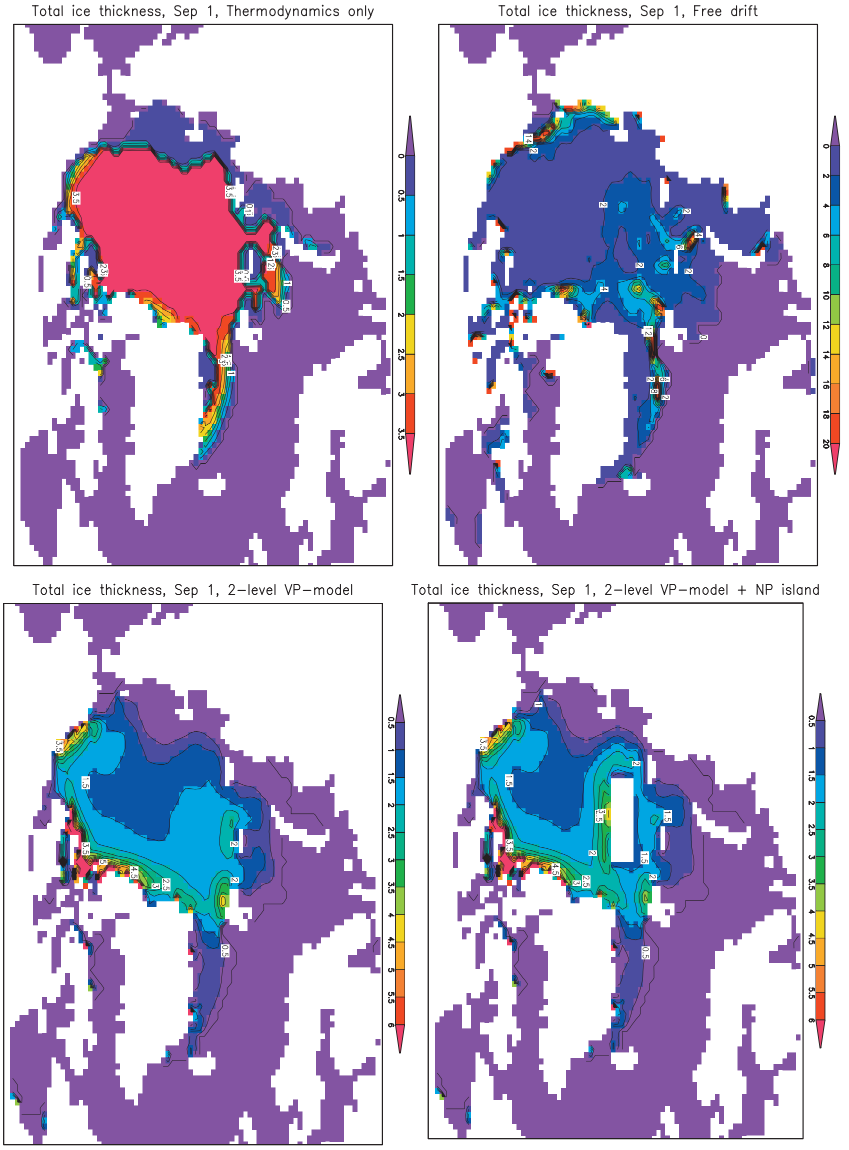

Figure 1. Modelled ice thickness on 1 September using different models for the present climate: a) thermodynamic only b) free drift c) two-level and d) two-level with island on the North Pole.

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Hydrological Changes in Polar Regions, Impacts of a Changing Climate

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Effects of a warming climate

The effects of a warming climate on the terrestrial regions of the Arctic are already becoming apparent; some subsequent impacts to the hydrologic system are also becoming evident. It is expected that the effects and consequences of a warming climate will become even more evident within the next 10 to 50 years. The broadest impacts to the terrestrial Arctic regions will result through consequent effects of changing permafrost structure and extent. As the climate differentially warms in summer and winter, the permafrost will become warmer, the active layer (the layer of soil above the permafrost that annually experiences freeze and thaw) will become thicker, the lower boundary of permafrost will become shallower and permafrost extent will decrease in area.

These simple structural changes will affect every aspect of the surface water and energy balances. As the active layer thickens, there is greater storage capacity for soil moisture and greater lags and decays are introduced into the hydrologic response times to precipitation. When the frozen ground is very close to the surface, the stream and river discharge peaks are higher and the base flow is lower. As the active layer becomes thicker, the moisture storage capacity becomes greater and the lag time of runoff increases. As permafrost becomes thinner, there can be more connections between surface and subsurface water. As permafrost extent decreases, there is more infiltration to groundwater. This has significant impacts on large and small scales. The timing of stream runoff will change, reducing the percentage of continental runoff released during the summer and increasing the proportion of winter runoff. This is already becoming evident in Siberian rivers. As permafrost becomes thinner and is reduced in spatial extent, the proportions of groundwater in stream runoff will increase as the proportion of surface runoff decreases, increasing river alkalinity and electrical conductivity. This could impact mixing of fresh and saline waters, formation of the halocline and seawater chemistry. As the air temperatures become higher and the active layer becomes thicker, we have reason to believe the surface soils will become drier. As the surface soils dry, the feedbacks to local and regional climate will change dramatically, with particular emphasis upon sensible and latent heat flux. This may impact recycling of precipitation, capabilities to predict weather and may indeed increase variability of many processes and variables, including convective storms.

Hydrological processes

The role of Arctic hydrology in understanding and predicting the effects of global climate change is emphasized in every comprehensive document on the subject and the responsibility inherent to responding to these issues is of paramount importance. In the Arctic, small changes in climate as well as direct human impacts can have enormous and lasting consequences on the land surface and its dynamics. Recent research studies have revealed many interesting changes in the hydrological processes in response to a changing climate in Sub-Arctic Alaska and portend potential impacts to Arctic regions (Oechel et al. 2000; Serreze et al. 2000). Complementary field investigations being conducted in tussock tundra regions of the Arctic and the Sub-Arctic reveal differences in climatic drivers and the consequent ecosystem responses (Henry & Molau 1997).

The primary control on local hydrological processes in northern regions is dictated by the presence or absence of permafrost, but is also influenced by the thickness of the active layer and the total thickness of the underlying permafrost. As permafrost becomes thinner or decreases in spatial extent, the interaction of surface and sub-permafrost ground water processes becomes more important (Woo 1986). The inability of soil moisture to infiltrate to deeper groundwater zones due to ice rich permafrost maintains very wet soils in Arctic regions. However, in the slightly warmer regions of the Sub-Arctic, the permafrost is thinner or discontinuous. In permafrost-free areas, surface soils can be quite dry as infiltration is not restricted, impacting ecosystem dynamics, fire frequency and latent and sensible heat fluxes. Other hydrologic processes impacted by degrading permafrost include increased winter stream flows, decreased summer peak flows, changes in stream water chemistry and other fluvial geomorphological processes (McNamara et al. 1999). Hydrologic changes witnessed among study sites include drying of thermokarst ponds, increased active layer thickness, increasing importance of groundwater in the local water balance and differences in the surface energy balance. By far, the most significant changes occur in response to changing permafrost extent or thickness. As permafrost becomes thinner, the sub-permafrost groundwater becomes more important, either by contributing groundwater to streamflow, or allowing surface water to drain (Hinzman et al. 2003; Carr 2003).

Geomorphological processes

Perhaps the most important geomorphological processes following a warming climate will be those associated with thermokarsting. Thermokarst topography forms as ice-rich permafrost thaws, either naturally or anthropogenically, and the ground surface subsides into the resulting voids. The important and dynamic processes involved in thermokarsting include thaw, ponding, surface and subsurface drainage, surface subsidence and related erosion. These processes are capable of rapid and extensive modification of the landscape and preventing or controlling anthropogenic thermokarsting is a major challenge for northern development. The active layer is that portion of the soil above permafrost that seasonally experiences thawing and freezing. The depth to which the active layer will thaw each summer season depends upon many local factors, especially site hydrology. Other seasonal factors that influence depth of thaw include temperature and levels of soil moisture due to variation in precipitation and evapotranspiration. The interannual variation of thaw depth at a site is quite large and consequently, utilizing depth of thaw as an indicator of climatic change may be quite difficult as one would be looking for the response to a subtle change amidst large annual variations; however, the deeper permafrost acts an integrator of meteorological variations and will respond to long-term changes in climate. It must also be recognized that changing the surface configuration and condition (i.e. disturbance such as wildfires, construction or mining) will also impact deeper ground temperatures and thus may mask changes in temperature due to changing climate.

Results

Extensive thermokarsting has been discovered near Council, Alaska, U.S.A. (Yoshikawa & Hinzman 2003). Our studies relating changes in hydrologic processes to changes in climatic dynamics have focused upon these permafrost features to better understand current hydrologic and climatic dynamics and to develop better prognostic tools for predicting future hydrological changes. This investigation of the physical factors that influence thermokarst pond formation indicates that in warming climate, the permafrost will degrade and these ponds will drain. In response to some imposed disturbance, such as a tundra fire or climatic warming, massive ice permafrost may differentially thaw, creating irregular surface topography. Depressions forming on the surface soon form ponds, accelerating subsurface thaw through lower albedo and additional heat advected into the pond through runoff. In time, a talik (a layer of unfrozen soil above the permafrost and below the seasonally frozen soil) may form below such ponds as the depth of water becomes greater than the amount that can refreeze during the winter. If the talik grows to a size that completely penetrates the underlying soil or connects to a subsurface layer that allows continued drainage, the pond may then begin to drain.

Conclusions

The important implications are that in regions over thin permafrost (~<20 m), surface ponds may shrink and surface soils may become drier as the permafrost degrades. In areas over thick permafrost, degradation of massive ice wedges could thaw and catastrophically drain an entire village's water supply. These processes depend upon many complicating factors, such as the regional hydrologic gradients (i.e. whether the region is a groundwater upwelling or downwelling zone). The same mechanisms that allow drying of the ponds may also cause soil drying with significant impacts to latent and sensible heat fluxes. In-depth study and collaboration with villages is needed to project current capacities, future needs and future threats. Arctic village residents need to be trained in appropriate water use, both for current and future water utilization.

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Feedbacks between Northern Terrestrial Systems and Climate

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Summary

Feedbacks between northern terrestrial systems and climate involve a broad range of scientific fields but the scope of this presentation is narrowed down to interactions between the terrestrial vegetation and climate and in particular the significance of main structuring components of the vegetation such as occurrence of forests, woodlands and trees at local, regional and circumpolar scales. High latitude forests and tundra areas provide essential feedbacks to the global climate through for example their capacity to store large amounts of carbon and their ability to transform solar radiation to sensible heat. A change in growth rate and area covered by trees would change the CO_2 uptake capacity of the regions, and replacement of tundra by forest (in particular evergreen) would decrease regional albedo and thus enhance warming. Further, changes in tree cover interact with the occurrence and distribution of continuous and discontinuous permafrost. Consequently, a change in range position would involve a series of large ecosystem consequences with repercussion on occurrence and vitality of forests and trees in the transition zone towards the Arctic and thus further affect the Arctic climate.

Introduction

Understanding the behaviour of climate systems requires profound knowledge of its components – atmosphere, hydrosphere, lithosphere and biosphere – and how these interact with each other, but holistic approaches are sparse and difficult to achieve. To a large extent, the diversity of processes and changes within the terrestrial component has been excluded or avoided in attempts to model future development of climate systems. The system is non-linear and natural variability of feedback processes and responses to external forcing make the system highly complex. Use of numerical models, present-day climate data and reconstructed paleoclimate data offer analyses of a large range of possible states, which are needed as future climate is foreseen to differ significantly from last centuries. However, the structure of ecosystems making up the biomes of today are not identical to the ones that formed the paleo- archives, and can not be assumed to respond in the same way to abiotic and biotic forces. Consequently, profound knowledge and data on recent and ongoing responses to climate is needed and has to be included in models of future responses and development of the Arctic climate.

A continuously shifting balance between dominating air masses has changed and is still changing the structure of, for example, the boreal forest, the forest-tundra ecotone and the tundra. These structural changes include shifts in range limits of species and systems. During the last 10 000 years or so the main range limit shifts have been retraction, and in a Holocene perspective a large part of the area north of the present Arctic treeline belongs to the zone with discontinuous to continuous boreal forest and woodlands. Thus, the zone between the closed boreal forest and the treeless tundra including latitudinal treelines has a long history of response and adaptation to changes in climate conditions (cf. Hofgaard 1997). It is generally assumed that this climate-sensitive ecotone, including the northern limit of the boreal forest, is determined by the summer position of the Arctic Front. However, the boreal forest itself may significantly influence the position of the front, mediated through forest structure and surface roughness at local, landscape and regional levels.

Modelling future vegetation

Models of future vegetation distribution suggest a rapid and dramatic invasion of the tundra by taiga. Such changes would generate both positive and negative feedbacks to the climate systems. However, the balance between negative and positive feedbacks is largely unknown as important variables such as snow (timing, duration, quality), changed forest cover, peatland performance and fresh water flows into the Arctic Ocean, are to a varying degree or not at all included in the models. Additionally, feedbacks amplifying climate warming have been of main concern during the last decades, thus restricting interpretation of a broader range of potential responses. Compilation of some commonly discussed feedbacks could be presented as follows: Increased CO_2 levels → Increased temperature • Reduced snow cover • Increased growing season length • Increased forest growth • Increased absorption of solar radiation • Heating of the lower atmosphere • Increased forest growth • Increased carbon uptake • Decreased CO_2 levels • Decreased air temperature • Decreased growing season length • Decreased forest growth. Decreased forest growth would in turn lead to increased CO_2 levels making the entire feedback loop circular. However, there are many uncertain links in the system. For example, there is no straightforward relationship between increased temperature and reduced snow cover, or between reduced snow cover and increased growing season length. The main reason is prevailing regional differences along latitudinal gradients, coast–inland gradients, and along altitudinal gradients conditioned by the mean position of 0°C in space and time (Crawford 2000; Solberg et al. 2002).

Consequences and feedbacks

Most models of future responses, generally based on present day biome distributions and future biome distributions as predicted by $2xCO_2$ levels, forecast a latitudinal shift in the location of the forest–tundra ecotone by several hundred kilometres. Such changes would have significant consequences for ecosystem diversity and functioning, regional socioeconomy as well as consequences for the climate system at the regional to global scale. Changes of the forest–tundra zone strongly influence energy exchange between the biosphere and the atmosphere. The involved feedbacks between the physical climate and vegetation are essential for the entire circumpolar north. These feedbacks occur through energy exchanges at the surface, but extend through the lower and, ultimately, the entire atmosphere. The primary control on the energy balance is the surface albedo (the proportion of the incoming solar radiation that is reflected) – which determines the overall absorption of solar radiation at the surface. The contrast in surface characteristics between the tundra and the boreal forest is considerable, particularly in the winter when the tundra is snow covered. This variation causes massive changes in energy fluxes at the surface and hence temperature on the ground and in the atmosphere, with consequences for vegetation development, carbon fluxes, permafrost and hydrology. The heat flux into the ground is determined by the surface roughness, the physiological properties of the vegetation and the thermal properties of the vegetation and upper soil layers. All these surface properties are substantially altered by the presence of a snow cover. However, the nature of the vegetation will itself influence the distribution of snow and thus the impact of snow cover – leading to a number of additional interactions.

A range of expected vegetation responses to climate warming includes for example increased growth and productivity through increased photosynthesis and nutrient availability, increased seed reproduction vs. vegetative reproduction, range adjustments and new structure and composition of vegetation communities. However, it can be questioned to what degree this has been recorded and if there are other major impact factors or processes that complicate the scene. The answer to the first question is not straightforward, and there is a lot of published information confirming both expected vegetation responses and indifferent or opposite responses. The answer to the second question is of course Yes, and the main factors are human land use, grazing pressure, vegetation inertia, recovery from previous disturbances and shift in disturbance regime. Depending on the nature (kind, frequency and magnitude) of these factors or disturbances to the system, it may take several centuries to recover to the pre-disturbance state. However, in the meantime climatic prerequisites for the previous vegetation state will have changed and the previous state will thus not be reached. Consequently, it can not be known where studied systems are situated along the disturbance–recovery time gradient, and systems situated at for example relict positions (i.e. stronger correspondence to previous environmental conditions than present) will respond completely different to imposed disturbances compared to systems in balance with recent environmental conditions. Further, recent estimates of annual change in plant growth (net primary production) using satellite images and climate data for 1982 to 1999 show large differences at the regional scale around the circumpolar north (Nemani et al. 2003). Increased growth is shown for central Alaska, eastern Canada, Fennoscandia and the European Russia, and decreased growth for northern Siberia to eastern Alaska, northwestern Canada and the Kola Peninsula. Considering this distribution of regions with net increase and decrease, respectively, there is an apparent risk of overestimating responses at the circumpolar level as available measured ground data for inclusion in models of future responses mainly represent areas with increased growth.

The forest–tundra boundary

The forest–tundra boundary includes a set of distribution limits (timberline, forest-line, treeline, species line, historical treeline) that are of significant importance in studies of spatiotemporal responses to environmental changes. Towards Arctic areas, the boundary may cover hundreds of kilometres and towards alpine areas some hundred altitudinal metres. Along mountain chains, intersecting into the Arctic a complex system of both alpine and Arctic responses has to be considered. The scientific field of treeline dynamics considers the entire transition zone from the closed forest to the treeless tundra. Different environmental factors and disturbance agents are acting at the different limits and their individual importance change through time. For example, fire may be a strong determinant in the forested part of the transition zone but of no significance at the species line in some regions, but be temporary significant throughout in other regions; and wind might be highly significant through time at the species line and treeline but only rarely or occasionally at the timberline.

Since the mid-Holocene, the forest–tundra boundary vegetation in the northern hemisphere displays generally decreasing stand density, retreating forest and treelines and changes in species composition as a consequence of deteriorating climate conditions. The climatic improvements since the Little Ice Age (i.e. end of the 19th century) have caused stand densities to increase and different distribution limits to advance. However, there are large regional differences including evidence of fragmentation and withdrawal, and large differences between specific time periods. For example, the climatic improvement (warming) during the 1930s resulted in general increased tree growth but the warming during the 1990s did not show the same response. Structural changes result from a series of mechanisms, governed directly or indirectly by climate variability and small to large-scale disturbances. Disturbance is a central factor in vegetation dynamics and in translating climatic change into vegetational response. Consequently, knowledge of the disturbance regime is essential for understanding and modelling system responses. Factors such as temperature and precipitation regime (trends and variability), precipitation quality, occurrence of warm spells in winter and cold spells in early growing season, moisture conditions (drought vs. saturation), wind activity, changes in frequency, intensity and distribution of fire, fungi, and herbivores (insects, ungulates, lemmings) and human impact are important to consider in order to reach reliable predictions.

Insect outbreaks have recently received renewed regional attention. In Alaska, for example, spruce bark beetle outbreaks over the last 15 years have killed trees over an area twice the size of Yellowstone National Park, possibly as a consequence of a long run of warm summers. In northern Fennoscandia, the area affected by autumnal moth outbreaks has increased. These insects, that during severe outbreak years may kill birch forests over large areas, have a cyclic occurrence with some 12-15 years between regional peak years. The outbreak frequency is predicted to increase in a warmer climate and could thus cause radical structural changes at local, landscape and regional level. Forest recovery to the pre-outbreak state at these different spatial scales depends largely on intensity and spatial extent of the outbreak, post-outbreak climate conditions and species interactions. Host species will suffer at the expense of non-host species leading to changed distribution and balance between for example evergreen and deciduous tree species. In addition to tree layer changes, the outbreaks generally cause field layer changes commencing new grazing regimes dominated by other herbivores operating with deviating intensity and frequency compared to the pre-outbreak system. These interactions may, together with changes in the dominating climate regime, cause a shift in the state of the system characterized by new dominating tree species or formation of Sub-Arctic tundra areas with long-term resistance to tree invasion. These alternative paths of development will produce completely different feedbacks to the climate system both when compared to each other and to the pre-disturbance vegetation.

The future

Considering that terrestrial areas cover vast parts of the circumpolar north and that the forest–tundra boundary stretches approx. 14 000 km around the Hemisphere, any change of its position or structure will cause massive changes of the feedback system between the climate, other components of the abiotic environment and the biotic environment. Additionally, the climatic significance of the terrestrial component in regional to global climate systems has changed through history due to variation in the area covered by forest, tundra, wet lands and water. In the predicted future, its importance is likely to increase along with for example the decrease in Arctic sea-ice distribution as a consequence of spatial changes in mean seasonal and annual albedo around the circumpolar north. Consequently, there is an urgent need to include a broader range of terrestrial components, including internal and external feedback mechanisms, into models aiming at realistic regional and circumpolar climate scenarios.

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Can Arctic Ice Cores Provide Insight to Climate Feedbacks?

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Introduction

Through various paleoclimatic records we know that climate in the past has been widely variable and that in many cases the changes have been very rapid with sudden jumps. Much of this knowledge comes from ice core records. Feedback mechanisms in climate are often defined as rapid changes occurring when the system is pushed over a threshold. Thus, one way to study feedback mechanisms is to study rapid climate change and the processes involved. Without doubt, ice cores have been one of the most important archives.

The longest ice core records extend as far back as 150 000 years in Greenland (e.g. Johnsen et al. 2001) and 900 000 years in Antarctica (The EPICA group, unpublished). One of the advantages compared to other proxy records is the possibility to study high resolution records – in the most favourable cases seasonal variations can be seen as far back as in the last glacial period (Alley et al. 1993). Through a number of different chemical analyses ice cores have the potential to provide proxies for a wide range of other climate indicators such as sea-ice, biomass concentration, marine biogenic productivity and atmospheric circulation intensity (Legrand & Mayewski 1997). Therefore ice cores can provide insight to possible feedback mechanisms for rapid changes because of the variety of analyses that can be performed in the same "time window". The drawbacks include the limited spatial coverage and the risk for sampling a "locally created" climate. The latter is particularly important to consider when studying ice core records from the large ice sheets such as Greenland.

Also, the large ice caps in the Canadian Arctic have provided valuable climatic records mainly through the Holocene (e.g. Koerner & Fischer 1990; Fischer & Koerner 2003), but there are many smaller ice caps around the Arctic that potentially could provide climate records from a wider geographical area. So far, ice cores from low-altitude Arctic ice caps have provided records over the last 1000 years in particular (e.g. Koerner 1997; Tarrusov 1992; Watanabe et al. 2001).

Rapid climate oscillations detected with ice cores

Rapid oscillations between low and high $\delta^{18}O$ and high and low dust concentrations regularly spaced with about 1500 years were revealed in a Greenland ice core in the 1970s (Dansgaard et al. 1982). These were later named Dansgaard-Oeschger events and are often used as the best example of how fast climate has been changed in the past. In the Greenland ice cores it was found that the methane changes lag about three decades behind the $\delta^{18}O$ changes, implying methane concentrations are driven by millennial oscillations. These rapid changes are thus not restricted to the polar areas. The lack of rapid oscillations in Antarctic ice cores led to the conclusion that the changes were somehow related to the intensification of the North Atlantic Deep Water (NADW) formation, thus an example of a feedback mechanism (e.g Ruddiman 2000). The Holocene does not show these large amplitude changes but there is some evidence that also the climate during the Holocene varied in 1500 year cycles and thus could be caused by similar mechanisms as the Dansgaard-Oeschger events (e.g. Alley et al. 2003).

The importance of smaller Arctic ice caps

Ice cores from these smaller ice caps have not been studied to the same extent as the Greenland ice cores largely due to uncertainties about the effect of melt water percolation on these lower elevation glaciers (Koerner 1997). The relatively low-lying ice caps in Svalbard are included in this latter category. However, results from recent Svalbard ice cores have shown that with careful site selection, high-resolution sampling and multiple chemical analyses, it is possible to recover ice cores in which annual signals are preserved (Pohjola et al. 2002). The distribution of species in Svalbard ice cores has probably been altered to some degree by melt but the records still provide information about major trends in atmospheric variability of both climate parameters (Figure 1) and pollution history (Isaksson et al. 2003).

Due to the precise dating, ice cores can be a great help in extending instrumental records both through individual records and in multi-proxy compilations (e.g. Overpeck et al. 1997). One example of such an exercise is the extension of the winter temperature records from Svalbard where one ice core record is used together with instrumental records (Kohler et al., in prep.) (Figure 2).

The low altitude Arctic ice caps provide a great potential for obtaining sea-ice extent proxies on shorter time scale that the high latitude interior ice cores cannot provide. Examples from such studies are from the Canadian Arctic where it was found that the sea salt record from one ice core was linked to the sea-ice extent (Grumet et al. 2001) and Svalbard, where both MSA (methanesulphonic acid) (O'Dwyer et al. 2000) and $\delta^{18}O$ (Isaksson et al. 2003) (Figure 3) seem to be closely linked to the sea-ice extent.

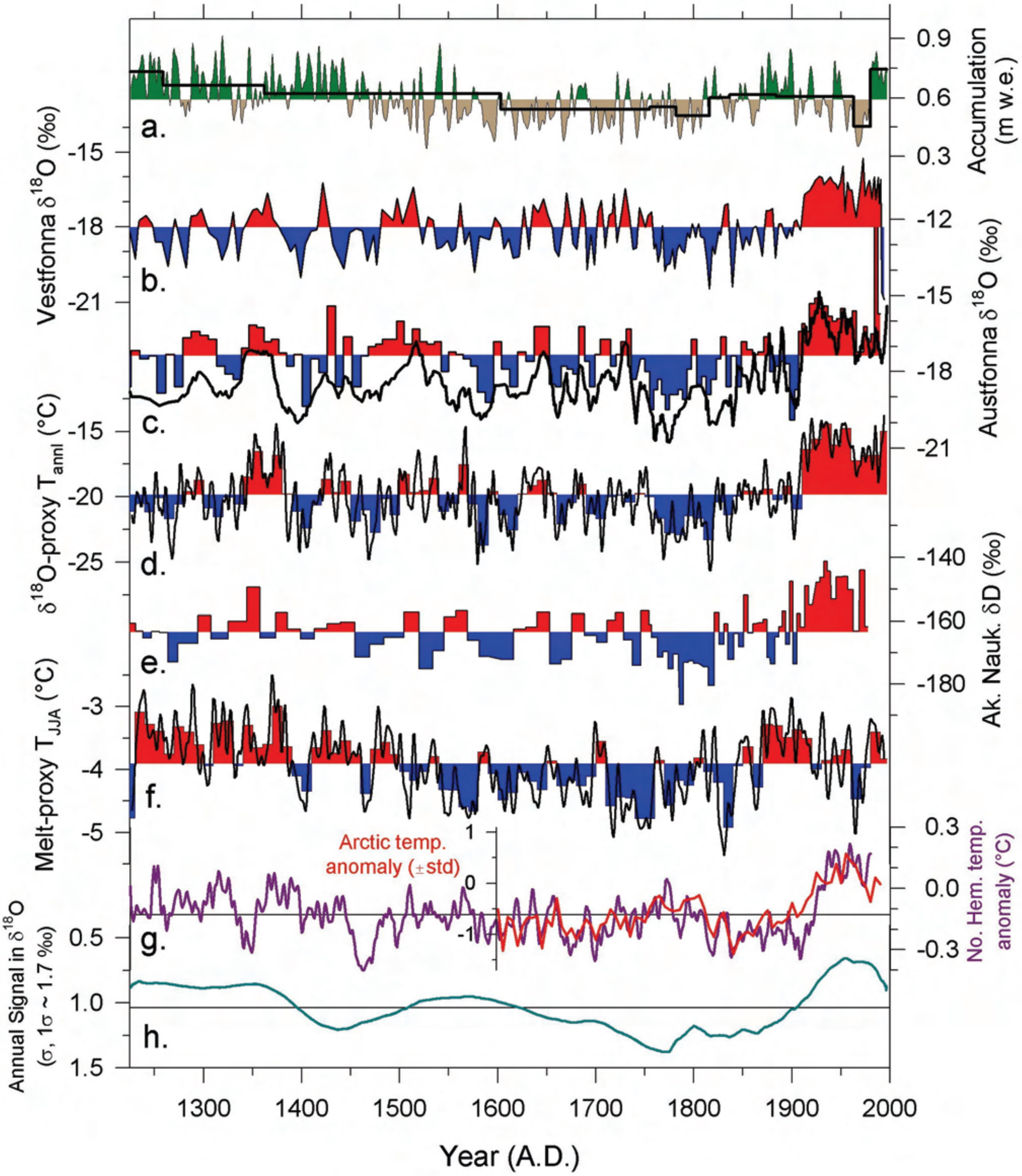

Figure 1. Temperature proxy records from Svalbard (Vestfonna and Austfonna) and Severnaya Zemlya (Akademii Nauk) ice cores (from Henderson 2002).

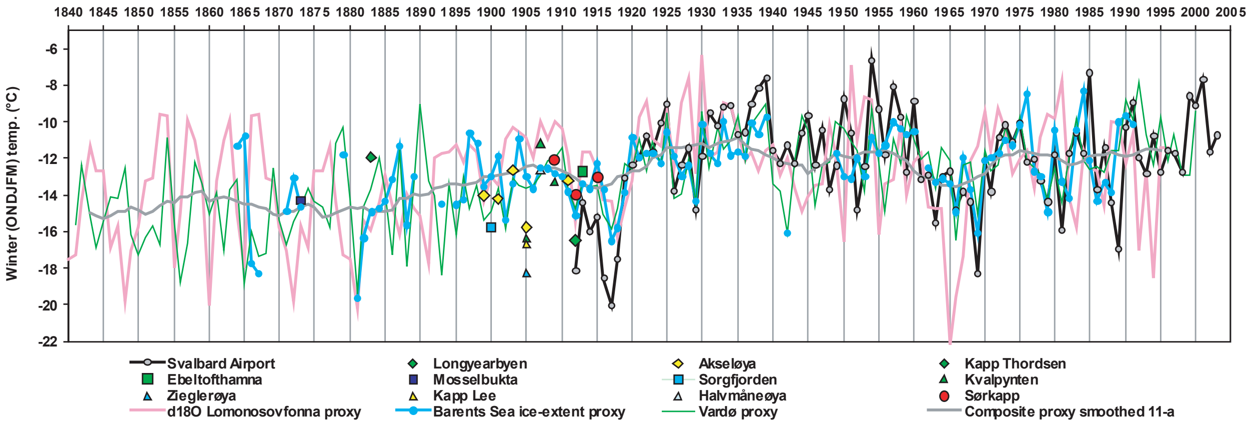

Figure 2. Multi-proxy extension of the winter temperature record from Svalbard using newly digitized expedition data (project funded by ACIA-Norway – J. Kohler and Ø. Nordli), the Lomonosovfonna ice core winter $\delta^{18}O$ (Isaksson et. al. 2003), the Barents April sea-ice extent record (Vinje 2001) and the Vardø and Longyearbyen temperature records (Kohler et al., in prep.).

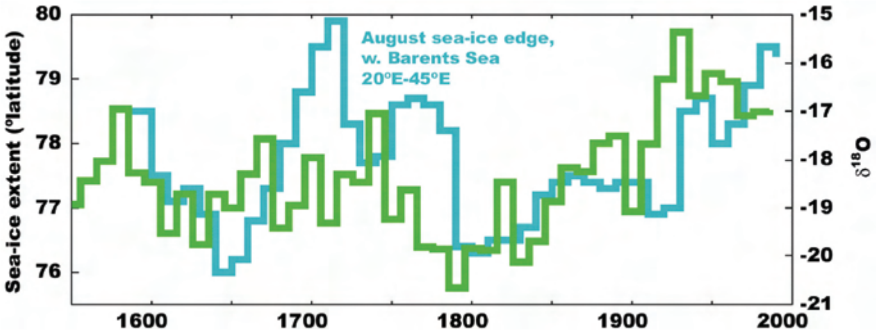

Figure 3. The decadal scale Austfonna $\delta^{18}O$ record corresponds to the August sea-ice record from the western Barents Sea (Vinje 1999). This suggests that $\delta^{18}O$ in the precipitation is directly influenced by the distance to the moisture source in the Austfonna ice core (Isaksson et al. 2003).

Summary and outlook

- Arctic ice cores can provide a wide range of environmental proxies, with accurate dating. The great advantage is that in ice cores many different parameters (i.e. climate proxies) can be measured in the same sample and thus the different processes can be correlated using the same time-scale. The spatial variability of climate is large, so geographical coverage needs to be improved. This can be done using shallow ice cores from a wide range of Arctic ice caps.The fast development of new analytical techniques within ice core research has made it possible to analyse ice cores with high-resolution and precision and thus making it possible to explore changes on short time scales.
- Ice cores are an important component in multi-proxy climate reconstructions and can provide sea-ice extent proxies.
- Ice cores can be an important tool when trying to quantify climate variability on different time-scales.

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Remote Sensing and Modelling of the Arctic Sea-Ice for Climate Studies

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Introduction

Remote sensing and modelling of the ice-ocean system in the Arctic and surrounding seas is important for several reasons: the highest predicted global warming will take place in the Arctic with the consequence that the sea-ice-cover will be reduced; increased human activity is expected due to offshore operations, ship transport and tourism; and the fact that the vulnerable environment requires particular careful management. The Arctic has over the last 2-3 decades experienced a larger temperature increase than other regions of the world, and the ice-cover has decreased in the order of 10% in the same period. Climate models furthermore indicate that anthropogenic global warming will be enhanced in the northern high latitudes due to complex feedback mechanisms in the atmosphere–ocean–ice system. The predicted warming in the Arctic over the next 50 years is in the order of 3-4°C, or more than twice the global average, while the ice-cover is predicated to be reduced by ~40% during summer and ~10% during winter time over the same period. At the end of this century, the Arctic Ocean is predicted to be “a blue ocean” during summer time.

Observing the Arctic sea-ice area and extent has been done successfully over the last two decades using passive microwave satellite data, as shown in Figure 1. These data have shown a significant decrease of the ice area and extent since 1978. However, decadal variabilities can only be detected in longer time series, but these are based on more irregular and scattered data that are less accurate than the passive microwave satellite data. Since observational data on Arctic sea-ice is far from sufficient to estimate variabilities and trends of the ice volume, it is necessary to use coupled ice–ocean models to simulate sea-ice variables on seasonal, decadal and century time scale. These models need to be validated against available data sets such as ice area from passive microwave data, ice thickness and ice motion from drifting buoys and satellite data. Furthermore, the data sets can be assimilated into the models to give optimal results.

Recent advanced sea-ice models

Examples of predicted changes in the Arctic sea-ice during this century are shown in Figure 2, where results from the Bergen Climate Model are presented. These results indicate that the ice area in the summer will decrease significantly, while the winter sea-ice will decrease less. It is noteworthy that most of the Barents Sea will be ice-free also in the winter according to this model. Ice-ocean models also simulate the ice thickness and reduction in ice thickness during this century. Figure 3a shows simulated ice thickness and variability for the period 1850 to 2100 for two of the IPCC scenarios (GHG and GSD) by the Max Planck climate model and the corresponding control runs. Available observed ice thickness time series are included, but the observed data are obtained by different methods and sparse data sets, which makes comparison difficult. Several of the data sets show that there has been a reduction of the ice thickness over the last few decades. The most dramatic results indicated a 40% reduction (Rothrock and Wadhams), while the Nagurny data set showed a decrease that is only a tenth of this value.

Recent advanced satellite developments

Ice thickness data are much more sparse than data on ice area and extent, because satellites have not yet been able to deliver regular ice thickness data. However, methods are

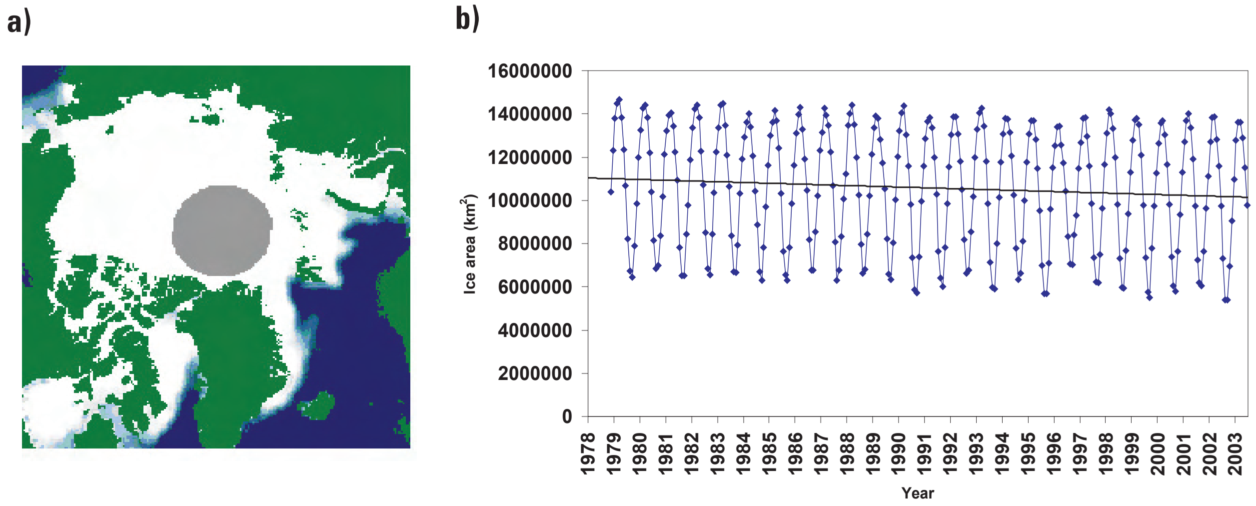

Figure 1. (a) Sea-ice area in the Arctic observed by passive microwave satellite data since 1978. The grey circle around the North Pole is a gap not covered by these data; (b) time series of monthly mean of the Arctic ice area observed by the microwave data, showing the predominant seasonal cycle and the trend indicating a decrease of about 7% of the ice area in the period.

being developed to derive ice thickness from radar altimeter data. The ERS altimeter has produced data covering the Arctic sea-ice up to 81.5°N for about ten years. An example of gridded ice thickness data is shown in Figure 3b. CRYOSAT, a new satellite to be launched by ESA in 2005, will carry an improved radar altimeter providing ice thickness data up to 88°N. Considerable validation work will be part of the CryoSat programme in order to establish the accuracy of ice thickness retrieved from the radar altimeter data.

To observe sea-ice processes on regional and local scale, Synthetic Aperture Radar (SAR) data have proven a unique capability to identify a number of parameters and processes of importance for climate studies in the Arctic. Ice volume fluxes through the Fram Strait are an important component of the Arctic freshwater flux, which is presently observed by an array of moorings with Upward-Looking Sonars (ULS). The ULSs measure ice drift, ice concentration and ice thickness at fixed positions. If wideswath SAR images, which cover the whole ice-covered part of the strait, are obtained at regular intervals (for example every three days), the area fluxes can be calculated by interpolation of the ice drift vectors retrieved by recognizing ice features in a sequence of SAR images. RADARSAT ScanSAR and ENVISAT ASAR wideswath images have been acquired and analysed for the Fram Strait and the area surrounding the Svalbard archipelago. An example of ice drift calculation in the Fram Strait from 24-27 October 2003 is shown in Figure 3a. By combining ice area fluxes from SAR data with point measurements from ULS and gridded data from CryoSat, the ice volume flux in the Fram Strait can be determined more accurately.

Ice velocity fields can be retrieved from several types of satellite data – passive microwave, scatterometer and SAR images – where SAR gives the most detailed resolution of the drift field, tentatively at a grid size of 5-10 km. These fields will be an important supplement to the Arctic Ocean Buoy Programme (AOBP), and the joint data sets from satellites and buoys can be used in ice model validation.

SAR data obtained on a regular basis will also be used to retrieve several other sea-ice parameters (ice edge, ice type classification, floe size, leads, polynyas, ridges and surface roughness), using the different SAR data available from RADARSAT and ENVISAT. In Figure 3b an example of ice classification from SAR is presented, showing the identification of multiyear, first-year and open water/thin ice in the area north of Svalbard. When SAR derived ice drift and ice type classifica-

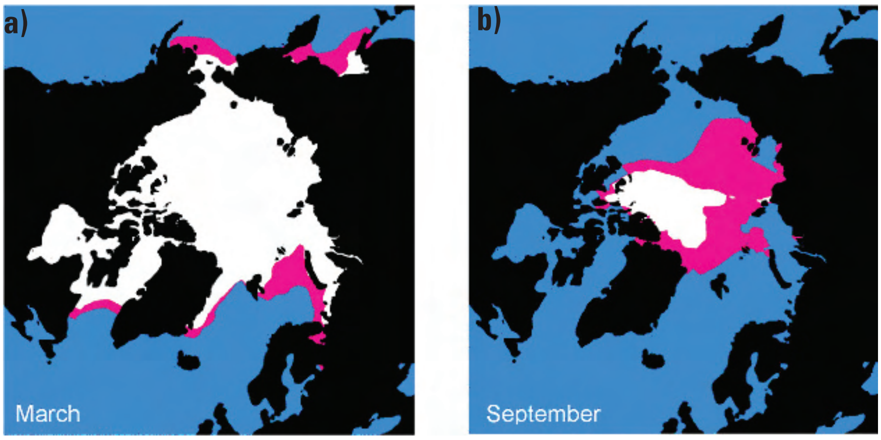

Figure 2. Predicted ice area towards the end of this century by the Bergen Climate Model, based on the IPPC scenario of doubling of CO_2 emission to the atmosphere, for the month of March (a) and September (b), respectively. The white area is the results of doubled CO_2, while the red area is the control run. Courtesy H. Drange, NERSC.

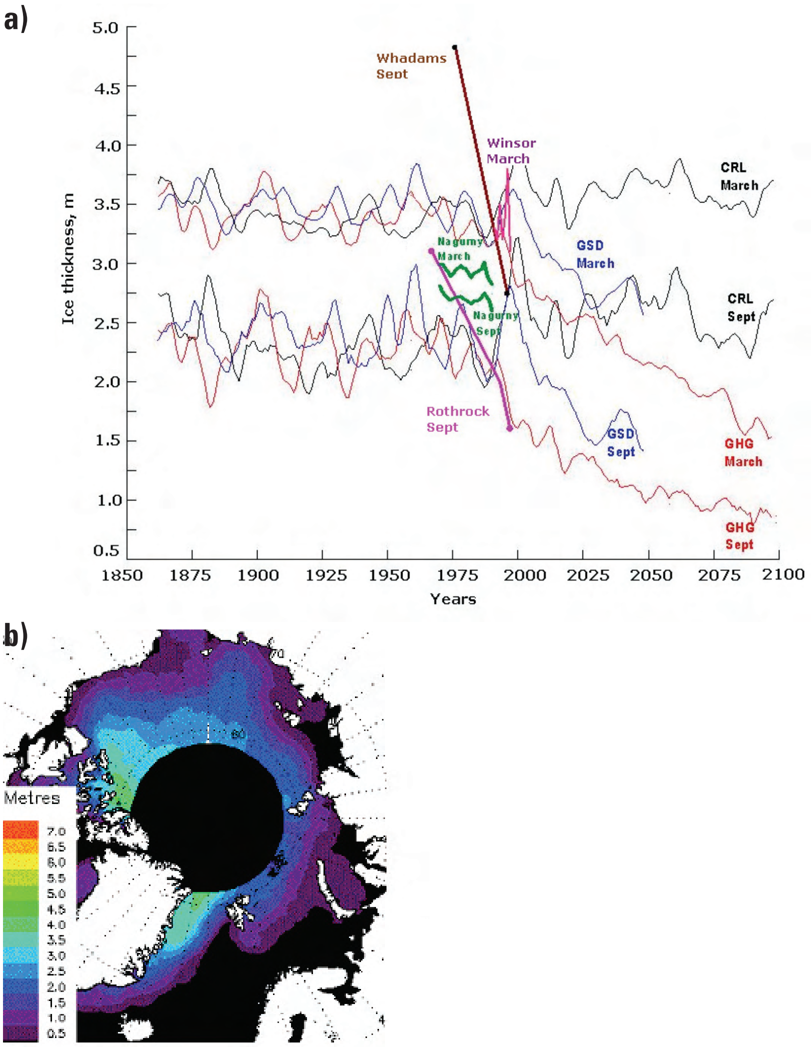

Figure 3. (a) Simulated ice thickness for the Arctic Ocean by the ECHAM model at Max Planck Institute of Meteorology based on the IPCC scenarios of global warming (GHG and GSD) as well as control run for March and September (Courtesy L. Bengtsson). The shorter time series indicate observed ice thickness data published by Wadhams, Winsor, Nagurny and Rothrock. (b) example of a map of mean ice thickness for the Arctic up to 81.5° N derived from ERS altimeter data for 1992-2001 (Courtesy S. Laxon).

tion are combined, it is possible to quantify the seasonal growth and variability of leads and coastal polynyas, including estimation of vertical heat flux and ice formation.

Operational monitoring of sea-ice by satellites

Operational monitoring of sea-ice by satellites is an important part of the monitoring systems presently under implementation in the context of the Global Monitoring for Environment and Security (GMES), where a number of institutions from many countries are involved. One GMES project with focus on sea-ice is ICEMON, which is one of the ESA service consolidation actions in 2003 – 2004 (http://www.icemon.org). ICEMON is preparing for the implementation of a coherent network of monitoring services for the high latitudes, including sea-ice, meteorological and oceanographic services (met–ocean services). ICEMON will serve operational users such as ships, offshore operators and others who need near real time information, as well as climate users who need longer time series of measurements for monitoring and modelling of seasonal, interannual and decadal variability of sea-ice and other met–ocean parameters.

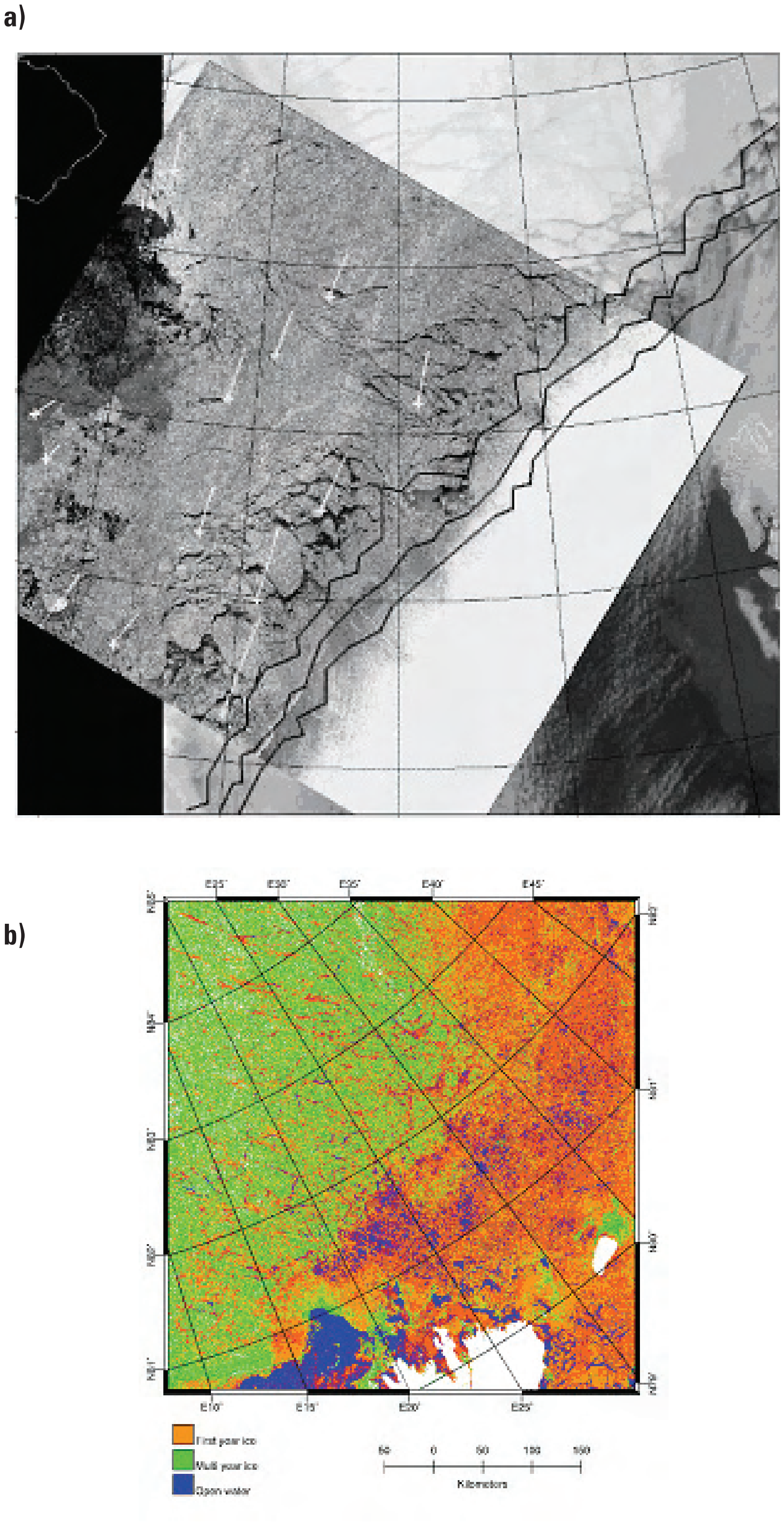

Figure 4. (a) ASAR Wideswath image from 24 October 2003, overlaid by an AVHRR image and SSM/I-derived ice concentrations isolines for 20%, 50% and 80%. The drift vectors represent the period 24-27 October; (b) Ice classification from RADRASAT ScanSAR image northeast of Svalbard 27 January 1997.

Arctic Ocean Shelf-Basin Interaction

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Introduction

The Arctic Ocean Basin is surrounded by large shelf areas, namely the Canadian Archipelago, and the Chuckchi, the East Siberian, the Laptev, the Kara and the Barents Seas. All of these shelves have a profound influence on the characteristics of the passing water masses and its constituents.

In the shelf areas of the Barents and Kara Sea (Figure 1) the inflowing water of Atlantic origin, runoff and polar water are transformed into water masses of different density classes that leave towards the adjacent seas, also feeding the mid-depth and deep central Arctic Ocean. While the local conditions in the Barents and Kara Seas are subject to influences from the atmosphere and farfield conditions of ice and ocean, the water masses which leave the shelves in turn have impact on the large scale at mixed layer, halocline or Atlantic layer levels. Part of this water supplies the outflow and overflow into the Sub-Arctic regions, forming the northern limb of the meridional overturning circulation. The interaction between large-scale and regional-scale in the communication between the Arctic Ocean shelves and basins is subject to large interannual variability.

We discuss some recent results from analyses of numerical model experiments and observations with respect to this variability and try to discriminate between the different responsible factors.

The Barents Sea

In the Barents Sea some areas are frequently occupied by large polynyas. Two of these are known to be active in transforming considerable amounts of surface water into high density water by heat loss and ice formation: the Storfjord polynya of Spitsbergen and the western Novaya Zemlya polynya (e.g. Martin & Cavalieri 1989). The Storfjord polynya has been subject to numerous research cruises in recent years, aiming at a quantification of the dense water outflow from the fjord as a contribution to the mid-depth and deep water of the Norwegian Sea, and subsequently to the Arctic Ocean. The maximum densities as well as the estimates on the dense water production intensities are variable on an interannual timescale (Haarpaintner et al. 2001). The years after 1999 have been characterized by high salinities in the Storfjord bottom water compared to the early 1990s (Schauer, pers. comm.). In summer 2002 the observed maximum salinities were on a record high (35.83). In the following winter, however, namely on the expedition WARPS (Winter Arctic Polynya Study) with RV POLARSTERN in March/April 2003, only low salinities (< 35.0) were found in the fjord.

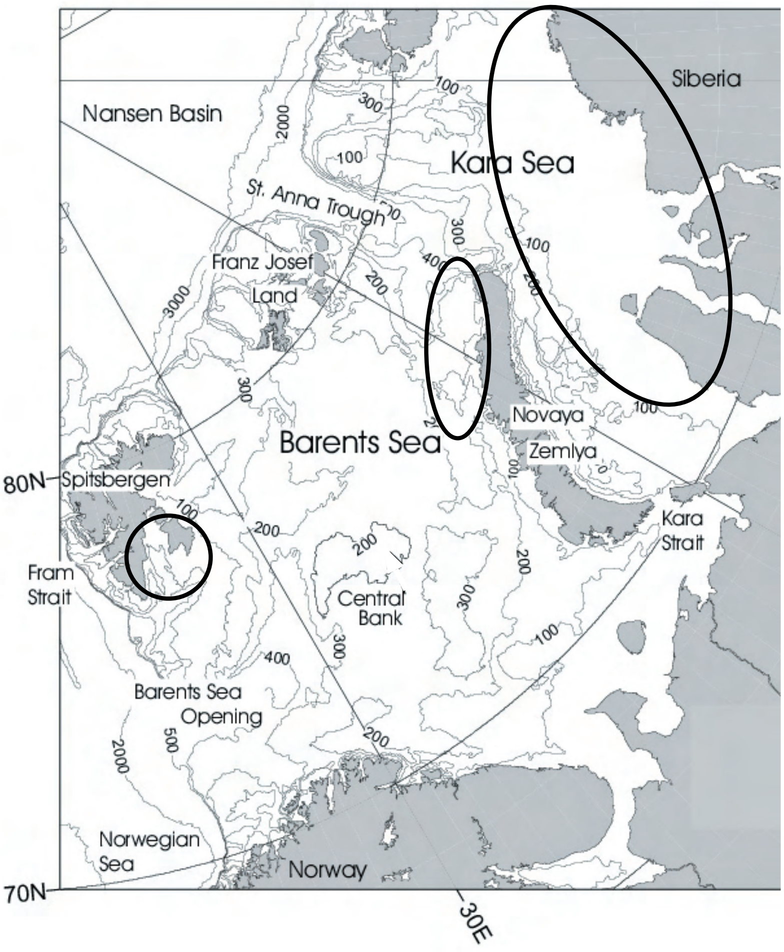

Figure1. Topography of the Barents and Kara Seas shelf. Areas of interest are encircled in black.

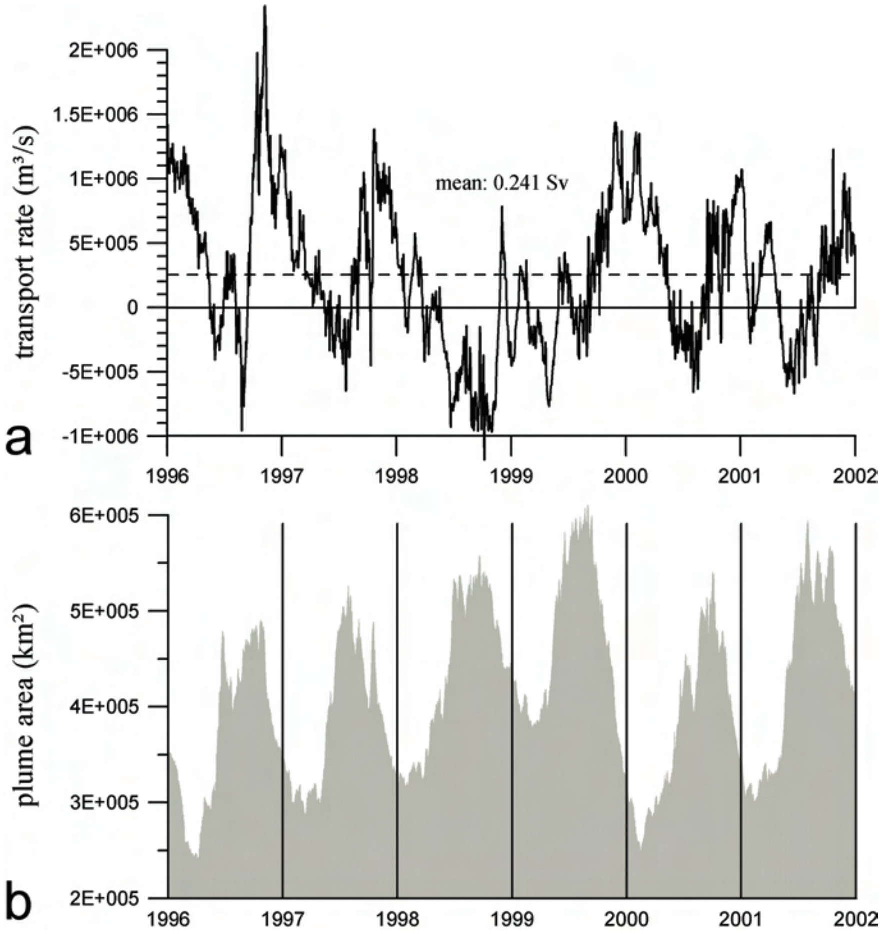

Figure 2. Simulated salinity in the northern Storfjord from 1980 to 2002 (monthly means).

To further investigate the interannual variability of the Storfjord bottom water salinities and the reasons leading to the large change between summer 2002 and winter 2003 the results of numerical simulation with the coupled ice-ocean model NAOSIM were inspected (see Gerdes et al. 2003 and Karcher et al. 2003 for details on the model). The model was forced with NCEP atmospheric reanalysis data for the period 1948 to 2002. It covers the Arctic Ocean, the Nordic Seas and the northern North Atlantic south to approx. 50° N.

The simulated salinities in Storfjord between 1980 and 2002 (Figure 2) exhibit a large seasonal cycle. Maximum salinities are reached in late winter, in late summer/early winter the salinities are minimal. As in the observed data, the period after 1999 shows a marked increase of late winter salinities, reaching the highest value in late winter 2002. Afterwards, however, salinities decrease to the lowest values of the last two decades in December 2002. The salinities in Storfjord in late autumn/early winter set the precondition for the onset of dense water formation in the course of the winter.

According to the model results, the source of the simulated low salinities in the second half of 2002 is evidently the intrusion of low saline water from the northeast. This low salinity water was released by an anomalously large amount of ice melt east of Spitsbergen in summer 2002, which subsequently advected into Storfjord and hampered the production of salty bottom water there. Obviously the local process of deep-water production in the fjord is not only dependent on the local atmospheric and oceanic conditions, but also on farfield influences like the intrusion of freshwater.

A well-known case of far-field influences on the conditions in the Barents and Kara Seas is the variability of inflowing Atlantic water in terms of volume, heat and salt transports. In the early 1990s, for example, an anomalously large amount of heat entered the Barents Sea as a result of a stronger atmospheric driven inflow from the Nordic Sea and a reduced heat loss to the atmosphere in the high NAO phase of this period (Karcher et al. 2003). Part of this anomalous heat survived the strong surface heat loss in the Barents Sea and was carried into the central Arctic, enhancing the warming which occurred in the Atlantic water layer at mid depth due to similarly anomalous heat input from the Fram Strait inflow. Another strong warm event entered the central Arctic in the 1960s during a low NAO phase in which the Barents Sea heat loss to the atmosphere was hampered by extensive ice import from the North. The same large net ice import and its subsequent melt lead to an unusually light Barents Sea water running off the shelf to feed the Atlantic Water layer of the central Arctic Ocean (Karcher et al. 2003b). The characteristics of the densest part of this Barents Sea water feeding the Arctic Ocean below the halocline, are determined by the conditions in the eastern Novaya Zemlya polynya. Its basic salt supply is delivered by the southeastern branch of Atlantic water crossing the Barents Sea shelf and strongly modulated by the intensity of local ice formation/salt release in the polynya (Karcher et al. 2003b, Hatten & Schrum, pers. comm.).

The Kara Sea

The pathways and intensities of the different Atlantic water branches in the Barents Sea, as well as the local ice formation rates are strongly influenced by the local wind stress pattern. In the early 1960s and late 1990s a large positive SLP anomaly covered the eastern Barents and Kara Seas (Karcher et al. 2003c). In the late 1990s it was strong enough to reverse the flow from the Barents Sea through the Kara Strait, one of the main driving forces for the Kara Sea. Results from a regional Kara Sea model for the winter 1998/99 (Harms et al. 2003) show Kara Strait trough flow rates, which usually are in the range of 0.5-1 Sv eastward into the Kara Sea, switch to a net westward flow of 0.5 Sv towards the Barents Sea for nearly the entire winter season (Figure 3a). The SLP anomaly also affected the spreading of river runoff in the Kara Sea, which receives more than one third of the total Arctic river runoff. Prevailing northeasterly winds caused an accumulation of the freshwater in central parts of the sea (Figure 3b). The freshwater export rates towards the north and east, normally at a maximum during winter, were largely suppressed in -98 and -99 and recovered only in 2000 and 2001 with strong freshwater pulses eastward into the Laptev Sea and northwards into the central Arctic Ocean. These strong freshwater pulses were also supplied by an increased amount of ice melt, as ice export to the north was also strongly reduced during the blocking situation in winter 1998/99.

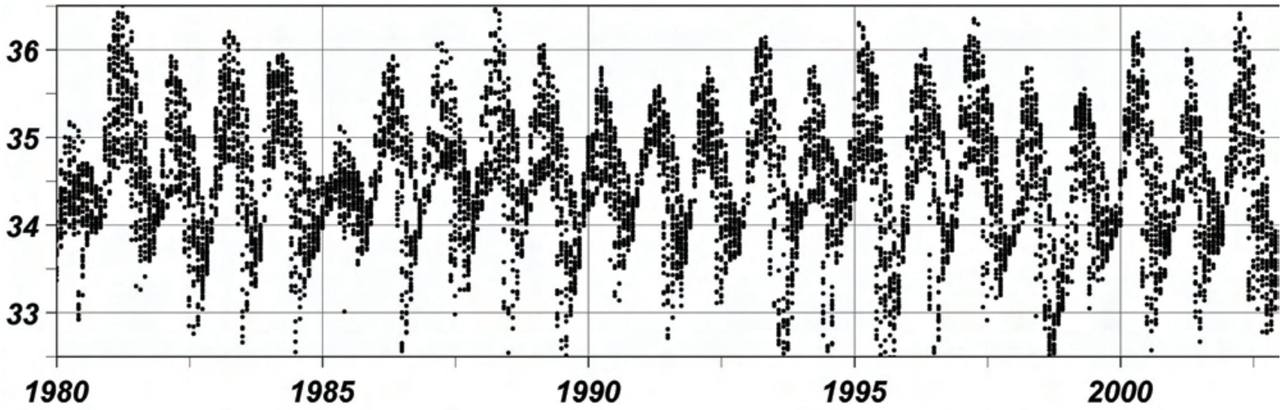

Figure 3. (a) Simulated volume flux through the Kara Strait (positive values denote inflow, negative outflow) and (b) freshwater plume extent (sal < 30 psu) in the Kara Sea.

Conclusions

The Barents and Kara Sead interact with the adjacent areas in terms of exchange of ice and water. The characteristics of the inflowing water are largely determined by farfield conditioning. These inflowing water masses, together with the atmospheric state, influence the local water mass properties on the shelves. They also influence the characteristics of the water masses that leave the shelves towards the central Arctic Ocean.

Model experiments support the observational evidence of dense water production in the Storfjord of Spitsbergen and west of Novaya Zemlya. In both areas the shelf-scale ice import/export balance seems to play a role for the final product leaving the shelf, which is as important as the local ice production or the variability of source water supply.

Furthermore, the experiments hint at a storage capacity and a switchgear role with respect to pathways of river runoff. An example was found in the late 1990s, when a strong atmospheric pressure anomaly lead to partially reversed ice drift and oceanic flow in the Kara Sea and the eastern Barents Sea. This lead to a blocking not only of the Atlantic water trough flow through the Kara Sea, but a complete shutdown of freshwater and ice export for one season with consequences for the large scale freshwater distribution.

The model experiments underline the role of the regional scale processes occurring in the Barents and Kara Seas in conditioning of the Arctic water masses, which subsequently feed into the Sub-Arctic seas.

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Simulating Climate Variability with a Coupled Regional Atmosphere–Ocean–Sea-Ice Model

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The Arctic Ocean and the Nordic Seas

The Arctic Ocean and the Nordic Seas are among the key regions for climate, as this area shows a high sensitivity to climate change (e.g. anthropogenic greenhouse warming). Furthermore, North Atlantic deep water, one of the main constituents of the deep oceans, is formed here. Beside this, strongly nonlinear processes (like e.g. sea-ice or deep convection) are essential and make climate research in this region more challenging (and interesting).

A coupled regional atmosphere–ocean–sea-ice model

In the following the results from a coupled regional atmosphere–ocean–sea-ice model are presented. The model consists of the regional atmosphere model REMO and the ocean model MPI-OM. REMO was originally developed by the German Weather Service and is used here with the physics package from ECHAM, which enables the use of the model for climate purposes. It has a rotated spherical grid and a resolution of approx. 55 km. The model area includes the entire Arctic Ocean, the Nordic Seas and a large fraction of the North Atlantic. The ocean model is formally global, but through the application of conformal mapping regionally variable resolution is achieved. We located the poles in Western Canada and near Moscow and projected the coordinate lines onto the plane that touches the globe at the intermediate point. The grid is thus focused on the Nordic Seas and the adjacent areas (see Figure1), where some of the climatically most interesting processes in the ocean take place. The model includes a Hibler-type sea-ice model and has a free surface, enabling use of a mass flux boundary condition for salinity. Inside the coupling area (identical to the domain of REMO) the ocean is driven with fluxes from the atmosphere model and river runoff, whereas the atmosphere sees SST and sea-ice parameters calculated by the ocean model. The coupling is updated every six hours.

Application of the model

The model was driven with forcing data from the NCEP Reanalyses for the years 1958 to 2002 (lateral boundary conditions for REMO, and input for bulk formula forcing outside the coupling domain for the ocean model). An ensemble of four members with almost identical starting conditions was carried out. The boundary forcing was identical in all runs. The setup allows us to investigate the question of which part of the regional climate variability is forced by global signals (represented here by the boundary conditions at the model domain) and which part is regionally generated (represented by the differences between the different ensemble members). Whereas large scale climate indices like e.g. the NAO index closely resemble the observations in all ensemble members, the sea-ice export through the Fram Strait shows a substantially different behaviour: The large-scale atmospheric circulation seems to modulate the ensemble mean, but the different members of the ensemble show a substantial spread. Here regional processes obviously are of equal importance.

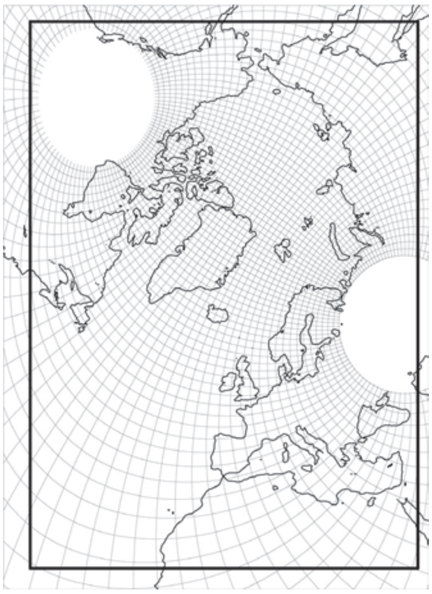

Figure 1. Configuration of the coupled model region. The rectangle is the domain of REMO with a resolution of 100 km. The curved lines represent the coordinates of the ocean model (every 4th line in each direction is drawn). Outside REMO the ocean is forced by NCEP reanalysis data that also serve as boundary conditions for REMO.

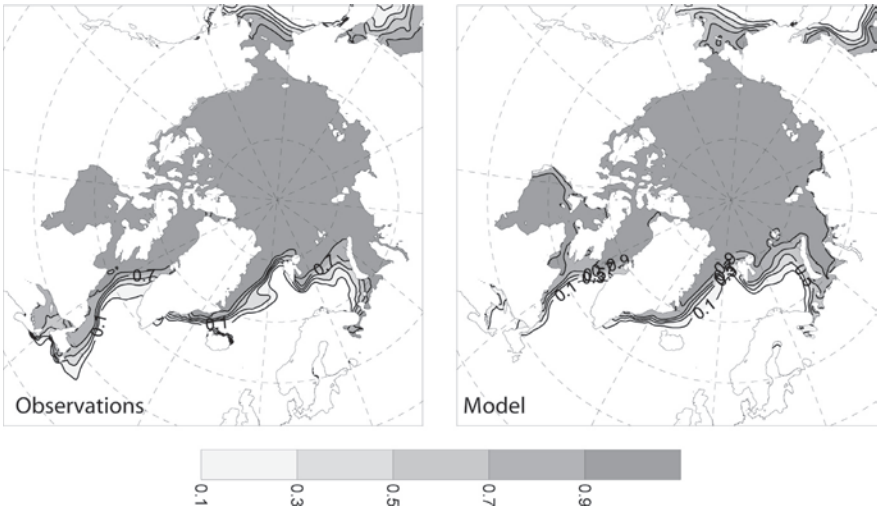

Figure 2. Climatological winter sea-ice concentration (DJF averaged 1978 - 2001) from the coupled model (a) and from satellite observations (b).

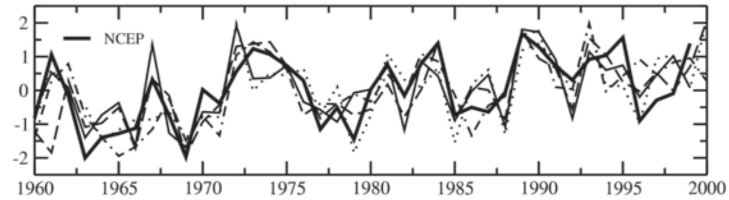

Figure 3. North Atlantic Oscillation index (DJF pressure difference Iceland-Azores) from the four runs and from NCEP reanalyses. The NAO index obviously is strongly determined from the NCEP boundary conditions.

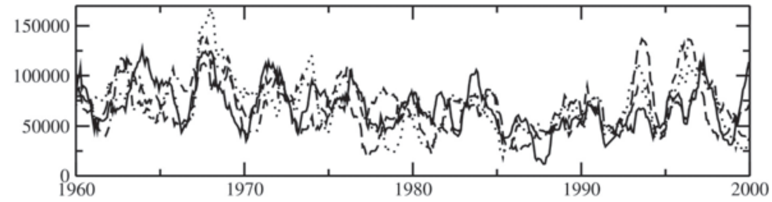

Figure 4. 12 months running mean of ice export through the Fram Strait (m3/s).

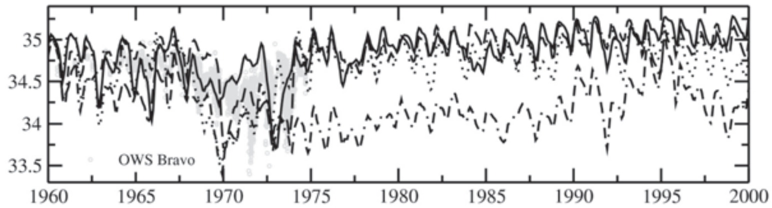

Figure 5. Monthly mean of salinity (22m-62m) in the Labrador Sea and data from OWS Bravo (solid line). All experiments exhibit strong reduction in the late 60s. In three experiments the salinity recovers soon in agreement with the data; in one experiment the low salinity persists over decades.

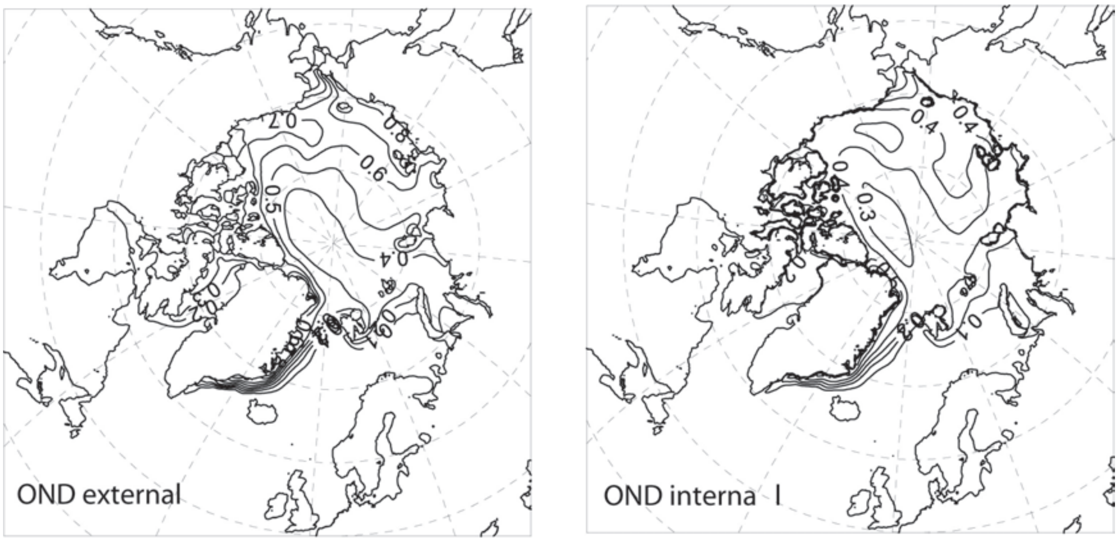

Figure 6. RMS (m) of the ensemble average of ice-coverage OND (a) and RMS of the different runs with respect to the mean (b). The former may be attributed to the variability of the NCEP forcing (outside REMO); the latter is clearly produced inside the REMO area by internal dynamics of the coupled model.

Seasonal and Long-Term Sea Level Variability in the Arctic Ocean

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Introduction

Many forces in the atmosphere and ocean influence the sea-level (SL). By integrating a number of physical processes in the environment the SL can serve as a representative parameter for monitoring climate changes in the Arctic. The dynamic component of the SL is connected with the reconstruction of water circulation, which depends on the modification of atmospheric dynamics and thermohaline structure of water masses. An analysis of seasonal and interannual variability of the SL was conducted using a more than 40-years long (1948-1993) time series of mean monthly data at 44 island and coastal stations in the Kara, Laptev, East Siberian and Chukchi Seas. Tide-gauges at most coastal and island stations are not connected with the geodetic network, so the relative values of SL oscillations were measured at these stations. In the analysis of seasonal and interannual variability we first removed the value for each station, calculated over the entire period of observations. Previous analyses of the seasonal and interannual SL variation have been conducted for different stations of the Siberian shelves by Dvorkin et al. (1978), Pavlov & Pavlov (1999), Pavlov (2001) and by Proshutinsky et al. (2001). However, Dvorkin et al. (1978) used observations up to the mid-1970s, so their results did not take into account modern SL trends. Pavlov (1998) and Pavlov & Pavlov (1999) conducted their analysis only for individual stations in the Laptev and East Siberian Seas. Pavlov (2001) did not estimate contribution of the inverted barometer effect to the SL variability. Proshutinsky et al. (2001) did not discuss features of seasonal variability of the SL.

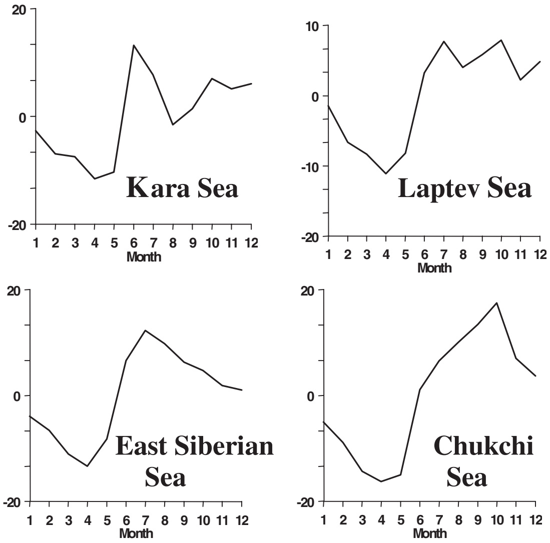

Figure 1. Examples of the seasonal variations of the sea-level (cm) in the Kara (Dikson station: 73. 50N 80.40E), Laptev (Terpiay-Tumsa station: 73.53°N 118.66°E), East Siberian (Chetyrekhstolboviy station: 69.51N 166.58E) and Chukchi (Netten station: 66.95°N 171.94°W) seas.

Seasonal variability

Figure 1 shows typical examples of seasonal variability of the monthly mean SL in the coastal zone of the Siberian marginal seas. At all the stations the minimum values are observed mostly in April. The SL sharply increases from May, reaching its maximum values in summer and autumn. In the Kara Sea the absolute SL maximum is observed in June or July at most stations located along the Siberian coast and especially near river mouths (Figure 1). Some stations have annual absolute maxima in October or December. However, these stations also show local maxima in the summertime. Seasonal variability in the Laptev Sea is nearly the same as in the Kara Sea. Minima are observed in April and maxima in June, July or October depending on the region. In contrast to the Kara Sea, the December maximum is completely absent in the Laptev Sea. The absence of the absolute October SL maximum in the East Siberian Sea is a specific feature of this region. Local maxima were observed at some stations in October. The absolute maximum at all the stations is observed in July. The minimum SL values are observed, as in the previous cases, in April. Stations in the Chukchi Sea show an absolute sea-level maximum in October (Figure 1). Small local extrema were observed at the Netten and Shmidt stations in June and July. Figure 2 shows the geographical variation in the seasonal SL amplitude (Pavlov 2001). Maximum values of the amplitude are observed along the western coast of the Kara Sea, in bays and gulfs of the Laptev Sea, along the eastern coast of the East Siberian Sea and in the western Chukchi Sea. The minimum amplitude of seasonal variability is observed at the island stations in the northern Kara and Laptev Seas. Previous analyses of the SL variations (Dvorkin et al. 1978; Pavlov 1998; Pavlov & Pavlov 1999; Pavlov 2001) have shown that seasonal changes of atmospheric processes in the Arctic and thermohaline conditions in the marginal Arctic seas give a major contribution to the annual SL cycle. In order to estimate the contribution of the SLP changes to the SL variations correlation coefficients were calculated between monthly SL

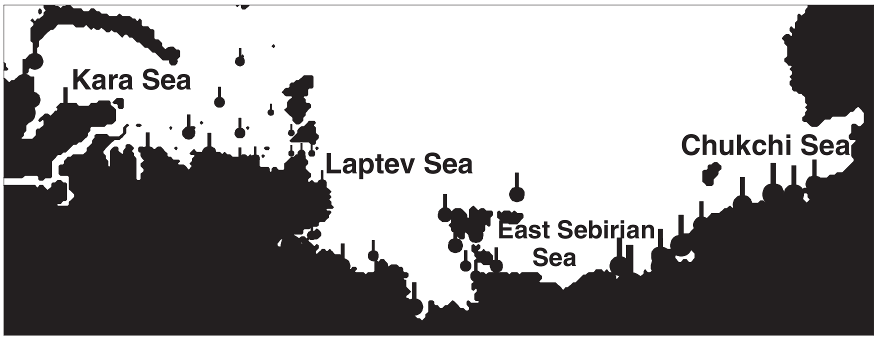

Figure 2. Amplitude (from peak to peak) of the annual cycle of the sea-level (Pavlov 2001).

at the coastal stations and monthly sea-level atmospheric pressure (SLP) for each point (size 5ºx5º) of the NCEP grid over the whole Arctic Ocean. The results of the calculations show that the contribution of the SLP static changes (inverted barometer effect) to the SL variations is high (73% on average) and show no significant difference for the stations in the different marginal seas.

The interesting feature of the correlation coefficients distribution is a location of coefficient maximum. Maximum does not take place at the point of station position but is located far from the station, in a deeper area of the Arctic Ocean (Figure 3). So, variation of the SL at the stations in the Chukchi, East Siberian and Laptev Seas has highest correlation with changes of the SLP over the deep area of the Beaufort Gyre. Monthly SL at the stations in the Kara, Barents and Norwegian Seas highly correlates with the SLP over the deep parts of the corresponding seas (Figure 3).

It was shown by Pavlov & Pavlov (1999) that extreme values of SL in the Kara and Laptev Seas were related to the appearance of a baroclinic coastal current (CC) in summer. This current is caused by large density gradients due to fast ice melting and increased freshwater inflow. Calculations of the monthly mean steric SL at the station near Yenisei Bay, based on annual observations of temperature and salinity which were carried out in 1961, show good agreement with the observed SL for the May-September period (Figure 4). The amplitude (from peak to peak) is close to the observed amplitude at the nearest station, Dikson. This implies that the major cause of the dramatic increase of the SL near the mouths of the big Siberian rivers in June-July is steric expansion resulting from increased river discharge at this time. The main contribution to steric expansion (about 90%) is the reduction of salinity in summertime.

Interannual and inter-decadal variability

Positive trends in interannual sea-level variability at almost all the stations in the coastal zone of the marginal Arctic seas were observed from the beginning of the 1950s until the end of the 1980s (Table 1). The magnitudes of interannual variability of the annual mean sea-level values in the Kara and Laptev Seas are comparable with the amplitudes of the seasonal variation. In the East Siberian and Chukchi Seas, interannual variability is greatest, reaching 40 cm and more. In the three decades spanning the beginning of the 1950s to the end of the 1970s, the positive sea-level trend was not as pronounced as it was in the 1980s.

What is the reason for such a pattern of long-term variability of the sea-level in the marginal Arctic seas? The positive trend of SL increase in the Arctic seas was mentioned earlier by Dvorkin et al. (1978). Analysing data from the most representative stations along the Siberian coast, Dvorkin et al. (1978) explain the increase by suggesting that the Arctic seas' coasts are lowering. This is not supported by the recent results (Pavlov 2001; Proshutinsky et al. 2001) because there are opposite trends at some stations in some months. The trend of SL increase in the Arctic seas can be more correctly explained by changes in the water properties and large-scale circulation of the Arctic Ocean. Shpaykher et al. (1972) proposed that physical processes (including the SL) strongly depend on the circulation of the Arctic Ocean. He reported that due to the dominance of cyclonic circulation above the Arctic Ocean, the SL in the Laptev, East Siberian and Chukchi Seas decrease and the level in the Kara Sea increase. In the years when an anticyclonic character of circulation predominates, there is an opposite situation. However, the analysis of the relationship between the winter North Atlantic Oscillation (NAO) index and the interannual SL

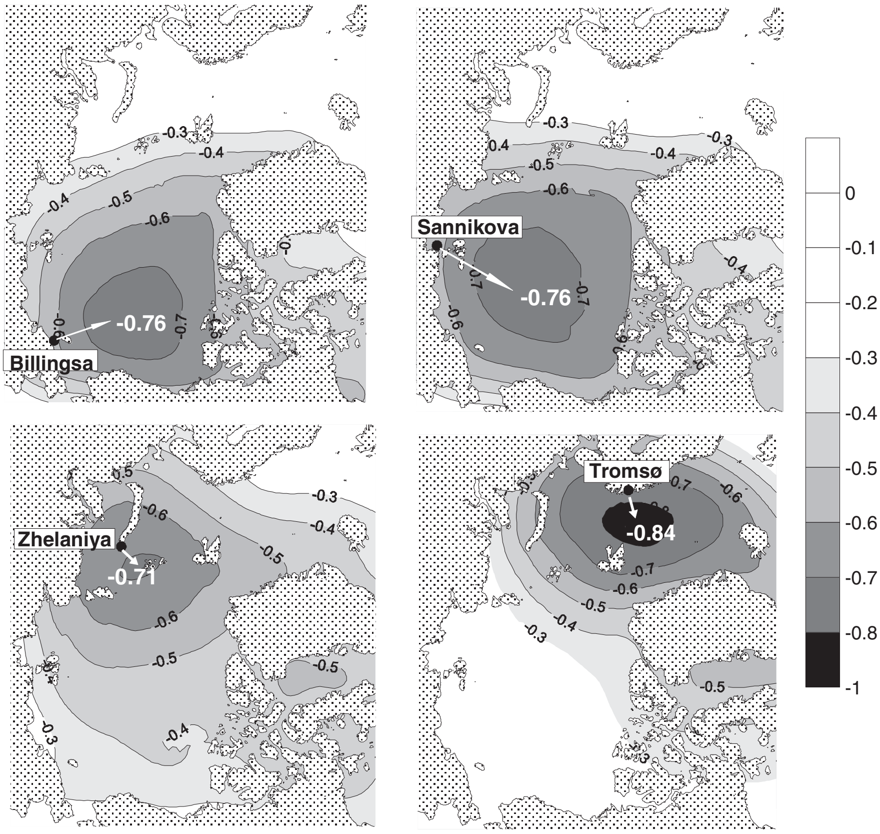

Figure 3. Distribution of correlation coefficient between the monthly mean SLP and monthly SL at the coastal stations. Arrow notes location of coefficient maximum.

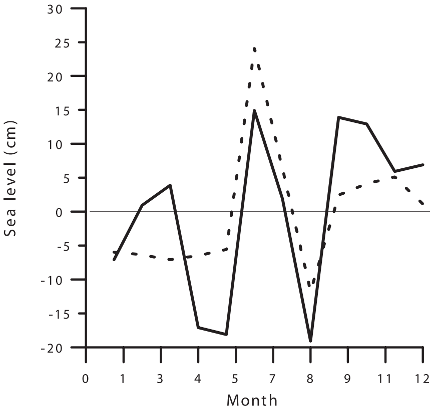

Figure 4. Seasonal variability of steric sea-level in 1961 at the station (73.31°N 80.29°E) near Yenisei Bay (dotted line) and observed monthly mean sea-level during the same year at the Dikson station (73. 50°N 80.40°E) (solid line).

elevations in the Arctic seas has shown that the correlation coefficients are significant, but not high. Recently the contributions of wind and baroclinic components of the circulation to interannual variability of the SL in the Laptev Sea were analysed in Pavlov & Pavlov (1999). It was shown that the baroclinic component dominates. For the East Siberian and Chukchi Seas the same conclusion was reached in Proshutinsky et al. (2001). To estimate the contribution of baroclinic circulation into the long-term variability of the SL of the marginal Arctic seas the simulation of mean decadal fields of level using numerical model (Pavlov & Pavlov 1999) was made. For the diagnostic simulation, mean decadal 3-D fields of water temperature and salinity of the Arctic Ocean for the years 1950-1980 (Joint US-Russian Atlas of the Arctic Ocean, Summer 1998) were used. Figure 5 shows the anomalies of the SL for each of four decades (1950s, 1960s, 1970s and 1980s) obtained from the model. The highest positive anomaly in the 1980s is located in the marginal seas along the Siberian coast. The maximum positive trend (about 8 cm for 40 years) is in the northern parts of the East Siberian and Chukchi Seas. There is a compensating negative anomaly of the SL over the continental slope. The maximum negative trend is located in northern part of the Laptev Sea. Both the observational data and modelling results show that the typical feature of inter-decadal variability of the level is the dramatic increase of the SL in shelf zone of the Kara, Laptev, East Siberian and Chukchi Seas in the 1980s (Figure 6).

Table 1. Linear trends of the interannual sea-level variation (cm per year)

	1950-1979	1950-1989
Kara Sea	0.031	0.111
Laptev Sea	0.043	0.122
East Siberian Sea	-0.219	0.130
Chukchi Sea	0.108	0.418

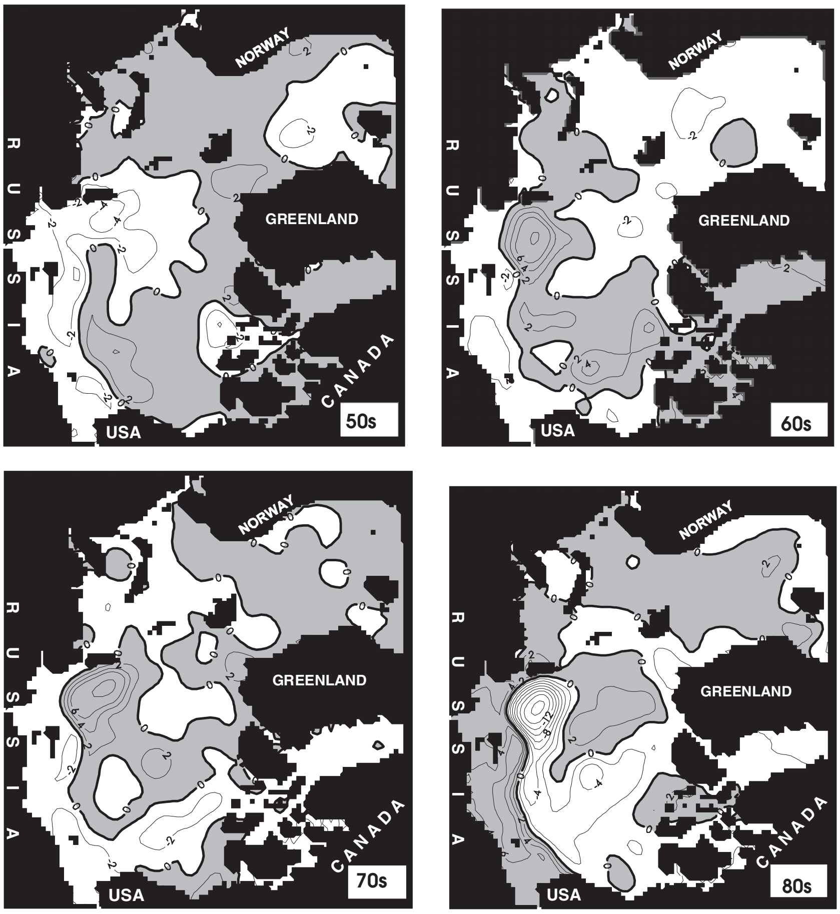

Figure 5. Decadal mean anomalies of the sea-level of the Arctic Ocean (modelling results) for the 1950s, 1960s, 1970s and 1980s. Areas with positive deviation have grey colour (Pavlov 2001).

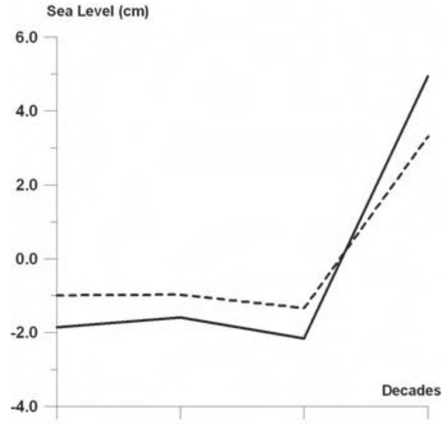

Figure 6. Inter-decadal variability of the sea-level anomalies averaged for stations in the Kara, Laptev, East Siberian and Chukchi seas. Observation: solid line; Model: dotted line.

Conclusions

- The SL in the coastal zone of the marginal seas of the Arctic Ocean has a significant annual cycle. The amplitude (from peak to peak) of the seasonal SL variability is 20-30 cm on average.
- The contribution of the SLP static changes (inverted barometer effect) to monthly mean SL variations is high (73% on average) and does not have a significant difference in the marginal Arctic seas.
- Steric expansion can be one of the major causes of the dramatic increase of the SL near the mouths of the big Siberian rivers in June-July, resulting from increased river discharge and reduction of salinity in this time.
- The agreement between observed decadal mean values of the SL and the results of diagnostic simulations gives grounds to believe that the tendency of the SL rise in the Arctic seas in the 1980s is connected with steric expansion and changes of the large-scale water circulation of the Arctic Ocean.

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Sunlight and Sea-Ice: The Ice Albedo Feedback and a Warming Arctic

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Introduction

The ice-albedo feedback mechanism plays a key role in the heat budget of the Arctic sea-ice-covers (Maykut & Untersteiner 1971; Curry et al. 1995). During the melt season, as incident solar radiation increases and the air temperature warms to the freezing point, the ice-cover evolves from a highly scattering, snow-covered medium to a darker combination of bare ice, melt ponds and leads. The ice-albedo is a positive feedback and is of great import to climate studies. A quantitative understanding of this feedback is needed to better treat the Arctic sea-ice-cover in climate modelling.

The albedo depends strongly on the surface conditions and exhibits considerable spatial and temporal variability during the melt season (Grenfell & Maykut 1977; Grenfell & Perovich 1984; Buckley & Trodahl 1987; Perovich 1996; Perovich et al. 1998; Radionov et al. 1997; Perovich et al. 2002). There has been significant recent progress in defining the key elements of the ice-albedo feedback, in quantifying the feedback, and incorporating improved treatments of the feedback into general circulation models (Curry et al. 2001). Recent research has determined that the ice-albedo feedback is strongly influenced by the timing of seasonal transitions and the duration of summer melt (Perovich et al. 2002). For Arctic sea-ice the development of melt ponds plays a critical role. It is often assumed that a warming climate would mean a longer melt season, with an earlier onset of summer melt and a later freeze up, more ponded sea-ice, and a stronger feedback. These changes could be incorporated into the existing theoretical framework in a straightforward way. It is possible, however, that the changes will be revolutionary, rather than evolutionary. Warming may be so great that a fundamental change in the nature of the sea-ice-cover will result, causing a profound change in the ice-albedo feedback.

Results

From a climate modelling perspective, it is of interest to know the temporal evolution of albedo on the scale of a single grid cell. This entails understanding how an ensemble of leads, melt ponds and bare ice evolves not only optically, but physically. The albedo can not be measured directly from the surface at this scale, and satellite observations are limited by the omnipresent summer cloud cover. Results from the SHEBA field experiment (Perovich et al. 1999; Uttal et al. 2002) include a summer-long time series (in 1998) of the albedo of individual ice types (Pegau & Paulson 2002; Perovich et al. 2002a), as well as the fractional areas of ice, ponds and leads averaged over 100s of km2 (Perovich et al. 2002b). The time series of areally-averaged albedo can be determined from these observations using the relationship

$$\overline{\alpha}(t) = \alpha_s(t)A_s(t) + \alpha(t)_i A_i(t) + \alpha_p(t)A_p(t) + \alpha_w(t)A_w(t)$$

(1) where $\overline{\alpha}$ is the spatially-averaged albedo, α is the albedo, A is the area fraction, t is time, and the subscripts denote snow (s), bare ice (i), ponds (p), and open water (w). The time series of spatially averaged albedo determined using Equation 1 and the SHEBA data is presented in Figure 1.

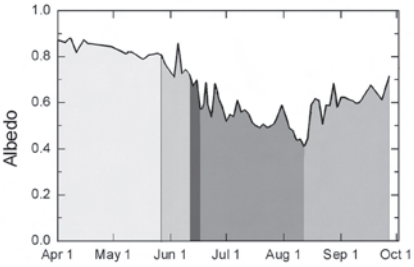

Figure 1. The evolution of spatially-averaged albedo of multiyear ice measured during the summer of 1998 at Ice Station SHEBA. The five phases are snow warming, snowmelt, pond formation, pond evolution and fall freeze up.

The albedo time series has an overall low frequency trend associated with the seasonal cycle, coupled with some high frequency fluctuations due to synoptic weather events. The seasonal cycle can be considered a “wave” with five distinct phases: dry snow, melting snow, pond formation, pond evolution and fall freeze up. In the first phase (April-May), the albedo was large (0.8-0.9), since the surface was cold snow-covered ice. During this time there was a slight and gradual decrease in albedo as the snow cover warmed and the snow grain size increased. Rain on 29 May initiated the second phase of melting snow, causing rapid coarsening of the snow grains to about 1 mm diameter, resulting in a drop in the average albedo from 0.8 to 0.7. Phase 3 started in mid-June, when melt pond formation resulted in a sharp drop in albedo from 0.7 to 0.5 in only a week. After this, there was a long period of a slow, steady decline in albedo as the melt ponds grew deeper and larger in spatial extent. By early August, the average albedo decreased to only 0.4. By mid-August air temperatures were usually below freezing and the fifth and final phase began: fall freeze up. Ice skims formed on the surface of ponds and there was occasional snow, resulting in an increase of 0.1 in average albedo. By the end of August, surface temperatures were consistently below freezing and the albedo increased as the snow deepened. Albedos returned to the springtime maximum of 0.8 to 0.9 by the end of September.

Recent studies (Curry et al. 2001) indicate that determining the timing of the transitions in the albedo evolution is critical to accurately represent the albedo. While Figure 1 represents only albedos from one location for one year, the overall form of the evolution should hold for the general case of any location and any year. Other scenarios will have different timing, e.g. changes in the onset on melt, or the length of the melt season. These changes will in turn vary the wavelength and the amplitude of the albedo cycle, but the general form will still apply. This should be true even under modest warming. For example, in a warming climate, it is expected that melting will start earlier, last longer and have a smaller

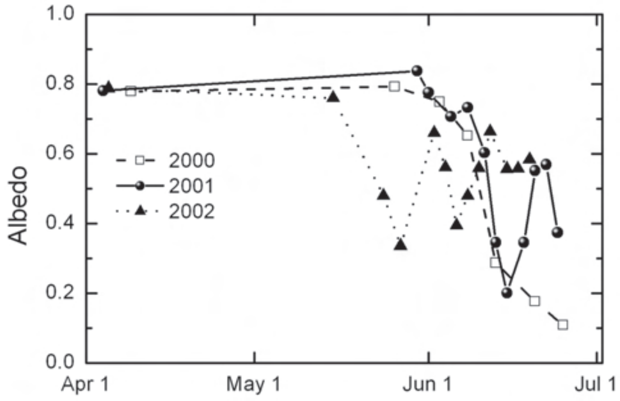

Figure 2. The evolution of spatially averaged albedo of shorefast first year ice at Barrow Alaska as observed in 2000, 2001, and 2002.

albedo minimum.

There may, however, be fundamental changes in the nature of the sea-ice-cover that will cause a profound change in the ice-albedo feedback. There are obvious impacts from a warming climate, such as larger open water fractions resulting in a decrease in albedo and greater heat input to the system. There are also more subtle consequences, such as those resulting from greater amounts of first year ice or changes in winter snow accumulation. As the amount of first year ice increases the general form of the albedo evolution characterized by Figure 1 will become less applicable. Somewhat deformed first year ice may have morphological properties, and an albedo evolution, similar to multiyear ice. In contrast, undeformed first year ice will likely have extensive pond coverage, no surface scattering layer, lower albedos and an accelerated ice-albedo feedback. Indeed recent observations made on undeformed shorefast ice near Barrow Alaska (Grenfell & Perovich 2003) show considerable interannual variability in the albedo evolution of first year ice (Figure 2). The results in Figure 2 are albedos averaged from measurements made over a 200 metre long line. In 2000 the form of albedo evolution was similar to the SHEBA results with a faster and deeper decrease. The 2001 albedo evolution started in a similar fashion, decreasing rapidly. However, the decrease was followed by a sharp increase from 0.2 to 0.6, then another decrease. The behaviour in 2002 was even more complex.

The ice physical properties provide insight into the interannual variability in albedo evolution. Once melting began in 2000, there was widespread flooding of the surface and temperatures above freezing. In contrast, there was flooding at the onset of melt in 2001, but then there was substantial drainage of the surface melt water, greatly reducing the area covered by ponds. Melt began two weeks earlier than usual in 2002. However, after 10 days of melting, air temperatures dropped below freezing for several days, freezing the melt ponds and increasing the albedo. When melt resumed, the frozen ponds melted and the albedo decreased rapidly. As the ice warmed the permeability increased allowing some of the surface melt water to drain, once again increasing the albedo.

The albedo evolution of underformed first year ice is a complex process. This complexity is due, in part, to the lack of surface topography that increases the sensitivity of melt pond development to the snow depth, melt rate, surface topography, and ice permeability. Depending on the degree of deformation, there may be even greater uncertainties. For example, in the active ice pack, ice dynamics will tend to deform the relatively thin first year ice forming ridges and rubble fields. Little is known about the albedo evolution of such highly deformed ice.

Summary

Because of the potentially powerful positive feedback the evolution of sea-ice albedo has potential climate importance. Under current environmental conditions, there are five distinct phases in the annual cycle of multiyear ice albedo evolution. This cycle is reasonably well understood, though questions regarding melt ponds formation and development still remain. For modest changes to the ice-cover the fundamental albedo formulation should still be applicable, with modifications to the wavelength and amplitude of the albedo cycle. However, major changes in ice conditions will bring major changes to albedo evolution. Thinner ice may well lead to an accelerated ice-albedo feedback, but this is more speculation than extrapolation. More work is needed investigating the albedo of both deformed and underformed ice and of the distribution of surface melt water.

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Land Ice (Glaciers and Ice Sheets) in the Climate System

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Land ice occurs as glacier ice, permafrost and seasonal lake and river ice. Climatically driven changes of permafrost, lake and river ice that may have important societal impacts are not treated in this contribution, which just deals with land ice occurring as glaciers.

Glaciers are conventionally divided into two main categories: (1) small glaciers and ice caps (SGIC) and (2) ice sheets, i.e. the Greenland and Antarctic ice sheets. In a climate change context the argument for making this distinction between glacier categories is the magnitude of difference in size of the glaciers, but also the fact that they display very different response times to climate change. Glaciers and ice sheets interact with climate on a wide range of time scales from annual changes of snow cover (albedo-changes) over decadal/century – and even longer scale variations of fresh water exchange with the ocean to millennial scale changes in landscape topography by down-wasting of ice masses and isostatic crustal movement. Several strong positive feedback mechanisms characterize glacier–climate interactions, e.g. melt-rate/albedo feedback, melt-rate/glacier sliding feed back, and mass-balance/surface-elevation feedback.

Arctic glacier area, volume and contribution to sea-level change

In the Arctic, SGICs outside Greenland cover about 275 000 km^2 distributed over the widely glacierized archipelagos of the Canadian, Norwegian and Russian High Arctic, as well as over areas north of about 60°N in Alaska, Iceland and Scandinavia (Dowdeswell at al. 1997). If glaciers in Alaska and neighbouring Canada between 55°N and 60°N (not usually counted as Arctic glaciers) are included, this will add another 90 000 km^2 to the glacierized area within the region of interest for AMAP/ACIA (Arendt et al. 2002). Of the glacierized area in Greenland, the main ice sheet covers about 1 710 000 km^2, whereas SGICs cover roughly 50 000 km^2 (Weidick 1995). Depending on whether the Alaskan glaciers are included or not, the total area covered by Arctic SGICs constitutes 81% or 63% of the total global SGIC-area of about 510 000 km^2. While the Earth's total glacier and ice sheet area is pretty well known, this is not the case for glacier volume. In particular, the total volume of the small glaciers and ice caps is based on estimates and not on observed ice thickness. The sea-level equivalents of the volume of the Greenland ice sheet and Arctic SGICs are about 7.2 m and 0.3–0.4 m respectively, if it is assumed that the average thickness of the Arctic SGICs equals the global average.

Table 1. Area, volume, and mass balance of Arctic glaciers and ice caps, excluding the Greenland ice sheet

Region of the Arctic	Ice-covered area (103 km^2)	Average Mass Balance Dowdeswell (m y-1)	Average Mass Balance Ohmura (m y-1)
Svalbard	37	-0.55	-0.38
Franz Josef Land	14	-0.03	
Novaya Zemlya	24	-0.14	
Severnaya Zemlya	18	-0.03	
Ellesmere Island	81	-0.08	-0.15
Axel Heiberg Island	12	-0.08	
Devon Island	16	-0.08	
Baffin Island	43	-0.23	
Iceland	11	-0.16	
Scandinavia	3	+0.03	-0.05
Alaska north of 60°N	15	-0.21	-0.37
Alaska/Yukon s. of 60°N	90	-0.52*	
Greenland	49	?	
Total	323 (413)		

In accordance with the worldwide trend, most Arctic SGICs showed significant mass loss during the second half of the 20th century (Dowdeswell et al. 1997; Ohmura in press). As regards the Greenland ice sheet, recent investigations indicate that also this ice-mass is presently loosing mass, although it is not yet clear whether the mass loss reflects a long-term trend or whether it reflects short-term temporal variability of snow accumulation and melt-rate (Krabill et al. 2000). The present state of balance of the Antarctic ice sheet is still an open question (Rignot & Thomas 2002). An overview of glacier and ice sheet area, volume and contribution to present global sea-level change is given in Tables 1 and 2.

Mass balance

The processes related to the surface mass balance of glaciers are pretty well understood and climate induced changes of surface mass balance can be modelled with reasonable accuracy. An exception is a gap of knowledge in respect to modelling the processes in the percolation zone, where melting/re-freezing processes dominate. Also our understanding of the processes at the glacier bed (glacier sliding) and glacier calving, and how these processes are influenced by climate change, is at best fragmentary if not completely missing.

No doubt, continued monitoring of glaciers, ice caps and ice sheets for example by using existing and future space-borne laser and radar altimeters (ICESAT/GLAS; CRYOSAT) will improve our knowledge of the current volume change of the ice masses. In order to interpret volume changes derived from such measurements in terms of mass changes – the relevant quantity influencing global sea-level – temporal changes of specific mass balance and near-surface temperature, and their influ-

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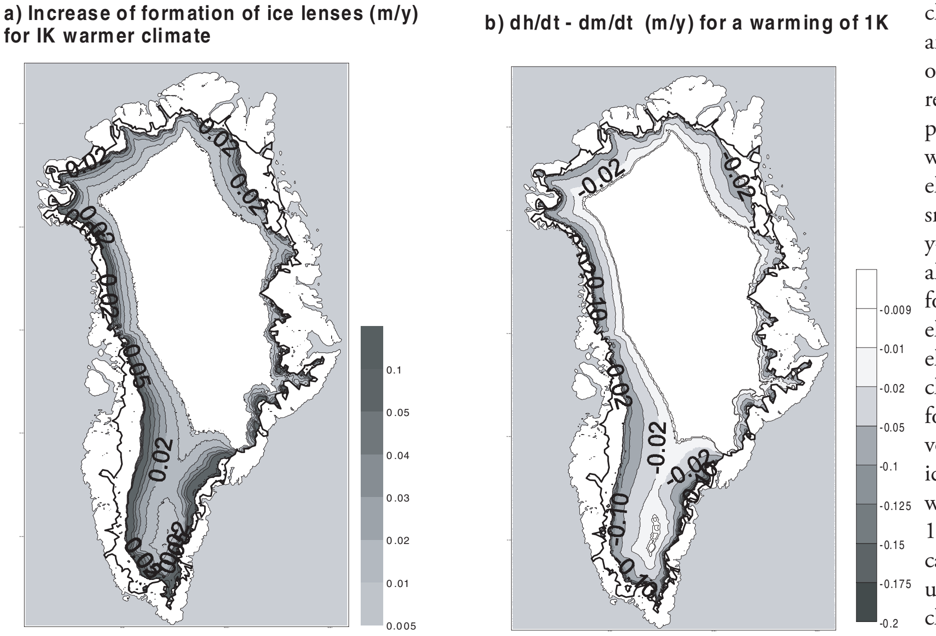
Dowdeswell et al. (1997): based on 20 – 40 years of net balance measurements of selected glaciers
Ohmurah (in press): based on 20 – 49 years of net balance measurements (1952 – 2000) of selected glaciers
*Arendt et al. (2002): based on measured volume change from mid 1950s to mid 1990s of selected glaciers


a) Increase of formation of ice lenses (m/y) for IK warmer climate

b) dh/dt - dm/dt (m/y) for a warming of 1K

Figure 1. (a) Calculated increase of ice lens formation at the surface of the Greenland ice sheet for a 1K warming. (b) Difference between surface elevation change and local mass change (in ice equivalent thickness) for a 1K warming.

ence on surface density and hence glacier and ice-sheet volume change, must be understood and accounted for. Changing snowfall over the accumulation area and changing ice meltrates in the ablation area will result in immediate surface-elevation (volume) changes that can be directly translated into mass changes by using the density of firn or ice. In the accumulation area, temporal variations of surface energy balance and accumulation rate will in addition cause temporal changes of the rate of densification of the near surface layers, giving rise to a surface elevation (volume) change, not accompanied by a corresponding change of mass (Arthern & Wingham 1998; Cuffey 2001). These surface-elevation changes are significant, but small. Much larger surface-elevation changes without a corresponding change of mass may occur in the widespread percolation- and wet snow zones on Arctic glaciers and the Greenland ice sheet. Here, the density–depth profile is to a large extent controlled by melting–refreezing processes, by which low-density surface snow is transformed into high-density ice lenses deeper in the snow pack. A temperature change – or rather a change of the energy balance at the surface – will change the amount of surface melting and the subsequent re-freezing of ice lenses or formation of superimposed ice. The resulting density change will cause an immediate change of surface elevation, the magnitude of which in the wet snow zone may reach values of 0.1–0.2 m y-1. As first pointed out by Braithwaite et al. (1994), this has important consequences for how observed changes of glacier surface elevation should be interpreted. Part of the elevation change, or all of it, may be due to a change of surface-layer density, and therefore, may merely reflect a change of ice-sheet volume without a corresponding change of ice-sheet mass. This is illustrated in Figure 1 which, for a surface-temperature increase of 1 K over the Greenland ice sheet, shows the calculated increase of ice-lens formation (Figure 1a) and the part of the surface elevation change not contributing to the mass change (Figure 1b). For the total Greenland ice sheet, the increased volume loss resulting from a 1K change of surface temperature amounts to 128 km3 y-1, whereas the increased mass loss by runoff only amounts to 96 km3 y-1 of ice equivalent, showing that only 75% of the volume change caused by a warming of 1K represents a change of mass. High-accuracy space-borne measurements of changes of the gravity field (GRACE) may in the future help to discriminate between volume and mass changes of glaciers and ice sheets.

Table 2. Contribution from glaciers and ice sheets to global sea-level change

Region	Area (103 km^2)	Volume (m sea-level equivalent)	Sea Level Change (mm y-1)	Period	Ref.
Arctic glaciers excluding Greenland	273		0.13	1950 - 1995	1
Alaska/Canada	90		0.14 ± 0.04 0.27 ± 0.10	1955 -1995 1995 - 2001	2
Small glaciers excluding Greenland	513	0.15	0.4	1970 - 2002	3
			0.7	1990 - 2002	
Small glaciers Total	680	0.5	0.3 ± 0.1	1910 - 1990	4
Greenland ice sheet	1710	7.2	0.05 ± 0.05	1910 - 1990	4
Greenland ice sheet below 2000 m elev.			0.13	1993/94 – 1998/99	5
Antarctica	12370	61.1	-0.1 ± 0.1	1910 -1990	4

1. Dowdeswell et al. (1997) 2. Arendt et al. (2002) 3. Ohmura (in press) 4. IPCC (2001) 5. Krabill et al. (2000)

Recommendations

It is important to stress that although the use of remote sensing measurements for studying changes of the land ice masses will increase in importance in the future, there will still be a large demand also for in situ measurements for calibration/validation purposes. It should also be emphasized that future mass balance changes are strongly dependent on future changes of climate. As a consequence, our ability to predict future mass balance changes of the Arctic ice masses is closely linked to the ability of General Atmosphere and Ocean Circulation Models to predict future changes of Arctic climate on a regional scale.

In order to improve estimates of future mass balance changes, the following studies should also be given high priority: (i) better understanding of albedo changes and feed back mechanisms, (ii) studies of outlet glacier dynamics with emphasis on their potential for triggering persistent, fast changes in glacier/ice sheet mass, (iii) studies of glacier calving dynamics, and (iv) improving ice-dynamic models for determining the long-term response of glaciers and ice sheets to past climate change.

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Monitoring of Climate Change Effects: The AMAP Trends and Effects Monitoring Programme 1998 – 2003

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Introduction

At the Alta Ministerial meeting for the Arctic Environmental Protection Strategy (AEPS) in 1997, the Arctic Monitoring and Assessment Programme (AMAP) was asked to monitor and assess the effects of increased UV-B radiation and climate change on Arctic ecosystems, with special emphasis on human health impacts and combined effects of multiple stressors. At the first Ministerial meeting of the Arctic Council (AC) in 1998, the working group on Conservation of Arctic Flora and Fauna (CAFF) was asked to monitor and assess, in collaboration with AMAP, the effects of climate change and UV radiation on Arctic ecosystems. During 1999, AMAP, CAFF, the International Arctic Science Committee (IASC), the International Panel on Climate Change (IPCC) and the World Climate Research Programme (WCRP) jointly explored the idea of an assessment of the consequences of climate variability and change and the effects of increased UV in the Arctic region. This work led to the formal proposals to the AC to plan and conduct the ACIA.

As a part of the planning for the assessment work two workshops were held, in Rovaniemi in 1998 and in Tromsø in 1999, to document ongoing activities within the Arctic region related to observation and assessment of climate change and UV effects, and to prepare proposals for an observation network and research programme. The scientists were asked to propose parameters that would be of most significance in documenting eventual effects today and in the future. The programme was a first attempt to establish a system that would gather data needed for the ACIA assessment.

Climate and UV in the AMAP Trends and Effects Monitoring Programme

Based on the results and proposals from these workshops, and work performed within AMAP, a 'climate and UV effects component' was approved by the AMAP working group and included in the AMAP Trends and Effects Monitoring Programme for 1998 – 2003. The intention was that countries that were at the stage of initiating climate and UV observations in the Arctic would follow these recommendations, and countries with ongoing activities would attempt to harmonize these with the programme specifications, in an attempt to get comparable data from all parts of the circumpolar Arctic. Table 1 and 2 show part of the recommendations in the AMAP Trends and Effects Programme (www.amap.no). It was hoped that some of the data gathered might be used in the 2004 Arctic Climate Impact Assessment (ACIA), although it was clearly recognized that the time available for collection and processing of data would be very short before that assessment was undertaken. In general, the programme specifications aim to establish a consistent set of observations that can be monitored over the longer term, for use in future climate and UV assessments.

In an attempt to ensure that assessment experts are aware of relevant ongoing scientific activities, countries and scientists were asked to register their programmes and/or projects in the AMAP Project Directory (PD), or the ENVINET PD that is fully integrated with the AMAP PD. A total of 156 projects linked to climate and UV studies have been registered by scientists working in 12 countries, including a few non-Arctic activities (e.g. projects implemented in Alpine areas). We recommend that experts update these two project directories.

Some of the national programmes/projects have been initiated based on the recommendations in the AMAP trends and effects monitoring programme, and some results have been used in the ACIA assessment. The Zackenberg station in Greenland is an example of a station that has implemented a large part of the AMAP recommendations.

There is a continued need to focus on harmonization of climate and effects monitoring efforts, and it is necessary to update this programme so that it also may serve the future work of ACIA.

The Arctic reflects global processes and Arctic people are among the most exposed in the world to a number of contaminants. It is therefore important to:

- Secure an Arctic network for monitoring and research of climate, UV and pollution;
- Study impact of climate and UV/ozone on humans, biota and contaminants;
- Perform effect studies on biota and humans, especially combined effects; and
- Study basic mechanisms linking global and Arctic processes.

Table 1. Monitoring climate change and its effects in the marine environment
(From: AMAP, 1999. AMAP Trends and Effects Programme, AMAP Report 99:7)

Recommended observations for climate change/variability in the marine environment

Observation	Approach	Additional information
Temperature and salinity	Regular observations along standard sections and at fixed locations	Nutrients, current patterns
Position of fronts	Particularly in Barents and Labrador Seas	Current patterns and flux measurements
Sea ice extent, thickness, concentration and mass fluxes	Remote sensing and upward looking sonar	Vessel/aircraft ice observations
Freshwater outflow	Particularly in Norwegian and Greenland Sea area	Current patterns and flux measurements
Other parameters		Sea level (rise), meteorological parameters including winds, carbon and CO_2

Examples of main options for monitoring climate change effects in marine ecosystems

Potential effects on	Species groups	Effects
Biodiversity	Benthos, plankton, fish populations	Changed Species composition
Distribution area	Plankton, commercial fish stocks, marine mammals, sea birds	New species introductions, reduction in existing species
Growth	Fish stocks	Change in maturation timing
Reproduction	Plankton, fish stocks, sea birds	Change in recruitment

Table 2. Climate monitoring at Zackenberg: site of coordinated long-term Bio-, Geo-, Climate- and Marine monitoring
(From: Zero, 2003. 8th Annual Report 2002. Danish Polar Centre)

MarineBasis in relation to AMAP's Effects Monitoring Programme (Climate Change effects)

Variable	Parameter subdivision	MarineBasis Sites	Level	Recommended by AMAP
Radiation	UV-B	2	Micro/Fjord	Yes
	Short-wave		Micro/Fjord	Yes
	PAR		Micro/Fjord	Yes
Air pressure		1	Micro/Fjord	Yes
Wind	Speed	1	Micro/Fjord	Yes
	Direction	1	Micro/Fjord	Yes
Snow cover	Depth	3	Micro/Fjord	Yes
	Extent	3	Outer fjord	Yes
	Duration	3	Micro/Fjord	Yes
Ice-cover	Depth	1	Micro/Fjord	Yes
	Extent	10	Outer and mid fjord	Yes
	Freeze time	1	---	Yes
	Thaw time	1	---	Yes
Current patterns	Vertical profiles	2	Micro/Outer and mid fjord	Yes
	Speed & direction	2	Micro/Outer and mid fjord	Yes
Nutrients	Vertical profiles and fluxes	2	Micro/Outer and mid fjord	Yes
Carbon and CO_2	Vertical profiles and fluxes	2	Micro/Outer and mid fjord	Yes
Salinity and temperature profiles	Vertical profiles	2	Micro/Outer and mid fjord	Yes
Water table, tides	Depth	Irrelevant	---	Yes
Plankton	Species composition, abundance, distribution	2	Micro/Sea ice/water column	Yes
Primary production	Ice algae/phytoplankton/underwater plants	2	Micro/Outer and mid fjord	Yes
Underwater plants/diatoms	Species composition, abundance, distribution	10	Outer and mid fjord	Yes
Benthos	Species composition, abundance, distribution	10	Outer and mid fjord	Yes
Walrus	Abundance	1	Outer and mid fjord	Yes
Ringed seal	Change in food choice and condition	---	Outer and mid fjord	Yes

Observed and Modelled Climate Trends used in the ACIA Project

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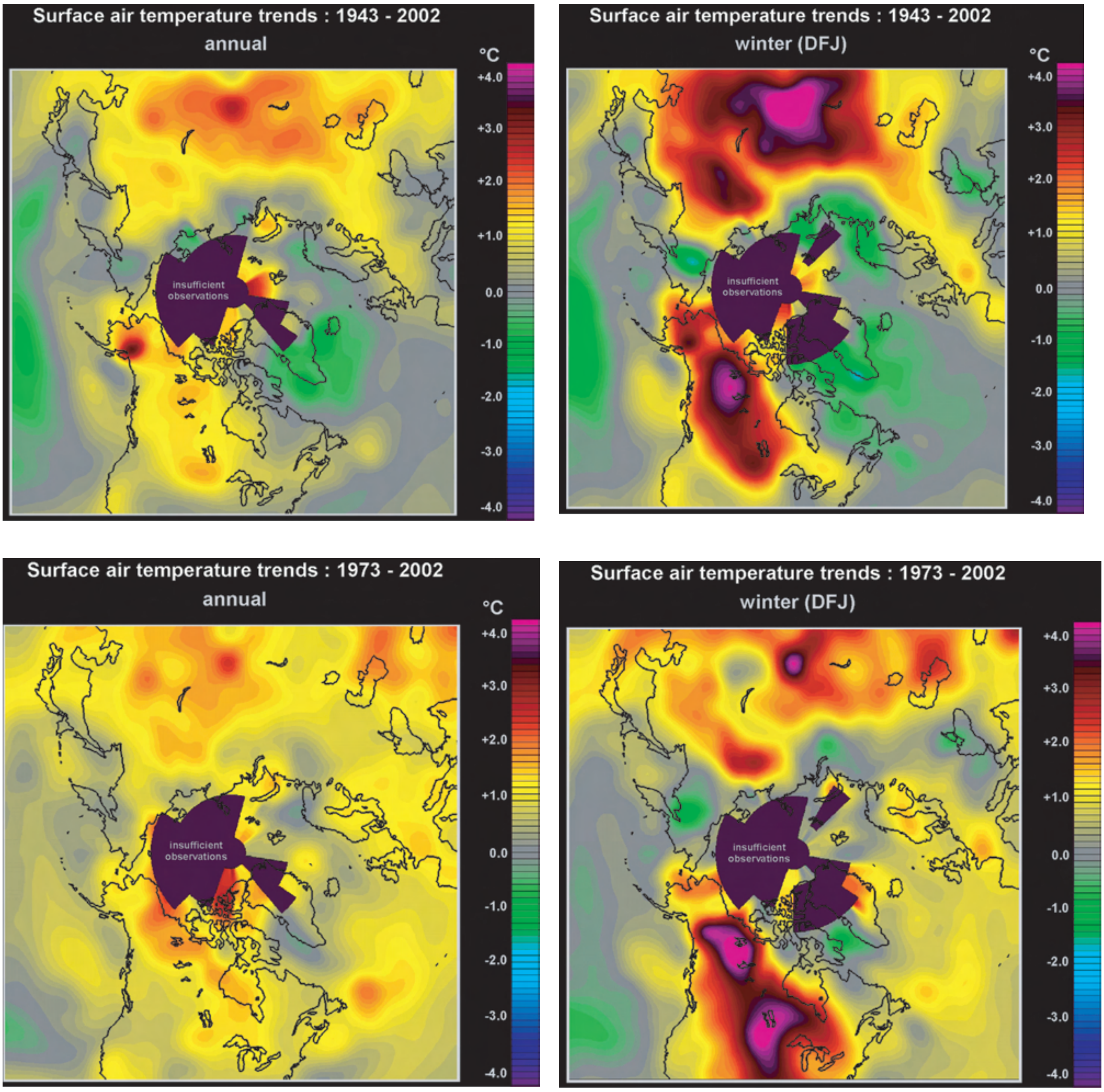
The ACIA

Climate variability and change have become important issues in the Arctic over the past few decades. It has become imperative to examine possible future impacts on the environment and its living resources, on human health and on relevant economic sectors. The Arctic Climate Impact Assessment (ACIA) is expected to lead to useful information for the nations of the Arctic region, their economy, resources and peoples. It is a four-year project of the Arctic Council (a high level intergovernmental forum of the eight Arctic countries) and the International Arctic Science Committee that started in 2000 and will be completed in 2004. The assessment is being conducted by international experts and with broad participation from many different disciplines and countries. About 180 lead and co-lead authors, contributing authors and consulting authors are involved in writing the assessment.

Figure 1



Past climate changes

To provide a basis for assessing the impacts of climate change, Arctic surface air temperature changes for two periods (1943-2002, and 1973-2002) were analysed by Chapman and Walsh (2003) for the ACIA. The database used for this analysis is from the Climate Research Unit of the University of East Anglia, UK. The observed changes from 1943-2002 showed a warming of up to 3˚C over Siberia and Alaska in the annual mean, and a cooling over Southern Greenland and parts of the Barents Sea of about 1˚C (Figure 1a). Winter temperatures in Siberia and Northwest Canada were warmer by up to 4˚C, but there was a cooling of 1˚C in southern Greenland, Labrador and Northern Scandinavia even in winter (Figure 1b). Trends from 1973-2002 were similar (Figures1c and 1d) but there were regional variations in both warming and cooling for the two periods.

Future climate changes

To project future changes in climate, the ACIA used two IPCC scenarios, i.e. the SRES A2 and B2 scenarios. Both are "moderate" climate change scenarios containing projections out to the year 2100. The scenarios were implemented on five GCMs: CCC, GFDL, CSM, UKMO, and ECHAM under agreements with each of these modelling centres, and the results were provided to the ACIA authors. Time slices around 2020, 2050 and 2080 were used, which are the ones also used by the IPCC. Results show fairly good agreement of the five model projections for the region north of 60˚N to mid-century, a warming of about 2-3˚C (Figure 2). At the end of the century there is more scatter between the model results and a projected temperature increase of 4-7˚C.

Discussion

Composite five-model projections of July and January mean surface air temperature changes between 1980-1999 and 2070-2089 are shown in Figure 3a (June) and Figure 3b (December) and can be found on the web page on ACIA scenarios (Chapman & Walsh 2002). Annual mean temperatures show a fairly uniform warming of 2-4˚C throughout the Arctic, with a slightly higher warming of up to 5˚C in the East Siberian Sea. Summer temperatures are 1-2˚C warmer on land, with little change in the central Arctic Ocean, while winter temperatures show the greatest warming of about to 5˚C on land, and up to 8-9˚C in the central Arctic Ocean. Regional and seasonal differences between the individual model results can be large, however.

The ACIA webpage, which contains general details on the ACIA, can be found at: http://www.acia.uaf.edu.

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Figure 2

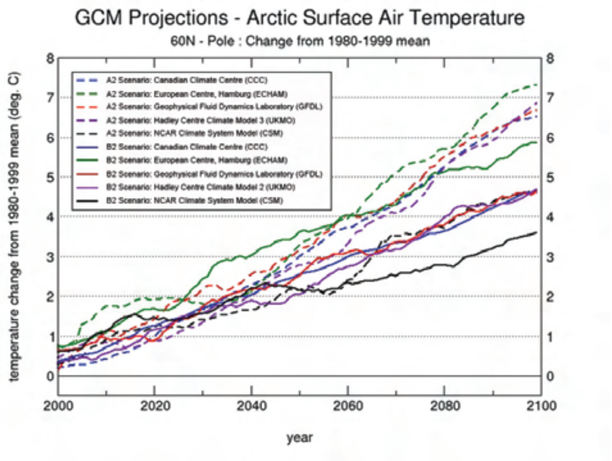
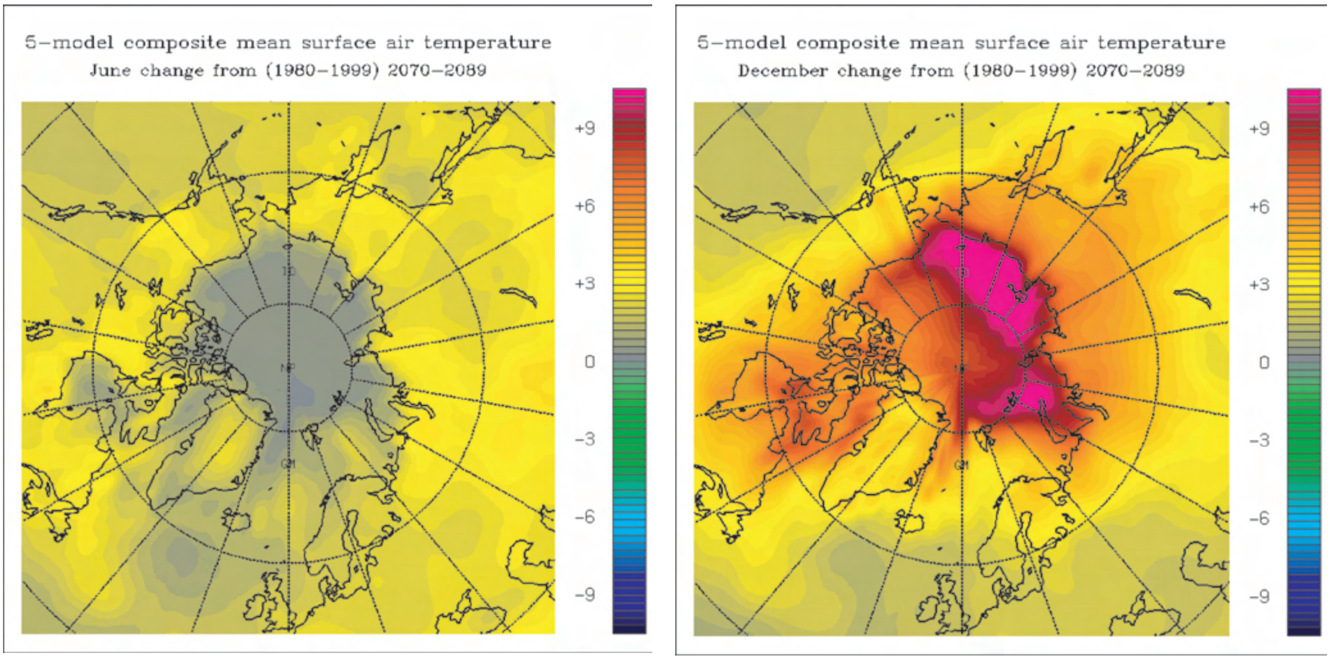


Figure 3



Variability and Feedbacks of UV-Radiation and Surface Radiation Budget in the Arctic

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Introduction

The Arctic climate is influenced by natural and anthropogenic forced changes in large-scale atmospheric and oceanic circulation patterns, the amount and distribution of sea-ice and snow, increasing greenhouse gases and solar cycle variability. The associated integrated modes of natural variability in the northern hemisphere are described by the Arctic Oscillation (AO) as well as the North Atlantic Oscillation (NAO) (Thompson & Wallace 1998). These systems have, together with regional patterns of low pressure over Iceland and relative high pressure over Greenland and the Arctic Ocean, large impacts on the weather patterns forcing warm and humid air along the cyclone tracks into the Norwegian and Barents Seas. In this area, the frequent exchange of air masses of Arctic or Atlantic origin cause highly variable weather conditions, especially during winter (Hanssen-Bauer et al. 1990).

The radiation climate in the Arctic is modified by the spatial and temporal variability of clouds, surface albedo, atmospheric aerosols, stratospheric ozone and the seasonal variation in solar elevation. Important inherent processes and radiative feedback mechanisms determine the amount of solar and infrared radiation that reaches the Earth's surface and is reflected/emitted from it, which in turn influences the hemispheric and global scale climates. Here we illustrate the variability of the cloud radiation and surface albedo feedback mechanisms with selected measurements from Ny-Ålesund, Svalbard (78.9° N, 11.9° E). The Arctic radiation and climate regimes are a mixture of oceanic, coastal and continental types, strongly modified by the regional variability of the surface albedo and clouds. Ny-Ålesund has a coastal type climate with significant oceanic influence during summer and autumn, and a continental type climate during the winter and spring (Førland et al. 1997). The radiation regime is also under strong influence by the open ocean to the west (and south) of Svalbard during summer and autumn, with the Arctic sea fog and low albedo playing a key role. During winter and spring, it is more similar to the central part of Svalbard, due to the frozen fjords, drifting sea-ice and seasonal snow cover (Ørbæk et al. 1999). The heterogeneous surfaces induce large local variations in the radiation fluxes during the melting season in spring.

Radiation climate variability and feedbacks

In general, clouds attenuate and snow/ice-cover enhances solar radiation at the surface. Arctic clouds and surface albedo, although not considered part of radiative forcing (IPCC 2001), are the major players in modifying the amount of solar radiation that reaches the surface and in balancing the surface infrared radiation fluxes over the Arctic Ocean. Figure 1 shows the seasonal variation of representative climate radiation parameters of Ny-Ålesund, Svalbard. The strong seasonal cycles in radiation fluxes with the net outgoing infrared radiation during winter are partly balanced by the poleward transport of warm water and humid air from the north Atlantic. The large seasonal variability in both solar and infrared radiation fluxes reflects the annual variation from polar night to polar day conditions, the significant fluctuation and exchange of air masses and the inhomogeneous surfaces in the Arctic.

The maximum surface solar radiation is shifted towards early June with increased mean radiation in spring as compared to the autumn values (Figure 1a). This is due to persistent snow cover and a higher frequency of clear days in spring (April, May) than in the summer months. Snow enhances surface radiation, because multiple reflections between the surface and the atmosphere occur. This effect also reduces the difference between clear and cloudy weather conditions. From late June till the end of August the multiple reflections are not as effective with the low tundra albedo, causing larger variation in solar radiation between clear and cloudy conditions. The shortwave net radiation is therefore still at the maximum after solstice (Figure 1c).

Infrared down-welling radiation peaks after the summer solstice during the months of July and August (Figure 1b), due to a warmer, stable and more humid summer atmosphere, which also reduces the variance induced by clear and cloudy days. The variability pattern is somewhat opposite of the solar radiation – highest during the winter months and much lower in summer. This reflects the fluctuating conditions during winter between periods of stable, cold Arctic air and warmer, humid air transported from the north Atlantic. The infrared net radiation (Figure 1d) over this coastal tundra site is persistently negative throughout the whole year with some increased variability during the intense melting season, which induces instabilities in the infrared radiation balance between the surface and the atmosphere.

Short-wave transmittance, cloud radiation and surface albedo feedbacks

The surface albedo of Ny-Ålesund (Figure 2a) shows large annual and interannual variations. During the intense melting season,

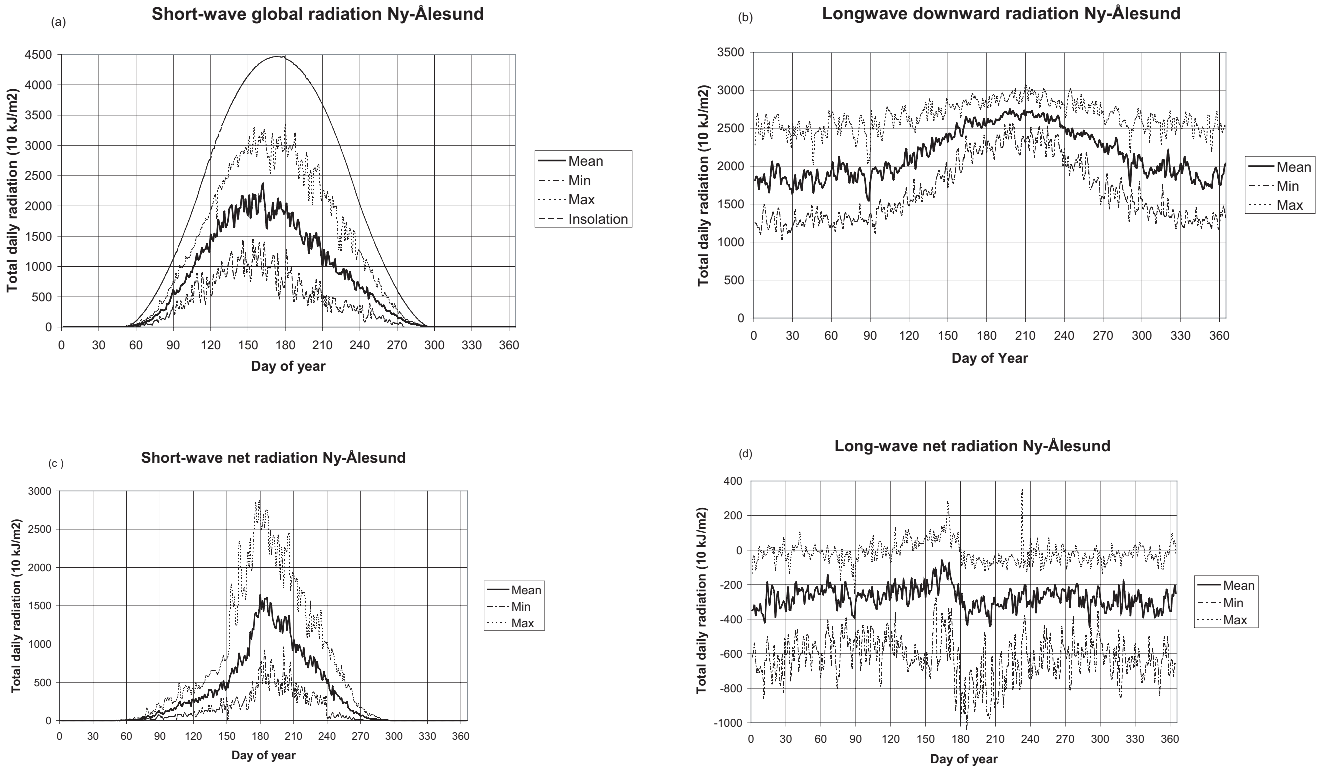


Figure 1: Daily mean, max and min. solar (a), infrared down-welling (b), shortwave net (c) and infrared net (d) radiation for the period 1981-1997 in Ny-Ålesund. (Ørbæk et al. 1999).

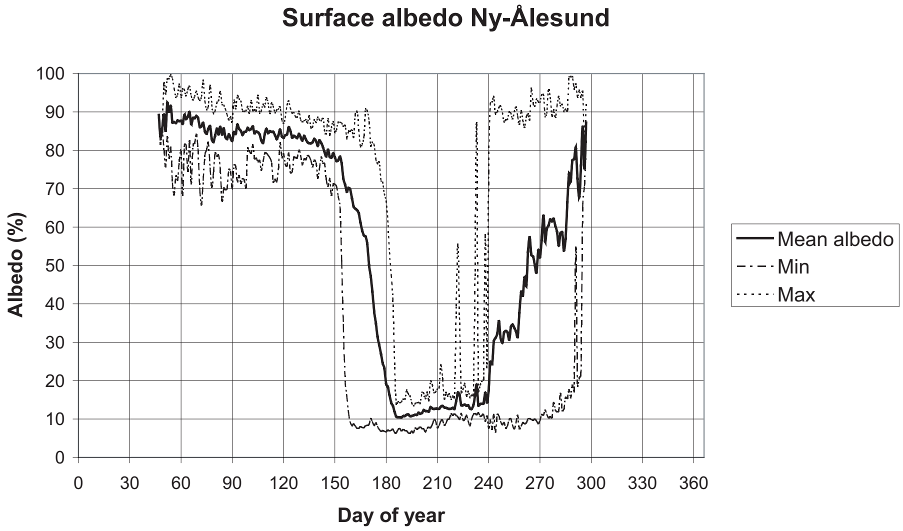
starting in the beginning of June, the albedo at this coastal tundra site drops down from winter/spring values above 80% to summer values below 20% within approximately up to three weeks time. From the end of August the snow returns, but does not stay permanent before the end of October. The successive reflections and multiple scattering of the solar radiation between the ground and the atmosphere and clouds enhance the solar radiation over the high albedo surfaces, hence apparently increasing the atmospheric transmittance during winter and spring. Based on the mean solar radiation fluxes (Figure 1), the mean, maximum (clear sky), minimum (cloudy sky) and cloud transmittance for solar radiation are plotted in Figure 2b. The mean, clear and cloudy sky transmittance are calculated with respect to the TOA-radiation flux, whereas the transmittance due to the clouds alone is retrieved as the ratio of cloudy sky to clear sky transmittance (clear sky 100%). Comparing the spring (doy 120-150) and summer (doy 210-240) periods, the dataset shows that the albedo feedback effect for solar radiation is of the order 10% for clear sky conditions and further increased to approximately 20% for cloudy sky conditions. The high surface albedo intensifies the effective multiple reflection and refraction processes between the snow covered ground and the atmosphere/clouds. The cloud transmittance shows that clouds more effectively attenuate solar radiation during summer by 20%.

UV-radiation, ozone and cloud/albedo feedback

The most important factors that affect the short- and long-term variability of surface UV-radiation are the stratospheric ozone, surface albedo, atmospheric aerosols and clouds. Arola et al. (2003) found that the effect of ozone on the short-term variability of monthly mean UV-irradiance could be as high as 100% with a mean value of 35%, whereas the effect of clouds was up to 40% with typical values at 12%. Their albedo-related effect showed to be up to 21% with average 7% on monthly basis. Figure 3 shows the effect of ozone, clouds, and surface albedo on surface UV-radiation in Ny-Ålesund. The UVSPEC (www.libradtran.org) radiation transfer model was used assuming clear sky conditions and summer albedo values (10%), comparing the simulations to real measurements of CIE (MacKinley & Diffey 1987) erythemal UV-doses for the years 1997 and 1998 (Figure 3a). Daily total ozone was taken from the TOMS archive. Contrary to spring 1998, an Arctic “ozone hole” persisted over the site during March/April 1997. The snow covered ground enhanced surface UV-doses by up to 20% during spring, whereas the cloud effect was strongest during summer reducing the surface UV-doses down to as much as 20% of its clear sky values. Josefsson & Landelius, (2000) reported comparable cloud modification factors for overcast days (0.35±0.2).

Figure 3b shows the relative amplification factor for surface CIE erythemal UV-doses with respect to changes in total ozone. The difference in the modelled surface clear sky UV-doses (excluding atmospheric aerosols) with respect to the real ozone values between the two years are plotted together with the similar differences in measured quantities (clear and cloudy sky), plotted separately for spring (high albedo) and summer (low albedo) conditions. The plot qualitatively shows that the increase in UV-dose with respect to the decrease in ozone follows the approximate relation: $\Delta UV\% = 5/4\ \Delta O3\%$. The increase in surface UV-doses between the two years was up to 50%, with an approximate 40% reduction in ozone. The figure also shows that the amount of clear days during winter/spring is higher than during summer, and that clouds more effectively reduce the UV-doses during summer than during winter.

a)



b)

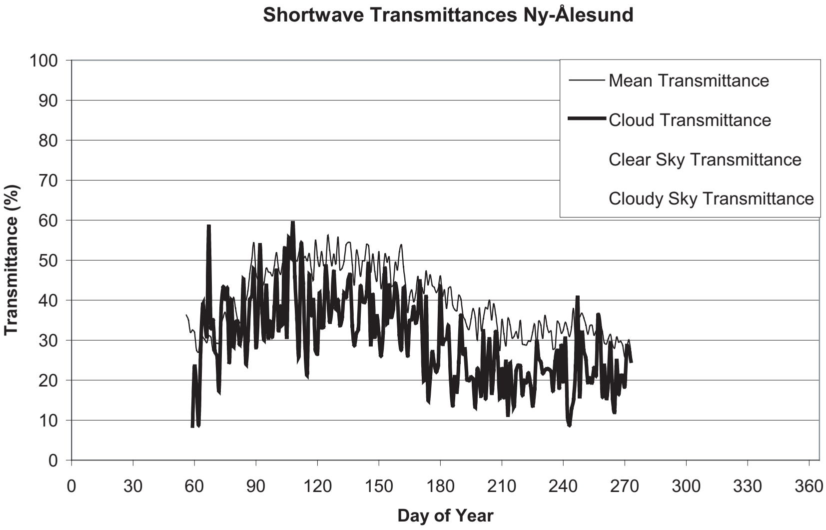


Figure 2: The mean, maximum, minimum albedo (a) and mean, maximum (clear sky), minimum (cloudy sky) and cloud transmittance of solar radiation for Ny-Ålesund (b).

Conclusions

Observations during the past decades show a trend towards the positive phase of the Arctic Oscillation (AO), induced by enhanced westerlies and advection of warm ocean water in the north Atlantic, giving increased surface temperatures over land and intensified transport of warm and humid oceanic air to the continents as well as to the Arctic (Thompson & Wallace 1998). This trend is believed to be sustained by the continued increase in atmospheric greenhouse gases as well as the associated strengthening of the polar stratospheric vortex due to the depletion of ozone (Shindell et al. 2001). The positive phase of the AO increases the poleward transport of heat and the Atlantic influence on the climate parameters in the Arctic. With a continued reduction in sea-ice distribution, increasing atmospheric temperatures and precipitation, the regional variability in radiation quantities is expected to increase, due to more inhomogeneous surfaces and larger difference in fractional cloud cover over land and sea. Reduced levels of solar radiation and harmful UV-radiation and a net positive infrared radiative forcing over the ocean and coastal areas should be associated with these changes. However, Benner et al. (2001) observed and modelled net negative forcing of inhomogeneous clouds and ocean ice surfaces, concluding that the uncertainty in the horizontally averaged surface albedo had a large impact on the radiation fluxes. The Arctic Ocean is the key area for such complex ocean–ice–atmosphere exchange processes. The radiation properties on different scales, the sign and size of the radiative feedback processes connecting fractional clouds and inhomogeneous sea-ice surfaces over the Arctic basin, their contribution to the climate trend patterns observed in the Arctic and their interaction with the large scale modes of the climate system, are still not well known.

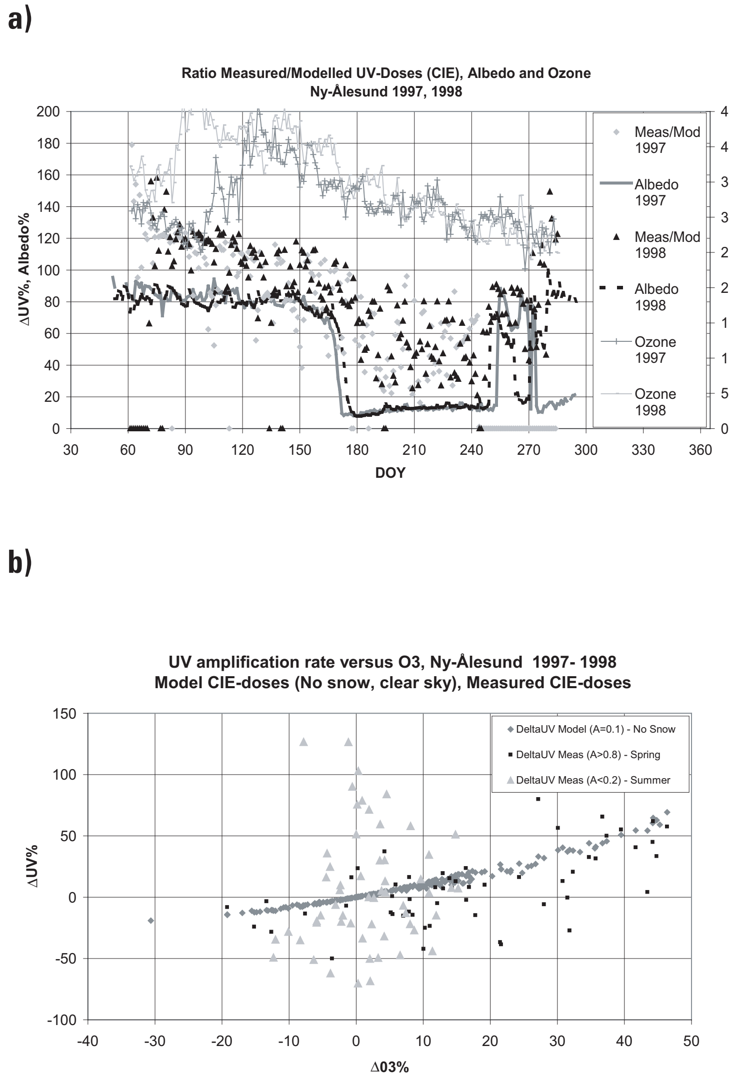


Figure 3: Intercomparison of measured and modelled solar CIE erythemal UV-doses for Ny-Ålesund during 1997, 1998 (a). The increase in surface UV-doses with respect to the decrease in ozone is plotted in (b) for simulated and measured UV-doses during spring and summer.

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On some Ocean Mixing and Transport Processes Involved in Feedbacks to Changing Heat and Fresh Water Fluxes

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Global climate change can be amplified by positive feedback mechanisms even if the process responsible for the feedback occurs only in limited areas such as the Arctic. On the other hand, regional impacts can be smaller than expected if negative feedbacks occur. Three ocean related feedbacks that may be involved in rapid response to global warming merit attention:

- Warm water inflow leading to ice melt (one way of activating the ice albedo feedback);
- Intensified water cycle, wind driven ice production and ice melt pushing open ocean convection regions and overturning circulation further south; and
- Reduced sea-ice, warmer water and changed ocean circulation giving increased ocean carbon uptake (and more general implications for biota).

The net result in all these cases is sensitive to small-scale processes related to vertical fluxes in the ocean. This may well be the case also for other feedbacks in the coupled atmosphere-ice–ocean system. It follows that there is a need for understanding and parameterization of high latitude ocean transport and mixing in climate models in order to obtain reliable regional and indeed global climate scenarios.

The concept of feedback is problematic when applied to complex, nonlinear systems such as the real climate or climate models. In particular, verification of feedbacks and comparison between modelled and observed feedbacks is complicated by the simultaneous action of several interrelated feedback mechanisms.

Aerosol Related Influences on Climate in the Arctic

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Introduction

Aerosols are liquid or solid particles suspended in the air. Aerosols influence the climate (radiative) forcing through the direct effect scattering and absorbing solar and infrared radiation in the atmosphere. Aerosols also influence the radiative forcing through the indirect effects both of the 1st kind by increasing the droplet number with increases in aerosol as well as the 2nd kind with a decrease in precipitation efficiency with increases in aerosol. In particular the indirect effects remain poorly quantified. In the Arctic there also is a further potential influence on the radiative balance by deposition of aerosol altering the albedo of snow and ice surfaces. Aerosols have most likely made a negative contribution to the overall radiative forcing. The level of uncertainty regarding the effect of aerosols is high in the Arctic due to several specific features discussed further below.

The role of aerosols in the Arctic

Aerosols are emitted directly to the atmosphere from the Earth's surface both naturally and anthropogenically. These aerosols are generally in coarser modes. A significant portion of the aerosols is formed within the atmosphere through conversions of precursor gases forming fine particles. Aerosols are characterized by short atmospheric lifetimes causing great spatial and temporal variability. This variability causes large uncertainties in our understanding of the aerosol influences in the sparingly sampled Arctic atmosphere.

Arctic aerosols are characterized by an essentially bimodal behaviour during the year. Wintertime aerosols are relatively low in number concentration but high in mass typical of aged aerosols with little new particle formation. During the winter precipitation rates are low giving inefficient washout and low oxidation rates prevail due to the lack of sunlight. At polar sunrise there is a rapid switch to the summertime mode with low mass concentration but high number concentrations. These two aerosol modes are further complicated by large differences in typical circulation patterns during the year; the wintertime situation brings in aerosols strongly influenced by anthropogenic emissions in Europe and Russia, whereas the summer situation mainly reflects natural sources within the Arctic. The potential influences of these two seasonal states are intriguing since the incoming radiation differs fundamentally between the seasons.

Challenges

The modelling of aerosols is complicated involving both formation, transport and, very significantly, interaction with the atmospheric water and precipitation. Great challenges remain in the Arctic: because mixed phase and pure ice clouds are so important, a significant portion of the transport can occur in thin layers that remain unresolved in models and little data coverage.

Climate feedbacks can occur through changes in sources of precursor gases, alteration of transformation rates and changes in transport pathways to the Arctic.

A significant component of the International Polar Year (IPY) should be a concerted action with models and measurements to determine the representativeness of existing measurement programmes, the limitations of the satellite platforms available and the resolution requirements of the models to depict the Arctic aerosol influences in a quantitative and robust manner.

The Changing Sea-Ice Mass Budget of the Arctic

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Thinning of Arctic sea-ice

Observations from satellites have shown that since the 1970s there has been a steady decline in the annual average area occupied by sea-ice in the Arctic, of some 3-5% per decade. Older observational evidence suggests that this decline began in the 1950s or earlier. More spectacularly, during the last 2-3 decades there has been a large decline in the mean thickness of Arctic sea-ice. The first measurements of the decline were done by submarine sonars, and suggested a thinning of 40% or more in summer in regions surveyed by the US and UK submarines, with a somewhat lower decline in winter. Other measurements have been carried out from moored sonars, drifting wave-ice buoys and satellite altimeters, and suggest lower rates of thinning, while some models predict a significant geographical variability in the thinning rate due to the changes in Arctic ice dynamics which occurred in the 1980s-1990s in response to a change of phase of the Arctic Oscillation.

Review of the ice thickness observations

We reviewed the data sources involved, in an attempt to answer questions such as when the thinning started, how much the ice has thinned seasonally and annually when averaged over the Basin, what (if any) are the significant spatial variations in the thinning rate, and whether the thinning has slowed or stopped. We then reviewed the question of the causes and mechanisms of the thinning, with effects such as enhanced bottom melt due to changed near-surface hydrography, or increased surface melt during a lengthened melt season, having been proposed as responsible. Model predictions of the likely continued course of thinning and retreat through this century are reviewed. Finally we looked at some of the implications of the changing mass budget of sea-ice. One of these, as proposed recently by Walter Munk, is that if the melt rate of sea-ice lies at the upper end of the quoted values (i.e. those derived from submarines) it is sufficient to account for the observed freshening rate of the world ocean as estimated by Levitus. This leaves no scope for a net contribution from land runoff, so glacier melt would have to be offset by other ways that fresh water is locked up, with implications for the rate of global sea-level rise.

Abstracts from Poster Presentations

Name /title

Snow and Vegetation Distribution

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Snow depth and snow distribution have major influence both on the length of the growing season and the access to water, and also have a major influence on the vegetation distribution in an area. This is especially evident in Arctic regions where both the growing season length and water access are limiting factors to plant growth. In this study snow depth and snow distribution were mapped in detail in a 1.5km x 2km area close to Ny-Ålesund (78°55'N, 11°56'E) in Svalbard, Norway. The snow depths were measured at the end of the snow accumulation period by a GSSI Georadar with a 500 Mhz antenna for every 0.25 m along lines with 100m separation. These data were interpolated in SURFER and the snow distribution was found. The average snow depth in the area was 0.7 m ranging from 0m up to a maximum of 2.7m. The snow distribution data were compared to a detailed vegetation map for the area and the correlation between snow depth and vegetation type was studied. This project was financed by the Research Council of Norway under the programme "Arctic Light and Heat".

Reconstruction of the Long-Term UV Climate at a North-Norwegian Location in Spring

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Solar ultraviolet radiation (UVR) has been recognized as an important parameter of the climate system. UVR may have a significant impact on ecosystems, but quantitatively such effects have only been investigated to a limited degree. In the frame of the EU project UVAC, the possible effect of UVR on the first life stage of fish stocks in spring was investigated. While ground-based UVR measurements only exist for the last decade, information on the most important parameters affecting UVR (total ozone, cloud cover) can exist over much longer time scales. In Tromsø, total ozone has been measured since 1935, while at a meteorological station in the spawning area of the Northeast Arctic cod, meteorological parameters, including clouds, have been observed on a daily basis since 1933. By combining these two datasets (and some supplementary data), daily UVR and PAR (photosynthetically active radiation) doses of the period March - May for the whole 68-year record were calculated by means of a radiative transfer model. The poster presents the parameterisation approach, data validation (comparison with measurements and other modelling) and an analysis of trends and variability at various time scales. Since 1936 a +2%/decade increase in UVR cod egg mortality rates in the spring months is observed. The time series has recently been extended and the UVR time series with respect to cloud cover and ozone have been analysed.

Atmosphere–Ice–Ocean Interaction Studies in Frozen Svalbard Fjords

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The project Atmosphere–Ice–Ocean Interaction Studies (acronym "AIO", duration 2002-2005) is designed to provide a rigorous basis for selected sea-ice-related aspects of climate hypotheses. The Norwegian project partners'activities are funded by the Research Council of Norway, whereas the complementary US activities are funded by the National

Science Foundation under a US-Norway collaboration scheme.

The main thrust of the project is to perform controlled field experiments using Svalbard fjords as a natural laboratory. The scientific objects include (1) to test proposed parameterizations for ice–ocean heat exchange during conditions of ice freezing, (2) to measure vertical heat and salt fluxes and develop parameterizations of ice–ocean heat exchange during melting, and in conditions when sea-ice is exposed to saline water which is colder than the melting temperature of the fresher ice, (3) to investigate the role of penetrating solar radiation, in particular effects of albedo and snow cover, in determining under-ice ocean temperature structure and bottom ablation of sea-ice, and (4) to study polynya processes where the atmosphere–ice–ocean processes are interactively linked to water mass formation and movement. Preliminary results from the first set of field campaigns, conducted in Kongsfjorden and van Mijenfjorden, show generally different hydrographic conditions for the two studied systems on large, medium and small scales. These differences can be explained by the principal settings of the fjords, resulting in different ice formation and current patterns/strengths.

Optical Properties of Arctic Sea-ice and Snow

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The optical properties of sea-ice and snow are a key element in the albedo feedback process in polar regions. Various studies related to these properties, initiated by researchers from the Norwegian Polar Institute, were conducted over the last decade in the Svalbard area and in the Barents Sea. The observations include both long-term monitoring and expedition work of a few weeks, covering certain characteristic snow and ice situations, such as during freezing and melting. To describe the optical properties of snow and ice, measurements were preformed using radiation sensors, photometers, and spectroradiometers above, in, and below the snow and sea-ice. Among the parameters measured were surface albedo, surface reflectance and snow and ice transmissivity. Near Ny-Ålesund (West-Spitsbergen), the change of optical properties of fast ice from before to after the onset of melt was investigated over several seasons both on land (Arctic tundra) and on the fast ice of Kongsfjorden. These data are supplemented by measurements from the Marginal Ice Zone in the Barents Sea. A longer time series with albedo data from the tundra at Ny-Ålesund illustrates the interannual variability since 1981. Long time series are necessary for the understanding and detection of interannual variability and long-term changes and the processes responsible for these, e.g. the influence of the North Atlantic Oscillation on the albedo in Svalbard.

Arctic Sea-ice Thickness Variability Observed Over a Decade in the Fram Strait

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Recent conclusions on widespread Arctic sea-ice thinning are based on data from submarine sonar surveys. The chief advantage of submarine sonars is that they allow basin-scale surveys to be carried out on a single cruise, giving the geographical variation in ice thickness characteristics. However, a submarine can not, unlike moored upward looking sonars, generate systematic time series of ice thickness at a fixed geographical point. The Fram Strait is the main gateway for sea-ice leaving the Arctic. This relatively narrow strait is therefore a natural location for the task of establishing a systematic time series of Arctic ice thickness at a fixed point in space. The Norwegian Polar Institute has maintained an array of upward looking sonars across the East Greenland Current at 79°N since 1990, with two to four instruments covering the ice drift in the past thirteen years.

With the recent controversy on Arctic sea-ice thinning as a background, the aim of this study is to use a decade of ice thickness measurements by upward looking sonars (ULS) in the Fram Strait to answer the following questions: 1) May a change in the Arctic sea-ice thickness characteristics be observed in the Fram Strait, and 2) What are the main features of the sea-ice thickness variability over the decade? Preliminary investigations reveal significant interannual variability in the monthly averaged ice thickness, but no general trend of thinning.

One main aspect of the study is to improve and verify the algorithms calculating ice thickness. The focus is presently on methods to detect open water, since this significantly affects the statistics of the time series. Two approaches are tested; one using satellite imagery and one using the variance of the timeflight and envelope voltage.

Downscaled Climate Scenarios for the Svalbard Area

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Atmosphere–ocean general circulation models (AOGCMs) are used to produce scenarios for future climate variation caused by increased atmospheric concentrations of "greenhouse gases". The spatial resolution of the AOGCMs is, however, too coarse for many climate impact studies. Regional modelling and empirical downscaling have been applied in order to make local climate scenarios for various regions, including Svalbard. Empirical downscaling from several AOGCMs indicates that the local temperature projections depend critically on the sea-ice projection. Thus, model runs with unrealistic present-day sea-ice conditions should not be applied for making climate projection in this area. The regional climate model HIRHAM was applied to downscale climate scenarios produced by the models ECHAM4/OPYC3 (with emission scenario IS92a) and HadAMh (with emission scenario A2). The former model was downscaled for the time slices 1980-1999 and 2030-2049, the latter for the periods 1961-1990 and 2071-2100. The scenarios were rather similar concerning projected temperature trends. E.g. both scenarios indicate, for the winter season, a warming exceeding 1°C per decade for parts of Spitsbergen and for areas north and east of this island. The precipitation scenarios were locally more different, although common features were found. E.g. in northeastern parts of Spitsbergen, both models project increased winter precipitation as well as increased snowfall during winter

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Phenology as an Indicator of Climate Change

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Phenology as an indicator of climate change effects, PhenoClim, is a five-year project

(2003-2007) sponsored by the Research Council of Norway. The objective of the project is to use in-situ and satellite based data to establish knowledge about ongoing large-scale changes in the phenological cycle and primary production of vegetation in Norway and Fennoscandia, in order to investigate selected biological, economical and social consequences of observed and predicted changes. Phenology, in its simplest terms, is the study of cyclic events of nature, in response to seasonal and climatic changes to the environment. In this project we measure phenological events like the start of the growing season, midsummer and end of growing season by combining remote sensing with phenological observations recorded in the field.

Ice Cores from Svalbard – Useful Archives of Past Climate and Pollution History

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Ice cores from the relatively low-lying ice caps in Svalbard have not been widely exploited in climatic and environmental studies due to uncertainties about the effect of melt water percolation. However, results from two recent Svalbard ice cores, from Lomonosovfonna and Austfonna, have shown that with careful site selection, high-resolution sampling and multiple chemical analyses, it is possible to recover ice cores with preserved annual signals. These cores are estimated to cover at least the past 600 years and have been dated using a combination of known reference horizons and glacial modelling. The δ18O data from both Lomonosovfonna and Austfonna ice cores suggest that the 20th century was the warmest century during the past 600 years. The anthropogenic influence on the Svalbard environment is illustrated by increased levels of non sea-salt sulphate, nitrate, acidity, fly-ash and organic contaminants particularly during the second half of the 1900s. Decreased concentrations of some components in recent decades most likely reflect emission and use restrictions. However, some current-use organic pesticide compounds show growing concentrations in near-surface layers. The distribution of species in these two Svalbard ice cores has probably been altered to some degree by melt, but the records still provide information about major trends in atmospheric variability of both climate parameters and pollution history.

Contamination of Sea-ice and Related Estimates of its Albedo

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The surface of the sea-ice-cover is exposed to natural and artificial contamination due to influence of various factors. The most significant quantity of organic matters is taken up by the ice from sea water during its freezing (phyto- and zooplankton) as well as during the melting period when colonies of microscopic diatom algae develop in melting ponds which are formed on a surface of old and first-year thick ice. Inorganic matters are taken up by the ice from the bottom layers (in shallow waters) or as a result of fresh water input and wind erosion from land. Ice contamination is partly a result of human activity. Artificial contamination is most frequent in the areas of intensive navigation, in river estuaries or some bights and bays where port and industrial facilities and towns are situated. In recent years human activity such as oil and gas mining and transportation within the areas of the Arctic shelf has become more and more real. The special field experiments were performed near point Barrow (Alaska) on the land fast ice and on Svalbard glaciers. We carried out spectral and integral measurements of albedo of natural sea-ice surface and snow–ice surface of glaciers and surface with artificial contamination. The correlations between surface albedo and surface concentration of artificial and natural sediments are discussed. The application of these results to mathematical modelling and applied (industrial activities) problems are considered.

Optimal Forcing of Atmospheric Flows

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A proposed paradigm for climate change given a relatively weak perturbation of the external forcing, is that it will be determined by a changed state population of flow regimes (e.g. Palmer, J. Climate, 1999). If this is the case, the global climate response and the regional differences will be determined by the properties of the flow regimes that increase its occurrence relative to those that become less populated. The climate system's selection of flow regimes will be made whilst the system is in states of transition between several flow regimes. During such transitions, this system will be sensitive to the actual external forcing being exerted, and a modest perturbation in this forcing may systematically change the preferred selection of flow regimes. As a consequence the regional pattern of climate change may differ significantly from the pattern of forcing perturbation. Furthermore, the climate response will be the result of an ensemble of very many regime shifts determined on natural time-scales for such shifts in the climate system.

This paradigm is the motivation for calculating optimal forcing patterns for atmospheric flow changes over five days, over which time-scale the atmosphere may swap between regimes. We have used both an intermediate T21L3 quasi-geostrophic AGCM and the ECMWF IFS model to calculate Forcing Singular Vectors targeted to Northern Europe and the adjacent sea areas and Forcing Sensitivities of selected flow regimes (see Barkmeijer et al, 2003, QJRMS, for definitions). Results so far point to the lower atmospheric temperature forcing in the northern parts of the North Atlantic Ocean as one important contribution to climate sensitivity.

Century to Decadal Scale Record of Norwegian Sea Surface Temperature Variations through the Holocene

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One of the main issues in paleo-environmental research today is to understand the stability and variability of the current climates and natural environments, whether the natural system is stable in its present mode of operation, what internal and external forcings are required to maintain or change it, and what impacts such changes might have. In order to provide such evaluations it is important to assess variability by obtaining time series that extend beyond the length of instrumental records. Sea-surface temperatures (SST) at decadal resolution have been reconstructed from core MD 95-2011. Core MD 95-2011 is located on the Vöring Plateau (66°58.18N; 07°38.36E, 1050 m water depth) along the main axis of the northward flowing warm Atlantic water. It is, therefore, in an ideal position to monitor changes in the northward heat flux to northwestern Europe. The core was dated by AMS C-14 and Pb 210 isotope profiles. SST variations were estimated by means of thee different diatom transfer function methods. Results indicate a division of the Holocene into three climatic periods: the Holocene Climate Optimum (HCO) (10,000-7000 cal years BP), the transition period (7000-3000 cal years BP) and the late Holocene cooling (3000-0 cal years BP). During the HCO SSTs were 4-5˚C warmer than today. The gradual cooling of surface waters since the HCO is in step with the decreasing insolation on the Northern Hemisphere. Superimposed on the general cooling trend there is a higher frequency SST variability, which is in the order of 1-2°C. This is the natural variability of the Norwegian Atlantic Current. There is also clear evidence for late Holocene climatic events such as the “Little Ice Age” and the “Medieval Warm Period”.

Sea-Surface Temperature Variability in the Eastern vs Western Nordic Seas during the last 2000 Years

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Sea-surface temperatures (SST) at decadal resolution have been reconstructed from core MD 95-2011 and core MD 99-2269 based on diatom transfer functions. The cores were collected during the IMAGES cruises of the R. V. Marion Dufresne. Core MD 95-2011 is located on the Vöring Plateau (66°58.18N; 07°38.36E, 1050 m water depth) along the main axis of the northward flowing warm Atlantic water. It is, therefore, in an ideal position to monitor changes in the northward heat flux to northwestern Europe. Core MD 99-2269 is located in the deep Hunafloi trough, off northern Iceland (66°37.53N; 20°51.16W, 365 m water depth). Today the core lies under the influence of the Irminger current, but it also may be influenced by the cold East Greenland current as the Polar front migrates eastward. Core MD 95-2011 was dated by AMS C-14 and Pb 210 isotope profiles and core MD 99-2269 by AMS C-14. Core MD 95-2011 has been studied at about 10-20 years resolution through the last 2000 years. Sea-surface temperature variations were estimated by means of three different diatom transfer function methods. The records show SST variability of 1-2 degrees on decadal timescales. There is clear evidence for late Holocene climatic events such as the “Little Ice Age” (LIA) and the “Medieval Warm Period”. The LIA starts in core MD 95-2011 with a SST fall of 1.5°C within a decade around 1400 AD and lasts until about 1750 AD. Core MD 99-2269 was studied at about four years resolution. The record shows SST variability of 3-4 degrees on timescales less than 100 years. Timing of century-scale late Holocene climatic events at the eastern versus western Nordic Seas is in anti-phase. North Atlantic Oscillation (NAO) type of atmospheric pattern might explain the observed anti-phase relation of SSTs. If so, then during the LIA a NAO ÷ state and during the last 3000 years a NAO + state dominated the atmospheric circulation.

The Effect of North Atlantic SST and Sea-Ice Anomalies on the Winter Circulation in an Atmospheric GCM

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This study addresses the question of whether the observed multidecadal trends in atmospheric flow (such as in the NAO index) may be caused by observed trends in oceanic boundary forcing. Experiments were carried out using the NCAR atmospheric general circulation model with specified sea-surface temperature (SST) and sea-ice anomalies confined to the North Atlantic sector. The spatial pattern of the anomalous forcing was chosen to be realistic in that it corresponds to the recent 40-year trend in SST and sea-ice, but the anomaly amplitude was exaggerated in order to make the response statistically robust. The wintertime response to both types of forcing resembles the NAO to first order. Even for an exaggerated amplitude, the atmospheric response to the SST anomaly is quite weak compared to the observed positive trend in the NAO, but has the same sign, indicative of a weak positive feedback. The anomalies in sea-ice extent are more efficient than SST anomalies at exciting an atmospheric response comparable in amplitude to the observed NAO trend. However this atmospheric response has the opposite sign to the observed trend, indicative of a significant negative feedback associated with the sea-ice forcing. Additional experiments using SST anomalies with opposite sign to the observed trend show a response that is more robust and similar to the sea-ice response.

We partition the total response in each case into a portion that projects onto the NAO (the indirect response) and a portion that is the residual from that projection (the direct response). This empirical decomposition yields physically meaningful patterns in which distinctive horizontal and vertical structures imply different governing mechanisms. The indirect response dominates the total geopotential height response. The direct response is localized to the vicinity of surface thermal anomaly and is baroclinic within the boundary layer and equivalent barotropic in the free atmosphere.

References

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Workshop Programme

Workshop location: Polar Environmental Centre, 9296 Tromsø, Norway - Auditorium.
Workshop time: 17-19 November 2003

Monday, 17 November 2003

12:00		*Workshop registration, coffee and poster setup*
12:30	Sebastian Gerland *Norwegian Polar Institute*	Workshop welcome
Introductory Talk		
12:40	Robert Corell *Harvard University, USA*	The role of the Arctic Climate Impact Assessment
Session 1: Climate programmes (chaired by O. Orheim and G. Weller)		
13:00	Gunther Weller *Alaska Climate Research Center, USA*	Climate data and models in the ACIA project
13:20	Olav Orheim *Norwegian Polar Institute*	ACIA and related climate programmes
13:40	Lars-Otto Reiersen *Arctic Monitoring and Assessment Programme Secretariat, Norway*	The climate monitoring programme of AMAP
Session 2: Terrestrial systems (chaired by N. Reeh & E. Isaksson)		
14:00	Larry Hinzman *University of Alaska, USA*	Hydrological changes in polar regions, impacts of a changing climate
14:30	Annika Hofgaard *Norwegian Institute for Nature Research*	Feedbacks between northern terrestrial systems and climate
15:00		*Coffee break*
15:30	Niels Reeh *Technical University of Danmark*	Glaciology I: Land ice in the climate system
16:00	Elisabeth Isaksson *Norwegian Polar Institute*	Glaciology II: Can Arctic ice cores provide insight to climate feedbacks?
16:30-18:00		**Poster session**
19:30-21:00		**Reception**

Tuesday, 18 November 2003

Session 3: Oceans (chaired by P. Haugan and V. Pavlov)		
8:30	Peter Haugan *University of Bergen, Norway*	On some ocean mixing and transport processes involved in feedbacks to changing heat and fresh water fluxes
9:00	Vladimir Pavlov *Norwegian Polar Institute*	Seasonal and Long-term Sea Level Variability in the Arctic Ocean
9:30	Uwe Mikolajewicz *Max-Planck-Institute for Meteorology, Germany*	Simulating climate variability with a coupled regional atmosphere-ocean-sea-ice model
10:00	Michael Karcher and Ingo Harms *Alfred Wegener Institute & Univ. Hamburg, Germany*	Arctic Ocean shelf-basin interaction
10:30		*Coffee break*
Session 4: Sea-ice (chaired by D. Perovich and P. Wadhams)		

11:00	Peter Wadhams *University of Cambridge, U.K.*	The changing sea-ice mass budget of the Arctic
11:30	Don Perovich *Cold Regions Research and Engineering Laboratory, USA*	Sea-ice and sunlight: the ice-albedo feedback mechanism
12:00	Stein Sandven and Ola M. Johannessen *Nansen Environmental and Remote Sensing Centre, Norway*	Remote Sensing of Arctic Sea-ice for climate feedback studies
12:30	Jari Haapala *Finnish Institute of Marine Research*	Evaluation of the sea-ice component of the ACIA Atmosphere-Ocean General Circulation Models (AOGCMs)
13:00		*Lunch break*
Session 5: Atmosphere (chaired by J. Christensen and K. Holmén)		
14:00	Jens Christensen *Danish Meteorological Institute*	Does climate change in the Arctic hold the key to understand Global change?
14:30	Kim Holmén *Norwegian Institute for Air Research*	Aerosol-related influences on climate in the Arctic
15:00		*Coffee Break*
15:30	Jon Børre Ørbæk *Norwegian Polar Institute*	Variability and feedbacks of the solar UV-radiation and surface radiation budget in the Arctic
16:00	Sigbjørn Grønås *University of Bergen, Norway*	Causes of Arctic climate variations – results from numerical simulations
20:00	Nalân Koç *Norwegian Polar Institute*	**Public talk** (in Norwegian)**:** Raske klimaendringer i Norskehavet: Eksempler fra fortiden (Rapid climate changes in the Norwegian Sea: Examples from the past). "*Driv*", town centre (near Polar Museum)

Wednesday, 19 November 2003

Working groups (divided after sessions) – State of knowledge, challenges and recommendations		
9:00-10:30		Working group meetings, part I
10:30		*Coffee break*
11:00		Working group meetings, part II
12:30		*Lunch break*
13:30		**Plenary working group reports** (15 min. for each group)
14:45		**Final discussion** and closing of the workshop

Acronyms: Institutions, Organizations and Programmes

AARI	Arctic and Antarctic Research Institute (Russia)
ACIA	Arctic Climate Impact Assessment
AMAP	Arctic Monitoring Assessment Programme
AWI	Alfred Wegener Institute for Polar and Marine Research (Germany)
CICERO	Centre for Climate and Environmental Research in Oslo (Norway)
DMI	Danish Meteorological Institute
EPICA	European Programme on Ice Coring in Antarctica
FIMR	Finnish Institute for Marine Research
IARC	International Arctic Research Center (USA)
IMR	Institute of Marine Research (Norway)
IPCC	Intergovernmental Panel on Climate Change
Met.no	Meteorological Institute (Norway)
NILU	Norwegian Institute for Air Research
NINA	Norwegian Institute for Nature Research
NIPR	National Institute for Polar Research (Japan)
NPI	Norwegian Polar Institute
UAF	University of Alaska Fairbanks (USA)
UiB	University of Bergen (Norway)
UNIS	University Centre in Svalbard (Norway)