DANMARKS OG GRØNLANDS GEOLOGISKE UNDERSØGELSE RAPPORT 2010/127

The Dynamics and Mass Budget of Arctic Glaciers

Extended abstracts
Workshop and GLACIODYN (IPY) meeting,
16 -19 February 2009, Kananaskis (Canada)
IASC Network on Arctic Glaciology

A. P. Ahlstrøm & M. Sharp (eds)

GEOLOGICAL SURVEY OF DENMARK AND GREENLAND
MINISTRY OF CLIMATE AND ENERGY

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Contents

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Preface

The 2009 annual workshop of the Network for Arctic Glaciology marked a change in several ways. Apart from a new Chairman and Vice-chairman, the former Working Group on Arctic Glaciology was also re-invented in the form of a IASC Network to accommodate the changes in the way IASC structures its activities.

The Network for Arctic Glaciology will strive to follow the successful tradition of the former Working Group in holding an annual workshop on the dynamics and mass budget of Arctic glaciers. The informal atmosphere and relatively small group of participants, along with the concept of encouraging the attendance of students and young scientists through financial support from IASC, has provided the perfect setting for strengthening the collaboration and knowledge transfer within glaciology across the Arctic.

To support this circum-Arctic trend, it was decided to aim for a venue that would alternate between Europe and North America. Thus, the 2009 workshop was held at the Biogeoscience Institute, University of Calgary, Barrier Lake Station, Kananaskis Country, Alberta, Canada, whereas the 2010 workshop will be held at the Universitätszentrum Obergurgl, University of Innsbruck, Obergurgl, Austria. For 2011 we aim for a venue in the USA. We are grateful to Prof. Martin Sharp and those assisting him for the considerable effort put into arranging this workshop and making it such a pleasant experience. The intimacy and scenic location of the Barrier Lake Station provided a perfect setting for a productive workshop.

Another structural change in the organisation is from having a group of National Representatives holding a closed business meeting during the workshop to an Open Forum with open participation for everyone attending the workshop. The former National Representatives take on a role of National Points of Contact to retain the necessary framework for communication to the National level. The Minutes from the first Open Forum of the Network is printed in this volume.

A total of 36 oral presentations and 12 poster presentations were given at the workshop, many of which have handed in extended abstracts for publication in this volume. The extended abstracts have not been reviewed, but provide useful information on the research being conducted and serve as an inspiration for collaboration and further studies.

I would to thank the University of Calgary for making the Barrier Lake Station available for our workshop and I am grateful to Signe Hillerup Larsen for helping with the editorial work.

Andreas P. Ahlstrøm
Chairman, IASC Network on Arctic Glaciology

Programme

Monday 16 February

<u>9.00-12.00: Volume and Mass Changes</u>

Chair: Martin Sharp

9.00: Anthony Arendt: Validation of Greenland Ice Sheet mass changes from high resolution GRACE masscon solutions
9.20: Tony Schenk: A novel approach for change detection from ICESat satellite laser altimetry – spatial and temporal pattern of ice sheet surface elevation change in NW Greenland 2003-2008
9.40: Andreas Ahlstrom: Programme for Monitoring the Greenland Ice Sheet (PROMICE)
10.00: Ginny Catania: Elevation change over the Canadian Arctic ice caps using ICESat data

10.20-10.40: Coffee

Chair: Jon-Ove Hagen

10.40: Geir Moholdt: Repeat track ICESat altimetry over Austfonna and Devon ice caps
11.00: Chris Nuth: Climatic and dynamic influences on an altimetric mass balance estimate for Svalbard
11.20: Norah Foy: Changes in Surface Elevation of the Kaskawulsh Glacier, Yukon Territory
11.40: Carsten Braun: The Surface Mass Balance of the Ward Hunt Ice Shelf and associated Ward Hunt Ice Rise

12.00-14.00: Lunch

<u>14:00-17.00: Mass Balance and Snow/Ice Facies</u>

Chair: Carleen Tijm-Reijmer

14.00: Jon-Ove Hagen: Four years of surface mass balance measurements on Austfonna Ice Cap, Svalbard
14.20: Thomas Schuler: An accumulation history of Austfonna, Svalbard, derived from reanalysis data
14.40: Tyler Sylvestre: Spatial Variability of Snow Accumulation Patterns across the Belcher Glacier, Devon Island, Nunavut, Canada
15.00: Regine Hock: Energy balance studies on Vestfonna, Svalbard

15.20-15.40: Coffee

Chair: Jason Box

15:40: Gabriel Wolken/Martin Sharp: Inter-annual variability in snow/ice facies distributions on Arctic land ice
16.00: David Burgess: Mapping Glacier Facies Across the Devon Island Ice Cap, Nunavut from RadarSat-2 Data
16.20: Joel Harper: Meltwater infiltration and firn densification in western Greenland.
16.40: Alex Gardner: Parameterization of Shortwave Absorption and Reflection in Snow and Ice

Tuesday 17 February

09.00-12.00: Ice Dynamics

Chair: Regine Hock

09.00: Bea Csatho: Long-term record of ice surface and velocity changes from historical photographs and satellite imagery: Investigating the dynamics of Jakobshavn Glacier since the 1960s
09.20: Hester Jiskoot: 2001-2007 East Greenland valley glacier surge and its implications for surge mechanisms
09:40: Carleen Reijmer: Svalbard GPS measurements
10.00: D.Puczko: Fluctuations of flow velocity of Hansbreen tidewater glacier in S Spitsbergen

10:20-10.40: Coffee

Chair: Ginny Catania

10.40: Thorben Dunse: Modelling flow dynamics and form of the Austfonna ice cap, Svalbard
11.00: Sam Pimentel: Coupling Glacial Hydrology into a High-order Numerical Ice Model
11.20: Charlotte Delcourt: Climate sensitivity of McCall Glacier, Alaska, based on radio-echo sounding and numerical flow modelling
11.40: Peter Nienow: TBA

12.00: Lunch

14.00 – 17.00: Glaciodyn Discussions – off site

Wednesday 18 February

<u>09.00-10.40: Radar and More</u>

Chair: Jacek Jania

09.00: Daniel Binder: Ground Penetrating Radar Investigations in the North East of Greenland
09.20: Francisco Navarro: On the hydrothermal state of Ariebreen, Spitsbergen
09.40: Jon-Ove Hagen: New low-frequency radio-echo soundings of Austfonna Ice Cap, Svalbard
10.00: Matt Nolan: 5 months on McCall Glacier in IPY4
10.20: Martin Sharp: Chemistry of a new Deep Ice Core from Ellesmere Island

10.40 – 12.00: Coffee and Poster Session

<u>14.00-17.00: Tidewater Glaciers and Iceberg Calving</u>

Chair: Julian Dowdeswell

14.00: Gabriel Wolken: Inventory and near-terminus velocity estimates of tidewater glaciers in Arctic Canada
14.20: Brad Danielson: Glacial-Marine Interactions at the Belcher Glacier Terminus
14.40: Jason Box: Terrestrial photogrammetry of Greenland glacier discharge variability: comparison with surface climate anomalies
15.00: Michelle Koppes: Submarine melting at the termini of subpolar outlet glaciers, western Greenland

15.20-15.40: Coffee

Chair: Andreas Ahlstrøm

15.40: Jacek Jania: Importance of topographic and geologic conditions for behaviour of Spitsbergen glaciers retreating from the sea to land
16.00: Cecilie Rolstad: Ground based interferometric radar for velocity and calving rate measurements of the tidewater glacier Kronebreen, Svalbard, from August 2007 and 2008.
16.20: Anne Chapuis: Investigating the calving rate variations of Kronebreen, Svalbard, using terrestrial photogrammetry and direct observations
16.40: Tad Pfeffer: TBA

19.30: IASC-NAG Business Meeting

Participant list

Name	Affiliation	Country
Agnieszka Izykowska	Polish Academy of Sciences	Poland
Alex Gardner	University of Alberta	Canada
Alexander Klepikov	Arctic and Antarctic Research Institute, St Petersburg	Russia
Andreas Ahlstrom	GEUS - Geological Survey of Denmark and Greenland	Denmark
Angus Duncan	University of Alberta	Canada
Anne Chapuis	University of Life Sciences, As	Norway
Anthony Arendt	University of Alaska, Fairbanks	USA
Bea Csatho	University at Buffalo, SUNY, New York	USA
Bernard Hynek	Central Institute for Meteorology and Geodynamics, Vienna	Austria
Bernhard Rabus	MacDonald Dettweiler Associates	Canada
Brad Danielson	University of Alberta	Canada
Carleen Tijm-Reijmer	IMAU, University of Utrecht	Netherlands
Charles Webb	University of Texas, Austin	USA
Charlotte Delcourt	Universite Libre de Bruxelles	Belgium
Chris Nuth	University of Oslo	Norway
Christine Dow	University of Alberta	Canada
Dan Juhlin	University of Lethbridge	Canada
Dan Moore	University of British Columbia	Canada
Ed Josberger	Washington Water Science Center, USGS	USA
Emilie Herdes	University of Ottawa	Canada
Emily Moss	University of Alberta	Canada
Evelyn Dowdeswell	Bristol Glaciology Centre	UK
Faye Wyatt	University of Alberta	Canada
Francisco Navarro	Universidad Politécnica de Madrid	Spain
Frank Pattyn	Universite Libre de Bruxelles	Belgium
Gabriel Wolken	University of Alberta	Canada
Gabrielle Gascon	University of Alberta	Canada
Geir Moholdt	University of Oslo	Norway
Ginny Catania	University of Texas, Austin	USA
Gordon Hamilton	University of Maine	USA
Grzegorz Gajek	Maria Curie-Sklodowska University In Lublin	Poland
Gwenn Flowers	Simon Fraser University	Canada
Hannah Milne	University of Alberta	Canada
Hester Jiskoot	University of Lethbridge	Canada
Inka Koch	University of Alberta	Canada
James Davis	University of Alberta	Canada
Jason Box	Ohio State University	USA
Jon-Ove Hagen	University of Oslo	Norway
Julian Dowdeswell	Scott Polar Research Institute, University of Cambridge	UK
Krzysztof Migala	University of Wroclaw	Poland
Laurence Gray	Canada Centre for Remote Sensing, Ottawa	Canada
Maria Ananicheva	Institute of Geography, Russian Academy of Sciences, Moscow	Russia
Mariusz Grabiec	University of Silesia	Poland
Martin Sharp	University of Alberta	Canada
Matt Nolan	University of Alaska, Fairbanks	USA
Michelle Koppes	University of British Columbia	Canada
Nick Barrand	University of Alberta	Canada
Norah Foy	University of Ottawa	Canada
Peter Nienow	University of Edinburgh	UK
Piotr Glowacki	Polish Academy of Sciences	Poland
Regine Hock	University of Alaska, Fairbanks	USA
Ruth Mottram	GEUS - Geological Survey of Denmark and Greenland	Denmark
Sam Pimentel	Simon Fraser University	Canada
Sarah Boon	University of Lethbridge	Canada
Shawn Marshall	University of Calgary	Canada
Tad Pfeffer	INSTAAR, University of Colorado, Boulder	USA
Tara Moran	University of Calgary	Canada
Thorben Dunse	University of Oslo	Norway
Tony Schenk	Ohio State University	USA
Tyler Sylvestre	University of Ottawa	Canada
Wendy Clavano	University of Alberta	Canada

Minutes from the IASC Network on Arctic Glaciology Open Forum

Wednesday 18th February 19.30 – 21.00
Barrier Lake Field Station, University of Calgary, Kananaskis County, Alberta, Canada

Agenda

1. Additional points to the agenda from the participants?

2. Open discussion on the future structure of the IASC-NAG
- should we abandon the idea of national representatives?
- should we establish a less formal group of national contacts?
- is the open forum a suitable replacement for the formal business meetings?
- what is expected of the network from IASC?
- can we learn something from the SCAR structure?
- how should we respond to the SCAR-originated ISMASS group coming to the Arctic?

3. Suggestions for science objectives of the network:
- determine the fraction of calving/tidewater ice masses in the Arctic (remote sensing), what part is below present sea-level (radar)
- intercomparison of ICESat elevation methods
- altimetry methods in general
- facies mapping of ice masses in the Arctic (incl. intercomparison of techniques)
- establishment of quantifiable glaciological indicators of climate change (blockbusters like melt area extent, or something equivalent of the Arctic sea ice reduction)
- have sessions at large meetings that are labelled with the IASC-NAG brand
- a full glacier inventory of land ice masses in the Arctic – define what the needs are provide justification for the GLIMS work, utilize the data already there
- have sessions specifically addressing a certain subject
- others…?

4. Future funding opportunities for cross-network projects?
- EU, NSF, ESF (Polar Climate, more?), ESA, Nordic Council of Ministers, …

5. Suggestions for the shape and role of the IASC-NAG website

6. Location of meeting next year

7. Publication of extended abstracts

8. Other

Introduction

The open forum was opened and introduced by APA who went through the details of the agenda. He also explained that the forum was an opportunity to have a discussion with IASC-NAG members about the future development of the network, particularly in the light of the restructuring of IASC overall.

Item 1. Additional agenda items

APA asked for any additional points to the agenda, but none further were raised.

Item 2. Open discussion on the future structure of the Network of Arctic Glaciology:

APA outlined the past working of the IASC-WAG (Working group on Arctic Glaciology) and the agreed changes to the structure of IASC as a whole and opened discussions by asking for any suggestions on how the new network should be organized. He suggested a loose structure of a network of informal national representatives comprising an executive committee.

Discussion on the “executive committee” followed, with TP asking for some clarification, given by JOH, who suggested the chair did most of the work, with national representatives being largely points of contact within countries.

MS emphasized that IASC want a report on what the network has achieved rather than just knowledge sharing about different countries. APA gave the example of the SWIPA report. JD suggested that a small committee works most efficiently and that national representatives may not be required to avoid duplication with the higher level of IASC. However, informal group contacts for each country are very useful, especially as the Network on Arctic Glaciology gets going. APA asked if the committee should be an open forum, which JD agreed was also useful. MS suggested this would be much more helpful in allowing young scientists to participate also. HJ asked if voting was ever required in business meetings which JD and MS agreed was never really required, there are unlikely to be few contentious issues in any case.

FP asked if network can decide if national representatives are needed by IASC. MS clarified that they are only required at the higher level of SSC, JD pointed out that “countries” will be asked to nominate the reps, individual nations will therefore decide how best to do this. MS suggested it will most likely be a scientific rather than administrative rep and JD further suggested the person or institution who pays the national subscription will likely be the nominating body for higher level national reps. APA further suggested that it would be useful for NAG to have a rep on the SSC.

MS suggested NAG write a proposal to SSC with aims, aspirations and overview to prepare the ground for future requests. JD agreed this would be wise and suggested aiming for a 3 – 5 year overview based on what has been achieved so far.

APA asked about funding opportunities from IASC. MS gave details on the ~€30,000 per year the SSC has for the cryosphere as a whole, not just the network. JD suggested that we ask for more rather than less as the network is already set up and running. JOH said annual grants are usually of the order of €8-10,000, which APA explained was a cap for the meeting; this amount can be extended for other activities. He then asked for details on what other activities could expect funding from the SSC. JOH suggested annual meetings and explicit single topic workshops. JOH asked for clarity on what was decided regarding the national reps and the network structure. APA and MS summarised that no formal national reps on the NAG committee were required and yes to an open forum set up. They further agreed to identify informal national reps and ask them to act as such.

Some Suggested Activities for the NAG

1. ISMASS

APA referred to ISMASS as an example, which MS and FN gave some more details about. The project is led by Kees van der Veen and it originally applied to the mass balance of the Antarctic ice sheets, the group were now looking for support for work on the mass balance of Greenland from IASC. MS has asked ISMASS to broaden scope to the whole of the Arctic and a memorandum of understanding signed. APA, JD and MS agreed to contact Kees van der Veen to find out more about the aims of ISMASS

2. Summer School on Glaciology

RH raised the possibility of financing a North American based summer school similar to Karthaus, possibly flying in tutors from overseas. JD suggested alternating Karthaus with a North American venue. CR gave some background on the demand for places, how the Karthaus school is organised and some of the issues behind organising it. RH agreed it could be worthwhile alternating. PN expressed concern that demand from Europe is enough for an annual school in Europe. RH pointed out that Alaska would be a good location for summer school and for North American students. CN said it was a good opportunity to attend Fairbanks. APA raised concern about the funds required to pay for travel. JB also suggested it may segregate the European and American groups further and suggested a 50/50 quota for each. JD pointed out that money from IASC will be limited. He emphasised that it would be nice to do and consultation with Hans Oerlemans would be the first priority. He particularly thanked Hans for the initiative and then asked about finding coordinators for such a summer school in North America. RH said this would not be a problem. It was agreed to contact Hans Oerlemans regarding the Karthaus Summer school and if the NAG can make any useful contribution to it.

Item 3. Suggestions for Science Objectives of the NAG

Project 1: To determine the fraction of calving ice masses in the Arctic

A suggestion introduced by APA. HJ mentioned her earlier work and offered to extend it. AA agreed it was useful but emphasised it should be done in conjunction with the glacier inventories. TP/RH/JB agreed and stated that land ice masses also need including. MS pointed out this had been included in other plans. JD contributed that formalising the standards of such inventories and a focus on output needs are important. APA raised the GLIMS project which is similar in scope. AA raised the issue of funding such a project. TP defined the two problems involved in such a project as 1) Accumulating data and 2) informing the community of the data availability. He also suggested that it is not generally understood that small glaciers have potential for dynamic response. MS emphasised that a strong statement on this would be powerful for the community. JD proposed a small subgroup should look at the planning and implementation of such a project. MS agreed but stated that finding the right people will be key. AA suggested that the data already exists it just has not yet been put into databases yet. MS then suggested that perhaps the purpose is to use GLIMS and to provide the scientific justification for it.

MN suggested that focused sessions within meetings on specific subjects would be helpful to move the scientific agenda forward and to stimulate the writing of papers, for instance. He suggested we identify one for the following year (e.g. glacier inventories). This was generally agreed as a good idea. MS pointed out that the GLACIODYN link with the meetings had increased participation and brought more like minded researchers together, he proposed land ice masses as a topic for next year. JB emphasised that it should just be a session to help focus research but that it would be good to include it. FN suggested that a session on ice grounded below sea level and TP suggested a next step would be marine channels. APA agreed to send out a focused session suggestion for the 2010 meeting.

Project 2: Intercomparison of ICESAT elevation methods

Introduced by MS, the aim is to develop and communicate a best practice around analysis of ICESAT data. AA concurred with identified problem but suggested much was going on around ICESAT outside the NAG and that such an exercise may give a range to the error estimates but was unlikely to find an answer. JD agreed but said it might be worthwhile to have a workshop with other users to find best practice for the community as a whole, also in preparation for ICESAT2. PN suggested including CRYOSAT2 especially in terms of calibration. Some general discussion followed this. JB and JOH suggested focus of one workshop could be laser altimetry to include ICESAT and MN asked to clarify whether workshops should be extra or part of NAG meeting. JD pointed out the choice is up to the network and MN went on to suggest a day or two preceding the meeting with a day in between for the workshops. APA suggested maybe more detailed meetings and MS responded that perhaps self identified groups could work together then report back to community through these meetings. JD gave the example of Texas group and stated that these workshops should be bottom up and demand driven but strongly linked back to the community. MN suggested that deliverables from the group could also be AGU/EGU sessions with a IASC-NAG label, even if these are already being organised. JB pointed out that would also help to advertise and grow the network. This was generally agreed to be a good idea. PN returned to the point about expertise with ICESAT, can it be useful to anyone? GW suggested working closely with ICESAT specialists to use their algorithms. APA and JD suggested contacting the ICESAT developers for their assistance.

Project 3: Blockbuster climate change indicators

APA suggested these would be useful in drawing attention to the work of the NAG and asked for some suggestions. These included: databases of glacier ELAs (JB), glacier lengths (HJ), glacier facies and AARs (MS). JD pointed out the problem of superimposed ice in using ELAs/AARs. MS also raised possibility of using the Arctic climate report card. JB responded that the cryosphere section of this is already growing and it must not be overloaded. MS suggested putting the indicators chosen on the IASC-NAG website for quick reference and communication. MN suggested volume change and AA suggested that for annual assessment mass balance is most practical. APA suggested contacting the WGMS and MS suggested that culling data on Arctic glaciers from there and making it available on the NAG site would make the data more visible and promote the work of the NAG. After some discussion (JOH/CR/MN/RH/APA) it was agreed to use the WGMS arctic data. AA suggested that in the future projections of glacier change would be useful too.

Item 4. Future Funding Opportunities

APA introduced the section and mentioned that the Nordic Council of Ministers would be putting some funding through soon. JOH encouraged an application to this.

Item 5. Website

APA asked for thoughts about updates and the function of the NAG website. Suggestions were: reports/publications/meeting details (CR), datasets outside of the GLIMS/WGMS projects (RH), publications from NAG participants (HJ). APA asked network members to email him with other suggestions.

Item 6. Location of next NAG meeting (2010)

APA asked for suggestions and had already made an enquiry with Michael Kuhn regarding the availability of Obergurgl venue. CR stated that it was nice to vary the venue and suggested alternating between North America and Europe. JB agreed and said 2011 would be good timing for him in Colorado. He offered to work on tentatively organising a meeting in Colorado for the following year. This was enthusiastically agreed.

Item 7. Abstracts

APA reminded participants of the submission deadline for the book of extended abstracts from the 2009 meeting

Item 8. Minutes

RHM promised to type up minutes and make them available on the NAG website.

25 Attendees, main contributors to discussion listed above:

AA – Anthony Arendt
APA – Andreas P. Ahlstrøm
CR – Carleen Reijmer
CN – Chris Nuth
FN – Francisco Navarro

FP – Frank Pattyn
GW – Gabriel Wolken
HJ – Hester Jiskoot
JB – Jason Box
JD – Julian Dowdeswell
JOH – Jon Ove Hagen
MN – Matt Nolan
MS – Martin Sharp
PN – Peter Nienow
RH – Regine Hock
TP – Tad Pfeffer
RHM – Ruth Mottram (minute taker)

Outcomes/Deliverables:

1. APA and MS to identify informal national reps for NAG business committee and ask for participation
2. MS/JD/APA to write short report on NAG for IASC SSC as brief communication on network aims, priorities and achievements etc
3. APA, JD and MS agreed to contact Kees van der Veen to find out more about the aims of ISMASS for possible collaboration
4. Contact to be made with Hans Oerlemans regarding summer school coordination and funding. (APA?/MS?/CR?/RH?)
5. MS/APA to arrange and send out a suggestion for a focused session at 2010 NAG meeting (possibilities include: calving glaciers, land ice masses, glaciers grounded below sea level, marine channels, glacier inventories, laser altimetry…)
6. ICESAT/Laser altimetry identified as an area of expertise, a workshop to be organized (by self selecting subgroup) and to include ICESAT team to look at best practice.
7. APA/MS to use WGMS arctic glacier data as climate change indicator on website.

Extended abstracts

PROMICE - Monitoring the mass loss of the Greenland Ice Sheet

Andreas Peter Ahlstrøm*[1], Dirk van As[1], John Peter Merryman[2], Michele Citterio[1], René Forsberg[2], Steen Savstrup Kristensen[2], Signe Bech Andersen[1], Jørgen Dall, Lars Stenseng[2], Dorthe Petersen[3], Erik Lintz Christensen[2], Robert Schjøtt Fausto[1]

[1] Geological Survey of Denmark and Greenland (GEUS)
[2] National Space Institute, Technical University of Denmark
[3] ASIAQ Greenland Survey

*apa@geus.dk

The Greenland ice sheet has been losing mass at an increasing rate during recent years, raising political concern worldwide due to the possible impact on global sea level rise and long-term climate dynamics. The Arctic region as a whole is warming up considerably faster than the global mean; it is necessary to quantify these climatic changes in order to assess the potential consequences and to provide the decision-makers with a firm knowledge base. To cover this need, the Danish Ministry of Climate and Energy has launched the Programme for Monitoring of the Greenland Ice Sheet (PROMICE) in 2007, designed and operated by the Geological Survey of Denmark and Greenland (GEUS) in collaboration with the National Space Institute at the Technical University of Denmark and the Greenland Survey (ASIAQ). The aim of the programme is to quantify the annual mass loss of the Greenland ice sheet and track changes in the extent of the glaciers, ice caps and ice sheet margin in Greenland.

The two main mechanisms responsible for the mass loss from the Greenland ice sheet are surface melting and iceberg calving. To quantify mass loss by melting, a network of automatic mass-balance stations is being deployed in the low-lying, melting part of the ice sheet, covering every distinct regional climatic zone around Greenland. Each station measures the climate factors causing melt as well as the subsequent local mass loss. In the past two years ten stations have been deployed by PROMICE, filling the gaps between the more than 20 additional Danish, US and Dutch stations already in place on the Greenland ice sheet and independent glaciers. Two more stations are scheduled to be deployed in summer 2009, followed by two more in 2010 completing the geographical coverage. Once in place, the observations will feed into a melt model to calculate surface melt – and changes/trends on a year-to-year timescale – over the entire ice sheet. The measurements will also provide a direct way to check the performance of climate models over Greenland, potentially improving accuracy of future-climate scenarios. Currently, live observations are transmitted to serve as input for weather predictions, as well as to inform people of the Greenland climate worldwide.

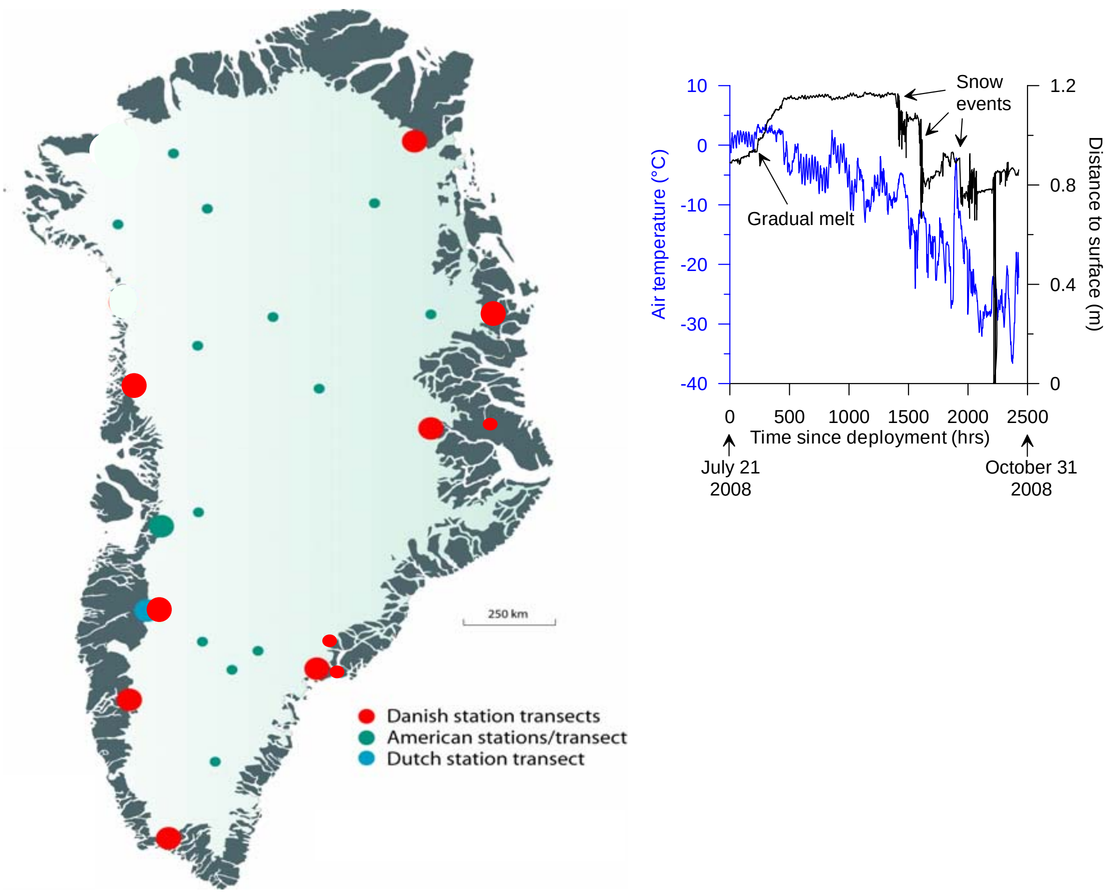


Figure 1: *Left: Map showing positions of currently active automatic weather stations on the Greenland ice sheet. Right: A sample of transmitted data from a station in Kronprins Christian Land, Northeast Greenland, showing air temperature (blue) and distance to surface whether ice or snow (black).*

The mass loss by iceberg calving is obtained from airborne surveys and satellite observations. A combination of the two allows us to determine the thickness and flow speed of the ice along the ice sheet margin, giving us the flux from the inland towards the ocean via Greenland's outlet glaciers. The first airborne survey was carried out in 2007 measuring ice sheet elevation by laser altimetry and thickness by ice-penetrating radar along the entire ice sheet margin. The next survey is scheduled for 2011. Repeating the effort every few years will show the dynamic response of the ice sheet to changing climate conditions.

In order to produce geocoded velocity measurements, a Synthetic Aperture Radar data processing software is currently under development at the National Space Institute. The processing chain implements coherence- and intensity-tracking of SAR images as well as differential SAR interferometry techniques. The core processing modules are provided by GAMMA Remote Sensing and Consulting AG, with additional functionalities developed at the National Space Institute, concerning error prediction, fusion of measurements from multiple satellite tracks and combination of offset-tracking and interferometric techniques. First results for Nioghalvfjerdsfjorden Glacier in Northeast Greenland is shown in Figure 3, utilizing ERS-1 and ERS-2 data from 1996.

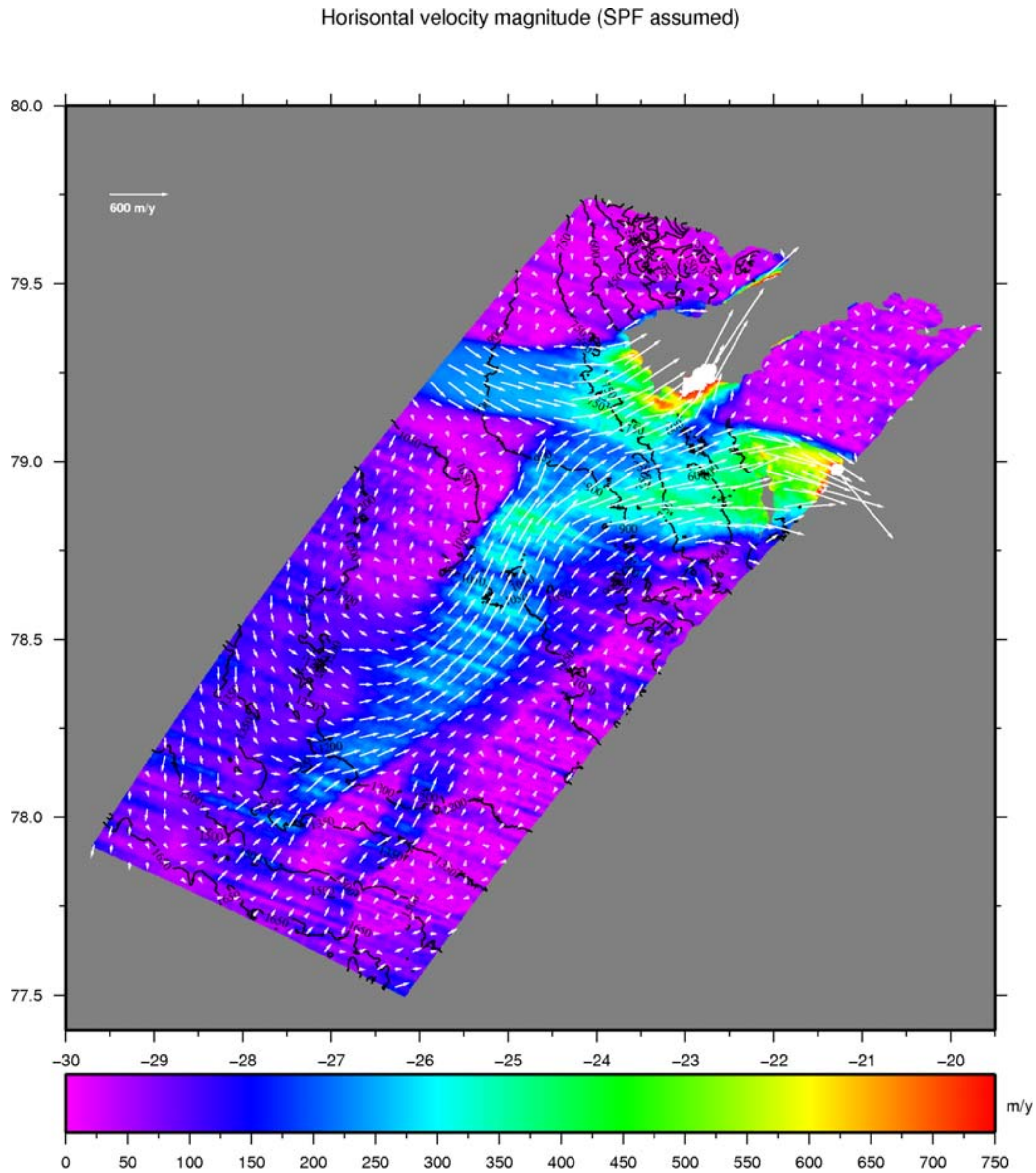


Figure 2: *ENU (east-north-up) velocities computed from InSAR double-difference line-of-sight velocity and coherence-tracked azimuth velocity. Height contours derived from InSAR double-difference. Surface parallel flow assumption (and height gradients from InSAR double-difference) used to relate UP velocity component to horizontal ones. Axes show latitude and longitude, arrows and colour scale indicate computed velocities.*

Investigating the calving rate variations of Kronebreen, Svalbard, using terrestrial photogrammetry and direct observations

Anne Chapuis[1*], Etienne Berthier[2] and Cecilie Rolstad[1]

[1]Department of Mathematical Sciences and Technology, Norwegian University of Life Sciences (UMB).
[2]CNRS, LEGOS, Toulouse, France.

[*]anne.chapuis@umb.no

The processes that control calving of tidewater glaciers are still poorly known even though they represent essential boundary conditions for glacier modeling. So far, long-term calving rates are determined using several laws which relate calving rates to independent variables such as water depth, ice velocity and stretching rate. But those laws only applies for annual variations. Indeed the knowledge of what controls short-term calving rates i.e. daily or hourly calving rates is very limited.
In this study we have led a multi-disciplinary approach on Kronebreen, Svalbard, to document what controls the calving rates of this glacier at different time-scales ranging from minutes to year. Each time-scale requires a specific method of investigation.

Seasonal calving rates

We used Formosat images to get data on both the front position fluctuation and the glacier veolicty. The calving rate displays a seasonal increase starting in mid-April and continuing until mid-september. Figure 1 shows the temporal and spatial pattern of calving rate throughout the melt season of 2007. The calving rate is highly spatially variable with doubling of calving rate at some places where the glacier lies in deeper water. Average glacier velocity is also an important factor that determines the calving rate seasonal fluctuations.

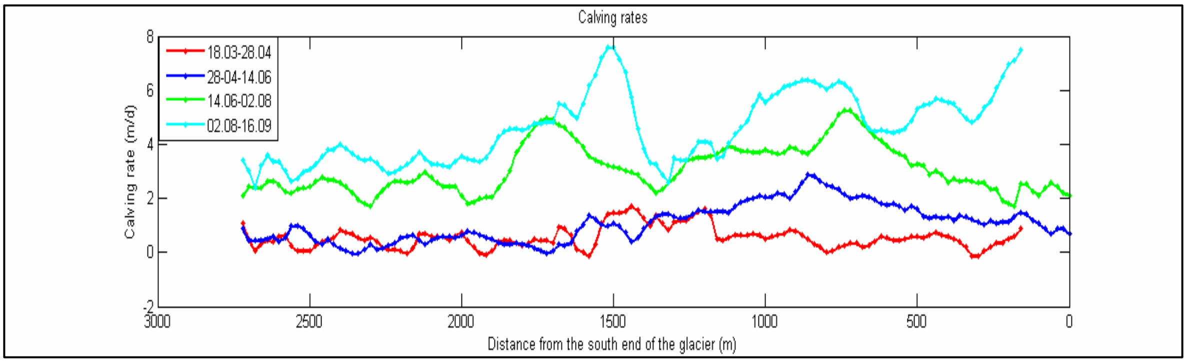


Figure 1: *Calving rate variations at the front of Kronebreen from March to September 2007.*

Frequency of calving events

Frequency of calving events has been monitored by direct visual observations of the calving front during a one-week period at the end of August 2008. We investigated several possible parameters that could influence the calving process such as tides, air temperature, precipitations, velocity fluctuations, wind speed, wind direction and direct sun exposure of the calving front. Figure 2 shows the overall very good agreement between air temperature and calving event frequency suggesting a strong control of the air temperature changes on the calving frequency.

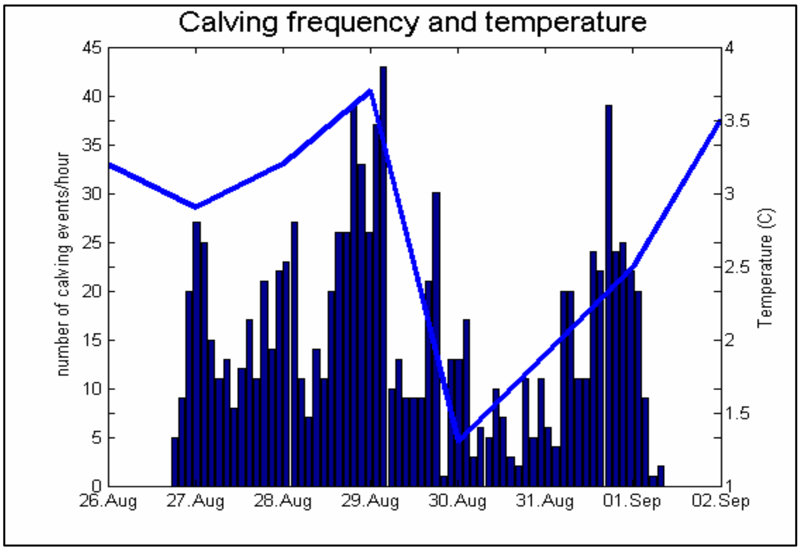


Figure 2: *Calving event frequency and air temperature for one week of observations at the end of August 2008.*

Glacial-Marine Interactions at the Belcher Glacier Terminus

Brad Danielson and Martin Sharp

Department of Earth & Atmospheric Sciences, 3-15 H.M. Tory Building, University of Alberta, Edmonton, AB, T6G 2E3, bdd@ualberta.ca

The Devon Island Ice Cap covers 14,000 km2 and contains 4110 km^3 of ice (Burgess and Sharp 2004) and is among the largest ice caps in the Canadian Arctic. The Belcher Glacier is an outlet glacier that drains ice from the ice cap interior to the ocean. It is the largest single source of iceberg calving, accounting for ~17% of the total mass loss from the Devon Ice Cap (Burgess et al., 2005). The glacier terminates in approximately 300m of water, and the ice cliff rises 20-40m above the water surface. While the terminus position has remained stable since 1960, the glacier appears to be thinning on average ~0.35m per year, but thickening near the terminus (Burgess and Sharp 2008). The glacier surface mass balance alone cannot explain this. This suggests that the glacier has accelerated in the recent past, and dynamic processes make a contribution to its thickness change and mass loss. Tidewater glacier systems such as the Belcher possess the potential for sudden changes in flow dynamics that may cause abrupt changes in mass balance. These systems must be studied in order to better assess their dynamic sensitivity to climate and their potential contribution to sea level change, which has not been specifically accounted for in IPCC AR4 predictions of global sea level rise.

In order to study the sensitivity of the flow dynamics of this glacier, observations of flow velocity, local meteorology, and surface hydrology were collected continuously over the 2008 summer. Five time lapse cameras were used to monitor surface hydrologic activity such as the filling and drainage of lakes, the development of surface drainage systems, and location of sites where surface melt water can enter the englacial drainage system. A sixth camera was used to observe the calving terminus. Continuously operating dualfrequency GPS stations were installed along the glacier centerline. Data spanning May through August were collected from three sites, and have been processed in kinematic mode using reference data from a differential base station. One of these stations was installed within 500m of the calving front, and reveals the changes in flow velocity of the terminus region. Air temperature data were collected from automatic weather stations and Hobo temperature loggers, and can be used to indicate when melt is occurring on the glacier surface.

Time-lapse photography of the calving front from May - August of 2008 reveals the timing of sea-ice removal from the bay, and episodic purges of sediment laden water into the sea sometime after major melt drainage begins up-glacier. Review of the photosequence has led to the three questions addressed in this study:

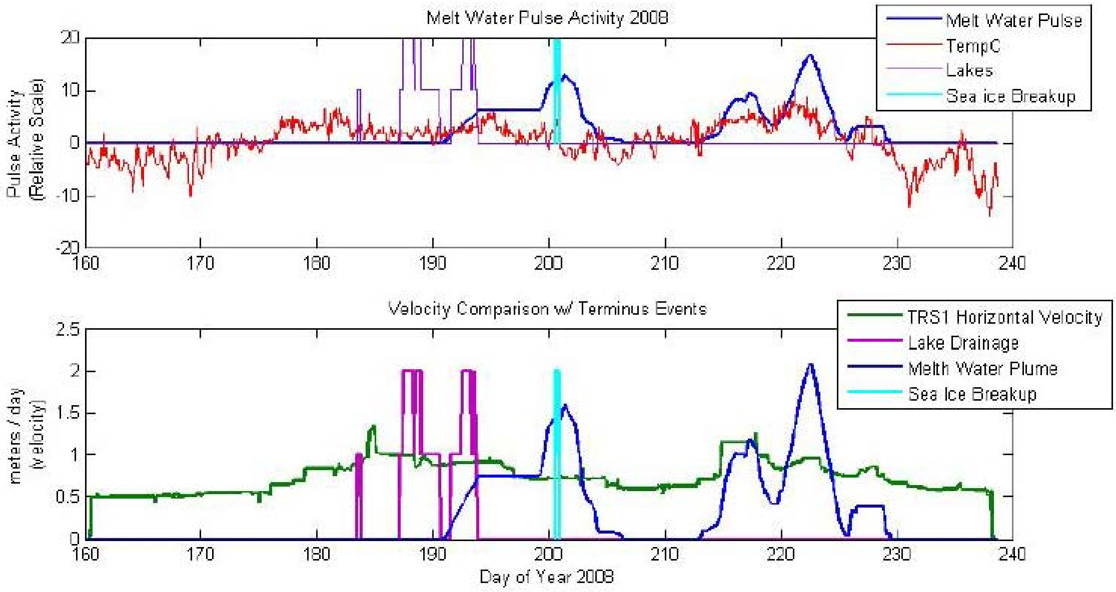
1. Does the melt water purge activity have an impact on horizontal ice flow velocity in the terminus area?
2. Does the glacier terminus accelerate when the sea ice breaks out?
3. Is horizontal or vertical motion of the glacier modulated by the ocean tide?

A frame-by-frame review of the terminus time-lapse sequence shows that there appear to be 2 large episodes of melt water purging. Based on the area of water affected by the melt water plume in each image, manual image classification was performed using a 0-5 relative scale to produce a time series of Melt Water Plume (MWP) events. The first large MWP was seen after ~15 days of strong melt (inferred from air temperature) had occurred, and ceased when air temperatures dropped below 0˚C during a mid-summer storm. The MWP activity increased rapidly with warm temperatures that followed in early August, and ceased completely at the end of the 2008 summer, around mid-August.

Examination of the photo sequences from the lake-observation time lapse cameras upglacier reveal that lakes formed on the glacier surface very quickly after the initiation of warm temperatures in late June - early July. One lake was observed to drain from full to empty in 3 hours, and the lack of a visible surface drainage channel suggests it drained englacially. Other lakes further up-glacier drained across the surface and the resulting floods are suspected to have eventually reached moulins in the mid-glacier region.
Comparing the timing of lake drainage events to the MWP time series reveals that the first MWP event immediately follows the sequence of lake drainages, providing a good indication that the sediment-rich water seen exiting the glacier marine terminus is surface melt water that has drained through subglacial channels.

The horizontal ice velocity at the terminus roughly parallels the air temperature curve. The ice was flowing fastest during the period when surface lakes were draining, slowed after the first large MWP released subglacial water at the marine terminus, and remained slow during the mid-summer cold period. Ice velocity increased again in early August, and closely followed both the air temperature and MWP fluctuations, suggesting that the englacial drainage system had become efficient at transporting melt water from the surface to the bed, and that glacier motion was responding to subglacial water pressure fluctuations. The MWP activity seen at the glacier terminus likely represents a release of subglacial water and thus a decrease of the water pressure that allows enhanced sliding at the glacier bed.

Sea ice removal appears to have less effect on glacier flow speed. Examination of the terminus time lapse sequence reveals that the sea ice showed visual signs of becoming mechanically weak between July 6-16. During this time, large leads formed, melt water ponds grew over the sea ice, and large floe chunks began to move independently in the tidal and surface currents. Areas of sea ice began to move appreciably when the first evidence of MWP activity appeared. However, all of this followed the first major acceleration of the glacier terminus, and occurred near the peak summer flow speed. By the time the sea ice collapsed and completely floated away from the calving front on July 18, the horizontal flow speed of the glacier terminus was decreasing with the rapid expansion of the MWP. It is not possible to ascertain the extent to which the landfast sea ice may have restrained the flow of the glacier terminus.



Several recent studies have shown strong velocity modulation by tidal swell on Antarctic glaciers and ice streams. These examples (Bindschadler, Whillans, Rutford Ice Streams) all have extensive floating ice tongues (Adalgeirsdottir et al., 2008). The lack of a floating ice tongue at the Belcher Glacier is likely to limit this effect, but tidal fluctuations in water column height at the calving front could induce a vertical (flotation) motion of the ice, and could also affect sub-glacial water pressure, thereby influencing horizontal motion. In order to study this potential effect, tidal amplitudes for the study period were generated with the Webtide tidal prediction model (Canadian Department of Fisheries and Oceans), and verified using a short recording of local tide level in July 2008. Unfortunately, the question of whether tidal forces affect the vertical or horizontal motion of this glacier remains unanswered at this point in the study. Kinematic processing artifacts revealed in the ice motion data series obscure potential motion signals at the periodicity of tidal fluctuations, and need to be removed before an definitive answer can be obtained.

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Changes in Surface Elevation of the Kaskawulsh Glacier, Yukon Territory

Norah Foy and Luke Copland

Department of Geography, Laboratory for Cryospheric Research, University of Ottawa, Ottawa, Ontario K1N 6N5, Canada. Email: luke.copland@uottawa.ca

The Kaskawulsh Glacier is located in Kluane National Park, in the St. Elias Mountains, SW Yukon (60°44'20"N, 138°55'53"W). The glacier has an area of ~900 km^2, is ~70km in length and flows generally northeast. The annual temperature trend in the Yukon shows warming of ~2°C since 1948 (Environment Canada 2008). Previous studies indicate that melting of glaciers in the Yukon/Alaska region has increased coincident with this warming; between the 1950s and 1970s, they thinned by an average of 0.52m yr^{-1}. From the 1990s to 2001, this thinning had increased to 1.8m yr^{-1}. This change amounts to a water loss of 96 km^3 yr^{-1} between 1995 and 2001 (Arendt et al. 2002). More recent work indicates that these losses are likely greater than previously calculated (Arendt et al. 2006; Larsen et al. 2007). The glacier changes over the Gulf of Alaska region are corroborated by data from the GRACE satellite, which indicates a water equivalent loss of 84km^3 yr^{-1}(+/- 5km^3) between 2003 and 2006 (Luthcke et al. 2008). However, snow accumulation records since 1736 from an ice core at Mt. Logan indicate increases in snow accumulation, particularly between 1976 and 2000, which saw an increase of 0.12m $decade^{-1}$ (significant at the 99% level; Moore et al. 2002). The Mt. Logan coring site, however, was located at 5340m, some 2km higher than the accumulation area of the Kaskawulsh Glacier, meaning that it may not be representative of changes in local conditions.

Using non-scanning LIDAR, Arendt et al. (2002) showed that between 1950 and 1995 the Kaskawulsh Glacier thinned by ~1.5m yr^{-1}, and that by 1995-2001 the rate of thinning had decreased to ~0.5m yr^{-1}. The present study expands upon these findings by extending analysis to 2007, as well as extending spatial sampling along the Kaskawulsh Glacier to cover the entire central arm. To map the changes, digital elevation models (DEMs) for 1977 and 2007 are compared. The source of each DEM used, as well as the resolution and vertical error associated with each data type, are detailed in Table 1. When two different DEMs are compared the vertical error is cumulative. Thus the cumulative elevation error is +/- 19m below the ELA, and +/- 49m above the ELA.

Type	Year	Resolution	Maximum Vertical Error
CDED DEM	1976-1977	1:50,000	+/-15m below the ELA (Larsen et al. 2007); +/- 30-45m in the accumulation area (Echelmeyer et al. 1996; Larsen et al. 2007)
Scanning LIDAR profile	2007	0.18m	+/- 4m (Chris Hopkinson, personal communication)

Table 1. *Data types and associated error*

The CDED DEM was created from stereo aerial photographs, and covers the glacier in its entirety. The scanning altimetry profile was flown in two segments in August 2007 by the C-CLEAR consortium using an Optech airborne LIDAR system. The individual laser returns were interpolated to create a DEM covering the entire central arm of the glacier. Changes in elevation observed along the profiles of the central arm are assumed to be representative of changes across the glacier as a whole. Furthermore, calculations of changes in volume do not account for any internal accumulation or changes in ice density.

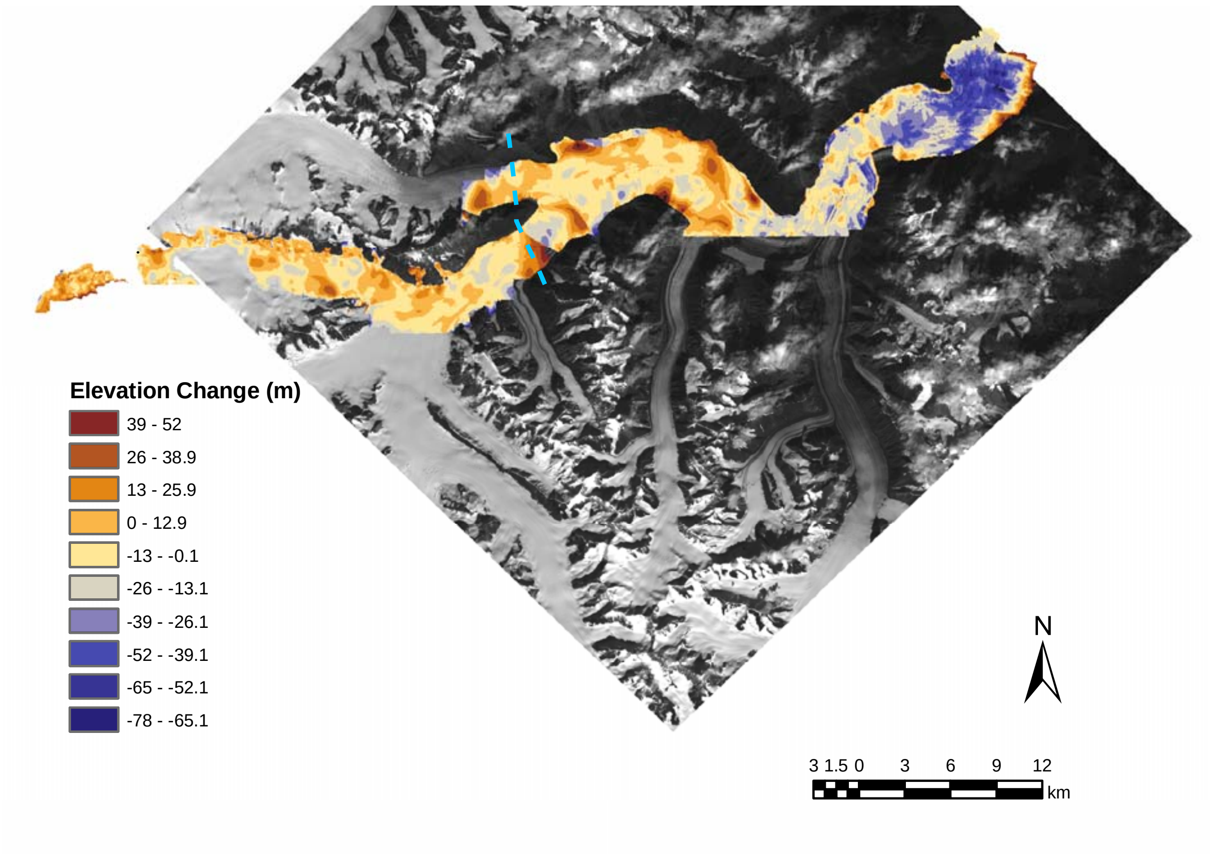

Figure 1. *Surface elevation changes of the Kaskawulsh Glacier (1977-2007)*

Between 1977 and 2007 the ablation zone thinned by an average of 10.9m (Figure 1). The thinning is greatest in the central portion of the terminus, where it reaches a maximum of 71.2m. In general, thinning decreases with increased altitude. In the accumulation zone thinning is more modest, and average thinning of 1.0m is observed. Some areas along the glacier do experience thickening of up to 48.5 m. Patches of thickening ice occur above 1470m, in the upper portions of the ablation zone and in the accumulation zone. One such area is located on the southern margin of the glacier, just below the ELA, downglacier of a surge-type tributary glacier. Therefore, the large increase in surface elevation at this location may be related to a surge event on the tributary. Another patch where thickening occurs is on the northern margin of the central arm, where there is a periodically draining ice marginal lake (Johnson and Kasper 1992). The marked change in elevation at this location may be due to a difference in timing of the drainage cycle captured by each DEM. Overall the mean thinning on the Kaskawulsh Glacier between 1977 and 2007 is 6.3m, which translates to an average rate of thinning of 0.21m yr^{-1}.

Therefore, over the past 30 years, warming in the Yukon is coincident with thinning of the Kaskawulsh Glacier.

Acknowledgements

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Importance of topographic and geolocic conditions for behaviour of Spitsbergen glaciers retreating from the sea to land

Grzegorz Gajek[1], Marian Harasimiuk[1], Jacek A. Jania[2], Mariusz Grabiec[2], Dariusz Puczko[3], Piotr Zagórski[1]

[1]Institute of Earth Sciences, Maria Curie-Sklodowska University, Lublin, Poland;
[2] Faculty of Earth Sciences, University of Silesia, Sosnowiec, Poland;
[3] Institute of Geophysics, Polish Academy of Sciences, Warszawa, Poland

Introduction

Contemporary evolution of Spitsbergen ice masses is leading to retreat of glaciers from the sea to land. The objective of this study is to examine the role of subglacial topography and geology for filtering of climatic signal expressed by behavior of tidewater glaciers in Spitsbergen conditions. Glaciers receded from the sea to land could be indicators for future evolution of glaciers under warmer climate conditions. Studies of Renardbreen and Recherchebreen have been conducted within the frame of the Polish segment of the IPY GLACIODYN project.

Both glaciers are located in the northwestern Wedel Jarlsberg Land, Southern Spitsbergen (Fig. 1). They are polythermal, valley type glaciers. Renardbreen is one of a dozen of Svalbard tidewater glaciers retreated from the sea to land during the last three decades (Blaszczyk, Jania, 2008). Recherchebreen was classified as the tidewater glacier by Hagen et al. (1993) and it is recently of the same type (Fig. 2), whereas, in 1936 and in 1960, its front terminated on narrow belt of land separating internal tidal bay (called here “lagoon”) from the fiord waters. Recherchebreen is the surge type glacier.

Methods and data

Topography of Renardbreen surface and the Recherchebreen lower reach is known from topographic maps based on aerial photos from 1936 (NPI), 1990 (Zagorski, 2005). Recent information on the Renardbreen surface topography has been obtained by the kinematic GPS survey done in spring seasons of 2005-2008. Present terminus position of Recherchebreen was surveyed by the kinematic GPS method in August 2008. Radio echo sounding of the glacier thickness was done by the MALÅ Geoscience Ramac GPR of 25 Mhz. Bathymetry of marine forefields of both glaciers was obtained as a result of a survey from a small motor boat by the Lowrance echo sounder in July and August 2008.

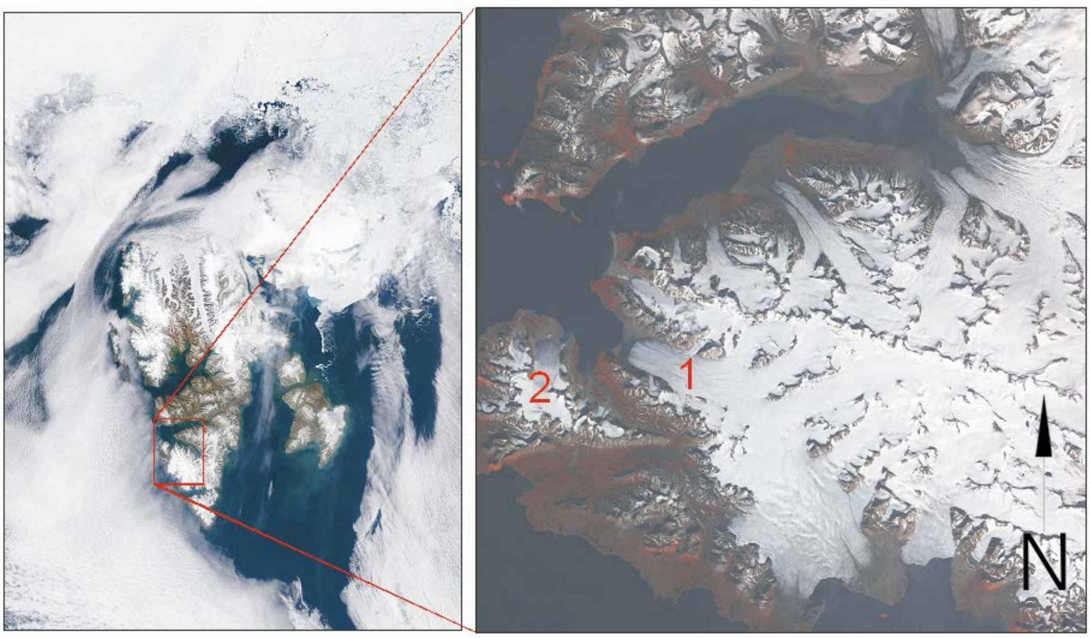

Figure 1 *Location of studied glaciers in Svalbard (MODIS image) and on NW Wedel Jarls berg Land (a portion of the Terra/ASTER scene of 23 July 2005): 1-Recherchebreen, 2-Renardbreen*

Figure 2. *Forefield of Recherchebreen (Photo by J. Jania, 7 August 2008)*

Tectonics and relief of the study area

The Recherche Fiord region is composed from several tectonic blocks (Birkenmajer, 2004, 2006). Their borders are dislocations of different importance and age. The protherozoic Renard Block is the largest tectonic unit in the area with a complex internal structure with three distinct overthrusts (Fig. 3).

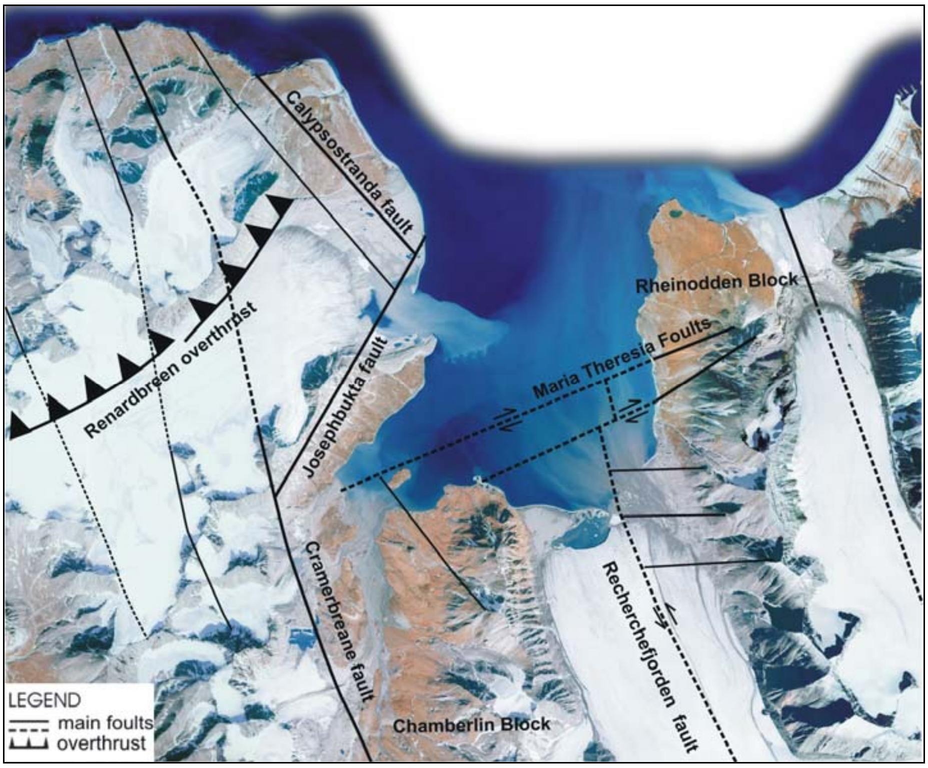

Figure 3. *Simplified tectonic map of NW part Wedel Jarlsberg Land (modified from Birkenmajer, 2004, 2006). Base map – the ortophoto map by Zagorski, (2005).*

The majority of superficial and subglacial relief elements are related to tectonics of the area. Faults are creating local horsts and grabens perpendicular or oblique to the flow direction of both glaciers (Fig. 4). They are also clearly visible in the subglacial topography of Renardbreen and in relief of subareal and subsea marginal zones of both glaciers (cf. Fig. 2 and Fig. 4).

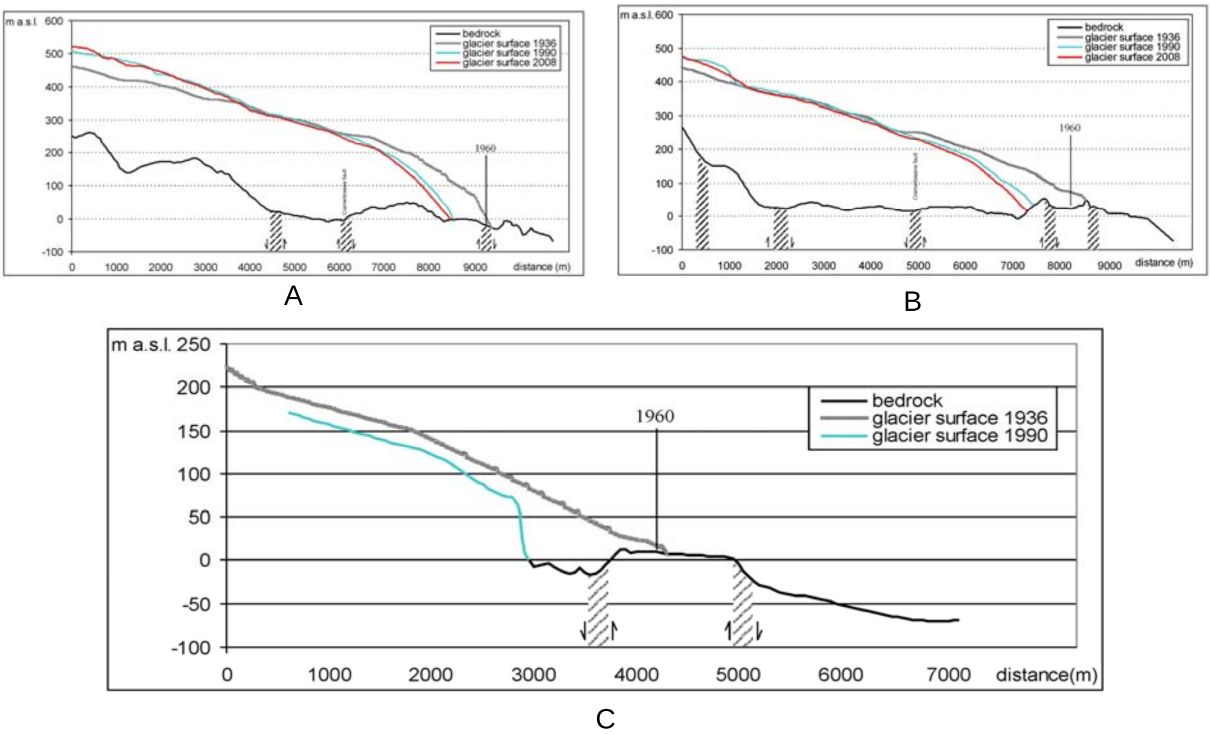

Figure 4. *Longitudinal profiles of surface and bedrock of studied glaciers (cf. Fig. 6 and Fig. 7): A - Renardbreen north profile, B - Renardbreen south profile, C – profile of the Recherchebreen lower part. Fault zones in the bedrock are marked*.

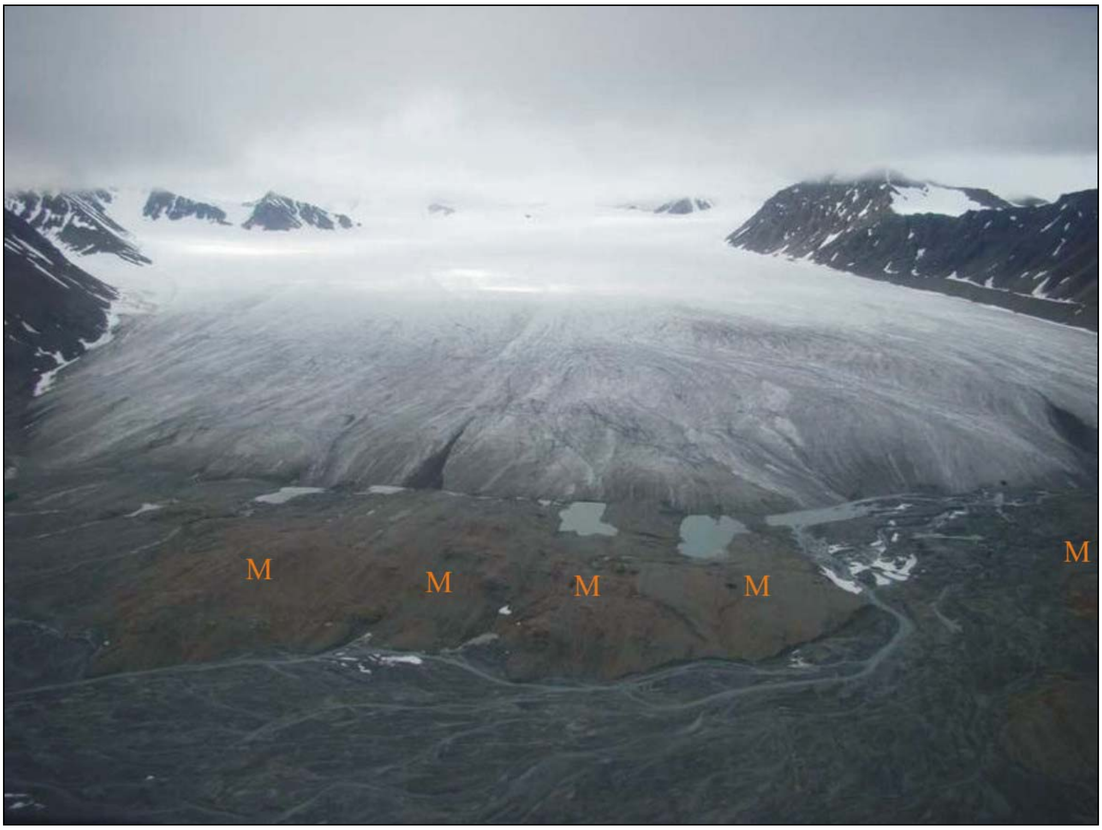

Figure 5. *Forefield of Renardbreen with line of roche-moutonées (M) developed along the horst zone. Humps are covered by a thin layer of fluted moraine (Photo by G. Gajek , 10 August 2008).*

Renardbreen valley is steeper and depressions in its floor are rather shallow, bed of Recherchebreen is less steep and has very distinct threshold on the mouth separating the inner lagoon from Recherche Fiord. The Recherche Lagoon is deeper (max. depth 58 m) than Josephbukta (max. depth 39 m) on the forefield of Renardbreen (Fig. 6 and Fig. 7).

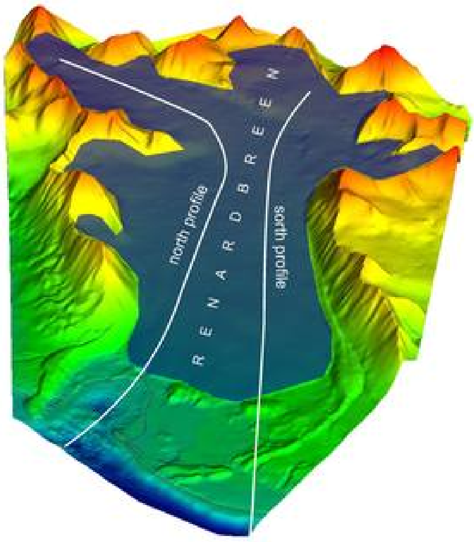

Figure 6. *Perspective view on Renardbreen based on the DEM. Longitudinal profiles are marked (cf. Fig. 4).*

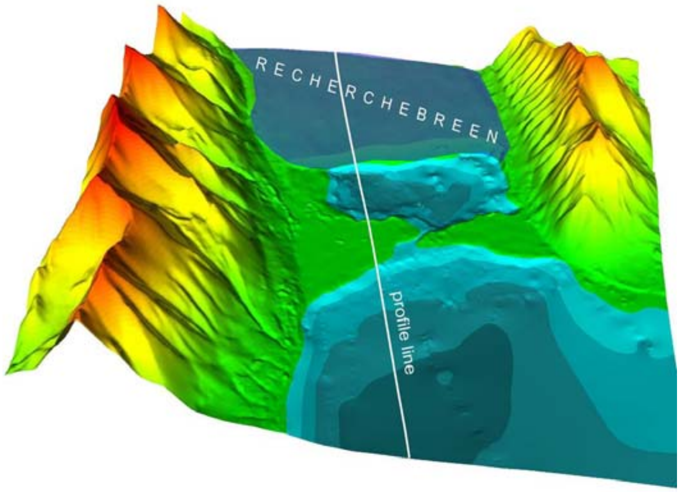

Figure 7. *Perspective view on the Recherchebreen front with the land and subsea forefield. Profile line is marked (cf. Fig. 4).*

Results of the Soviet airborne radar sounding of Recherchebreen (Macheret & Zhuravlev, 1985) suggest that its bedrock lies below sea level up to ca. 10 km upstream from the front. It makes a distinct difference to bedrock topography of Renardbreen, which is steeper and has small depressions with bottom lying below the sea level, but isolated from connection to the sea.

Bedrock Topography and fluctuations of tidewater glaciers

Interaction of a surging tidewater glacier with morphology of its fiord forefield can be considered for Recherchebreen. The maximum Little Ice Age extend (Fig. 8) of the glacier was a result of the surge dated back around 1838 (Liestøl, 1969).

Figure 8. *Changes of Recherchebreen terminus position*

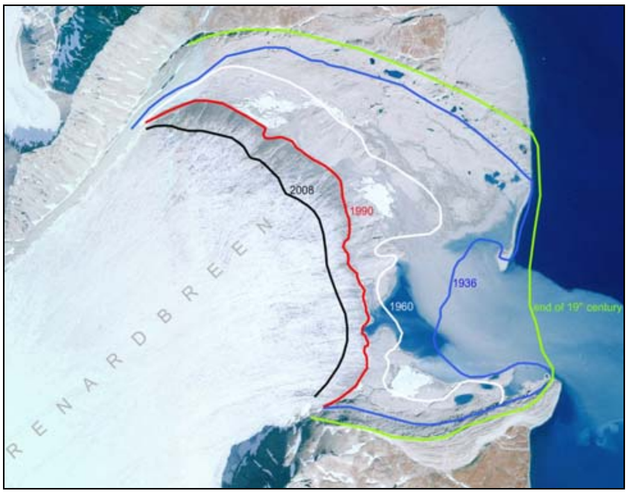

Figure 9. *Changes of Renardbreen terminus position*

One can reconstruct occurrence of intense calving afterwards, caused retreat of the front to shallow water over the tectonic horst (cf. Fig. 8). Recession diminished then from 3-4 dozens of meters per year to less than 10 m yr^{-1}. Further melting resulted in retreating of the western part of glacier front to deeper water in the contemporary lagoon (position in 1936) in an average rate of ca. 36 m yr^{-1}. The next surge event in the forties resulted probably in overpassing of the horst to unknown distance and deposition of sediments with some dead ice over the threshold. An advance of the front by an average rate 10 m yr^{-1} was noted between 1936 and 1960. In 1960, glacier front was situated on land. Subsequent retreat into relatively deep waters of the Recherche Lagoon continuing up to nowadays with increasing intensity (13 m yr^{-1} in 1960-1990 and 50 m yr^{-1} in 1990-2008). Glacier flow velocity in the quiescent phase was very low less than 15 m yr^{-1} (Blaszczyk et al. 2009). For this reason, despite the contact with the sea water in the lagoon. Currently, calving is very rare and mass loss due to calving is low in comparison to superficial ablation.

There is no direct evidence of surge of Renardbreen. The maximum extend of its tongue was reached in the end of the 19^{th} century (Fig.9). Distinct fluted moraines noted in the inner marginal zone together with transverse mini-ridges (reproducing probably an infill of the bottom crevasses) suggest an active surge episode sometime in the past. Retreat of the front from the land margin and on the Josephbukta waters was rather slow (12 m yr^{-1} and 8 m yr^{-1}) for the both periods 1936-1960 and 1960-1990 respectively. After complete retreat of the terminus to land, the recession rate was similar (10 m yr^{-1}). An average recession rate increased to 14 m yr^{-1} after 2005 (Zagorski et al., 2008). The glacier snout retreated from the line of roche-moutonée elevations after 2000. It is worth to note low flow velocity of Renardbreen of ca. of 11 m yr^{-1}, measured in the period 2005 - 2008. One can assume that mass loss by calving in time of the glacier retreat from Josephbukta contributes little to overall mass balance. Both are characteristic for the quiescence phase of the surge cycle. Therefore, recession rate of the marine margin front was not dramatically different from noted for the land margin one. Such prerequisites are leading to statement that Renardbreen had also undergone active surge before 1936. Exceptionally high and wide southern ice-cored moraines could be treated as an additional indirect evidence of a surge event.

Conclusions

Studies of fluctuations of two surge type tidewater glaciers in Spitsbergen lead to couple of conclusions.

- Shallow submarine zones or thresholds lying over the sea level are important factors influencing fluctuation of the front position of tidewater glaciers. They stabilize position of ice-cliffs due to limited calving intensity in the shallow water. Such well known relation filters climate signal recorded from fluctuation of glacier front position.
- Retreat of the tidewater glaciers from the sea to land depends strongly on presence of depressions in their valley bottom lying below the sea level and linked to fiord waters. Renardbreen retreated effectively due to higher altitudes of their bedrock. In case of Recherchebreen retreat was temporary on the area of the structural threshold.
- Retreat of tidewater glacier from the sea to land during quiescent phase, when glacier flow is very slow and calving rate insubstantial, is driven by the intensity of superficial ablation. Transition from the calving type glacier to the land based one is smooth without significant perturbations for the glacier mass turnover due to significant reduction of mass loss by calving in the overall mass budget. Renardbreen is a good example of such a gentle transformation.
- Bedrock topography and dynamics of tidewater glaciers are modifying climatic signal coming to the glacier system and demonstrated by glacier front fluctuations. For easier monitoring of response of Spitsbergen glaciers to climate warming tidewater glaciers emptying into fiords of relatively uniform depth and being in a quiescent phase of surge (when calving intensity is low) have to be selected.

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Radar Results from the Devon Ice Cap

L. Gray[1] and D. Burgess[2]

[1]Canada Centre for Remote Sensing.
[2]Geological Survey of Canada.

Results from 2006 and 2008 ASIRAS spring over flights of the Devon Ice Cap are discussed and compared to surface data acquired with a high bandwidth (5-18 GHz) radar[1] based on a network analyser. As others have shown elsewhere, the 2006 airborne ASIRAS altimeter results showed that often the dominant scattering surface was not the surface but corresponded to the buried denser melt layers from the previous summer and fall. Further, closer to the ice cap summit multiple layers could be distinguished. These results pose the question; while the ASIRAS altimeter data can under some circumstances map spatial accumulation variation, could this be done with CRYOSAT?

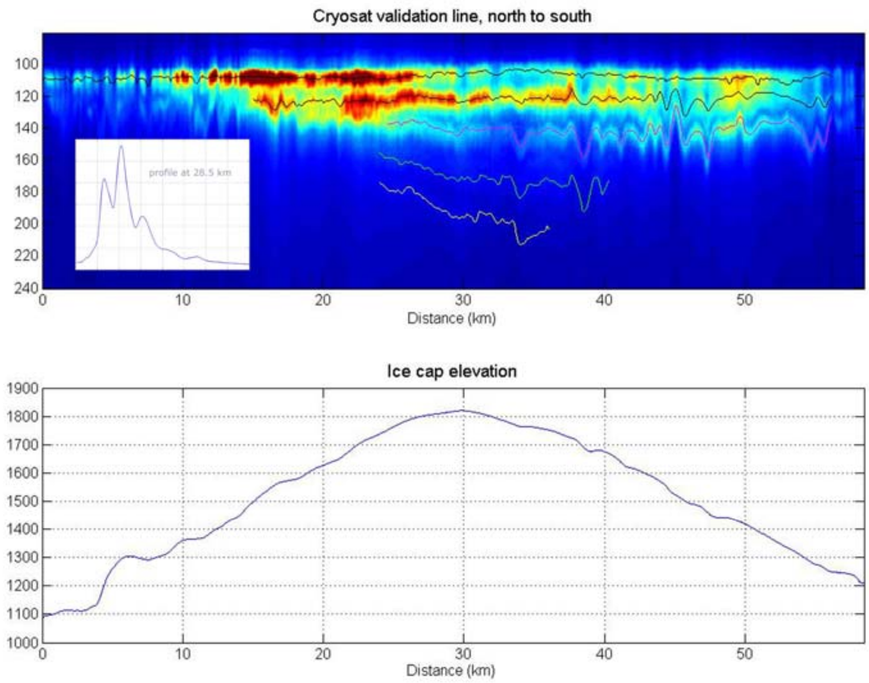

Figure 1. *ASIRAS results for the north-south CRYOSAT Cal/Val line across the Devon Ice Cap. The upper plot illustrates the relative variation in returns with position along the line and depth beneath the surface. The lower curve shows the ice cap elevation.*

[1] In 2008 the radar was borrowed from the UK (thanks BAS, UCL and U. Edinburgh) under the CRYOSAT programme and used for specific sites.

While the CRYOSAT altimeter has a lower resolution than ASIRAS, the more difficult aspect to simulate is the different geometry (wave front curvature and footprint size). To help understand how the wave front curvature can influence returns, some knowledge of the angular variation in backscatter is useful. We have developed code to process ASIRAS data to do this. However, this approach is limited particularly by the aircraft roll, and near-nadir results are only possible for regions of the flight line with low roll angles. Some results are illustrated in Figure 2 below for a region with strong specular reflection (in the northern region of the CRYOSAT Cal/Val line). ASIRAS results from close to the ice cap summit showed more distinct layering than that visible in Figure 1 above. The layering appeared to be related to summer melt layers. The strongest returns at the ice cap summit are from the previous summers melt layer, but neither the surface layer nor the first melt layer exhibited an angular variation as strong as the surface specular returns illustrated in light blue in Figure 2 below.

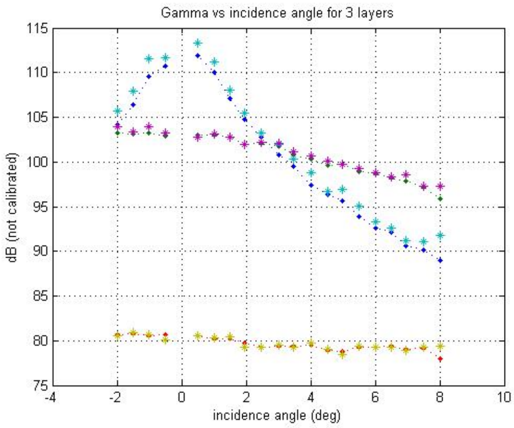

Figure 2. *The relative backscatter returns in dB from the northern section of the Cal/Val line in the region of strong specular backscatter (dark red surface returns to the left in the upper part of Figure 1). The surface returns (light blue in Figure 2) show the most variation with incidence angle, more than that from the previous year melt layer (purple) and much more than a deeper layer (yellow).*

The complexity of altimeter ice signatures can be illustrated further using results from the 5-18 GHz surface radar. In this case, the relatively broad beam antennas are ~ 1.4 m above the surface and subsurface nadir returns can be time-ambiguous with off-nadir returns from other layers, and even the surface. With this geometry one would expect that layering would be more readily visible close to the surface when the returns are pulse-limited, rather than beam-limited which is the situation for depths greater than ~ 50 cm. This is what we observe:

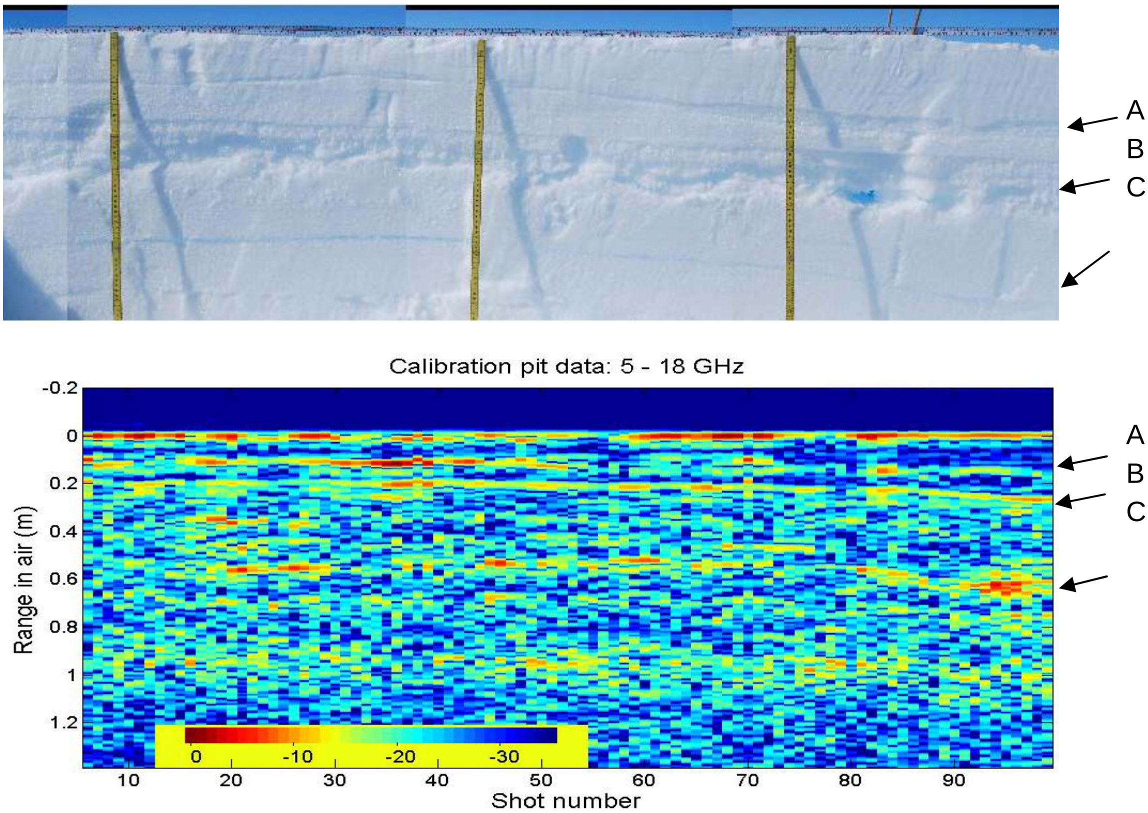

Figure 3. *The photo mosaic (top) covers the top two thirds of the 5 – 18 GHz radar results illustrated in the lower part of the figure. With this geometry and resolution the strongest and most distinct returns are from the surface and near-surface layers in the winter accumulation, not from the deeper ice layers.*

Four years of surface mass balance measurements on Austfonna Ice Cap, Svalbard

Jon Ove Hagen[1], Thorben Dunse, Trond Eiken, Geir Moholdt and Thomas V. Schuler,
Department of Geosciences, University of Oslo, Norway
[1]e-mail joh@geo.uio.no
Andrea Taurisano, Ola Brandt and Jack Kohler Norwegian Polar Institute, Tromsø

Austfonna (79.5°N, 25°E) at 8120 km^2 is by far the largest ice cap in Svalbard and one of the largest in the Arctic. Within the framework of CryoSat calibration and validation activities, annual field visits to Austfonna started in April 2004 and continued as a part of the IPY GLACIODYN project.

In spring 2004 a stake net of 20 mass balance stakes was deployed in profiles across the ice cap (Fig. 1). This network of stakes has been maintained since then. The stakes have been remeasured once a year, during the annual visits in springs 2005 - 2008. During the spring measurements the snow thickness at each stake is measured and thus we get the height of the stakes above the previous summer surface. From these measurements we derive the annual net surface mass balance at each stake location. Riming is a serious problem for all installations on Austfonna, so every year some of the stakes disappear probably because they break down due to heavy riming (Fig. 1).

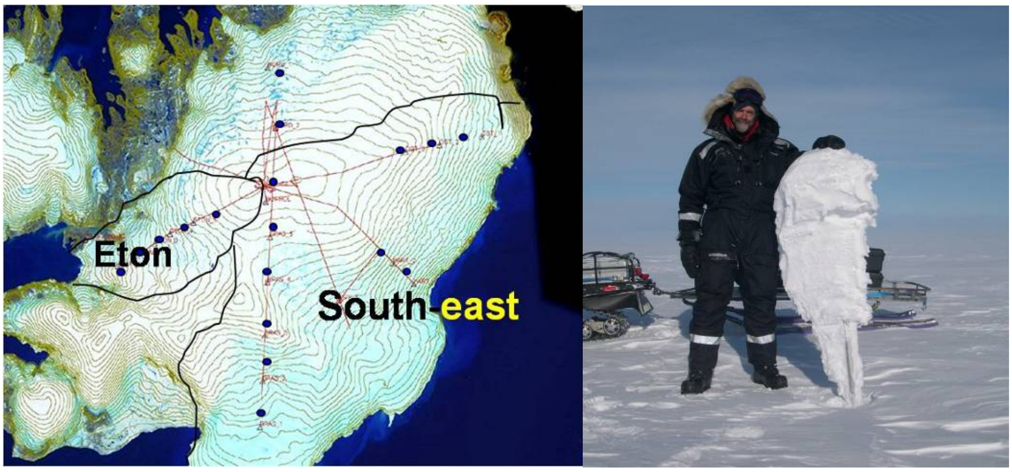

Figure 1. *Austfonna and the stake net marked with dots. To the right typical riming on the stakes is shown.*

The dominant precipitation direction for Austfonna is from the east. This explains the general pattern of snow accumulation across the ice cap, which shows a pronounced gradient from high values in the SE to lower ones in the NW (Pinglot and others, 2001; Taurisano and others, 2007). The distribution of snow thickness is measured using extensive GPR-profiling together with manual snow thickness probing, and the snow water equivalent is obtained from density measurements in snow pits. The winter snow accumulation varies from year to year by 100 %, but the pattern of the snow distribution is fairly stable (Fig. 2), often with about three times more snow in the south-east than in the north-west (Schuler et al 2006, Taurisano and others, 2007). The snow thickness is usually between 0.5m up to 3 m.

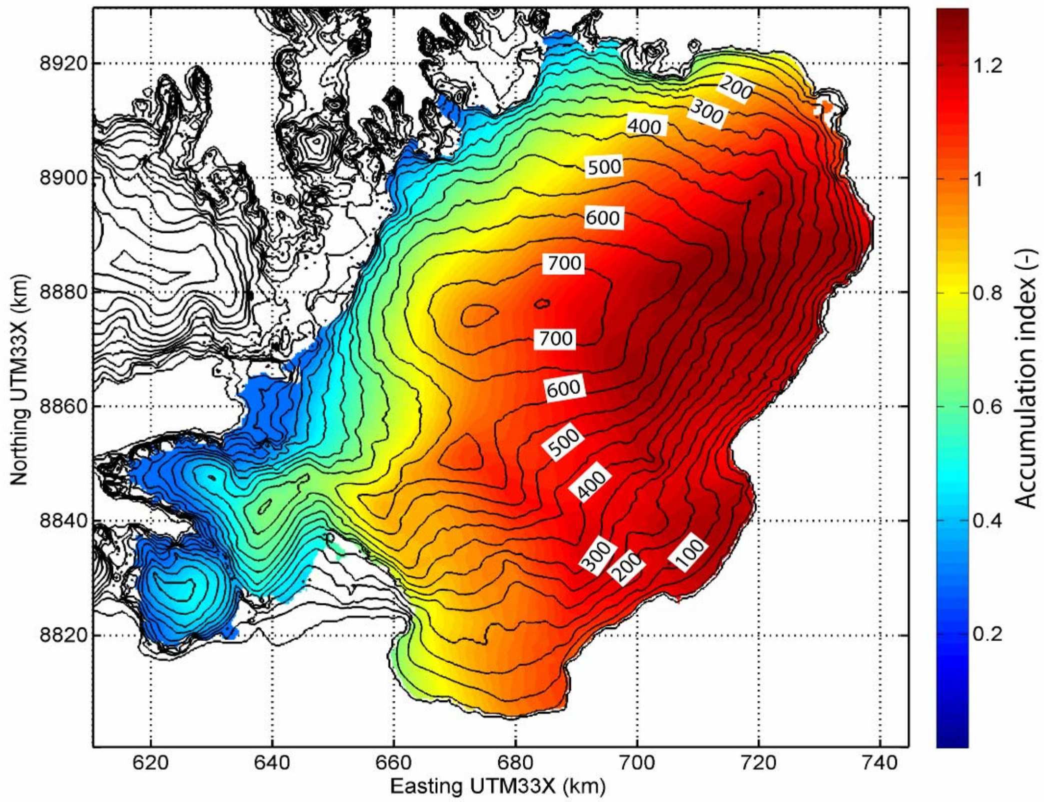

Figure 2. *Accumulation index map of Austfonna (from Schuler and others 2006)*

The average total winter mass balance (B_w) over the measured seasons 2004 – 2008 was calculated to B_w = 4,5 ± 1,5 km^3 water equivalent (w.e.) or a specific value of b_w = 0,55 ± 0,2 m w.e.. However, the inter-annual variation is larger than 100 %, and the snow accumulation varies from a minimum value of 0,5 m in west up to 3 m in east.

Only in one basin, Etonbreen, it has been possible to visit all stakes every year. This basin is draining to the south-west and covers an area of c. 630 km^2, so less than 10 % of the ice cap (Fig. 1). The measurements show a high year-to-year variability of mass balance. The winter of 2003/2004 received less snow than the following years and temperatures during summer 2004 were above average, resulting in a strong negative mass balance of b_n = -0,69 ± 0,1 m w.e. with the ELA at ca. 650 m a.s.l. The following years have received more snow and the minimum ELA for the period was in 2007 at ca. 400 m a.s.l. with b_n = 0,03 ± 0,1 m w.e. Mean specific net balance over the period 2004-2007 for Etonbreen was bn = -0,19 ± 0,1 m w.e.

Summarizing all stake measurements for the period 2004-2007 representing an area of ca. 60 % of the ice cap indicates a mean overall net surface mass balance close to or slightly more negative than on Etonbreen, with overall mean surface mass balance for the period 2004-2007 of B_n ~ -1,8 km^3/y ± 0,5 km^3/y w.e. or bn ~ - 0.22 m/y. The cumulative mass loss over these four years is then -7,1 km^3/y, however, -5,6 km^3 or 80 % of loss occurred in

2004, so for the last three years 2005-2007 the mean net balance was Bn~ -0,5 km^3/y bn = - 0,06 m/y. The yearly variations are large (Fig. 3).

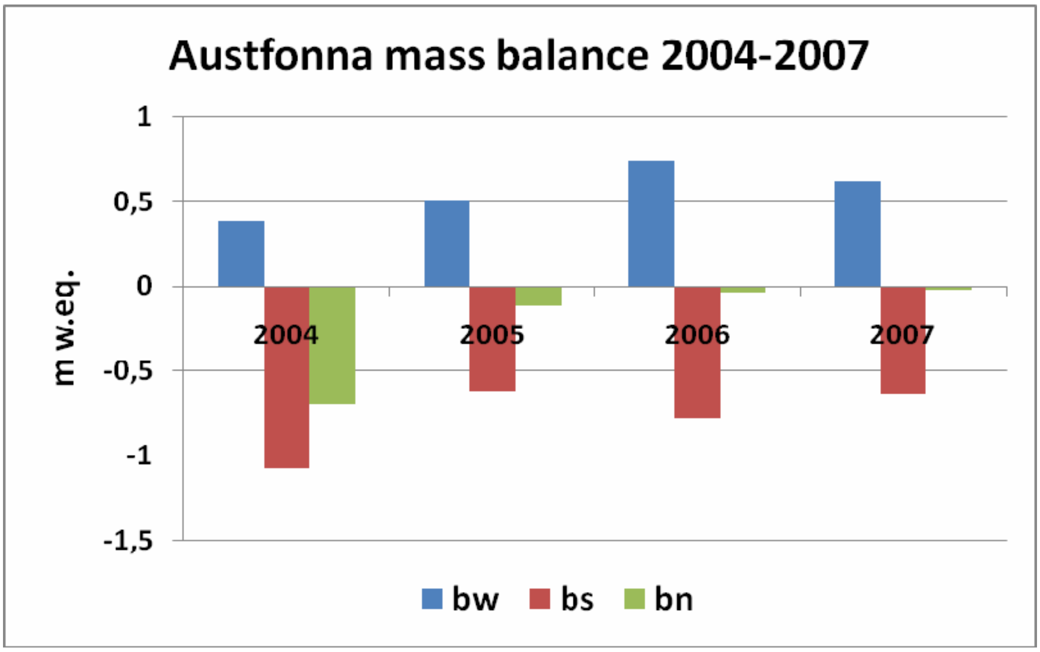

Figure 3. *Yearly specific surface mass balance on Austfonna 2004-2006.*

There is no sign of increased melting over the measured period, and the mean net balance is only slightly more negative than previous estimates form the period 1986 – 1999 when the net surface balance was calculated to be close to zero based on shallow core data (Pinglot and others, 2001)

Dowdeswell and others (2008) calculated the mean annual calving loss to be Mc ~ -2.5 km^3/y or -0.30 m/y. When this number is added to the surface mass balance the total overall net mass loss from Austfonna is clearly negative by Bn ~ - 4.3 km^3/y or bn ~ - 0.52 m/y ± 0,10 m.

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2001-2007 East Greenland valley glacier surge and its implications for surge mechanisms

Hester Jiskoot and Daniel T. Juhlin
Department of Geography, University of Lethbridge, Canada.

Introduction

About half of the glaciated area peripheral to the Greenland ice sheet, ~55000km^2, is located in central East Greenland (67º-72ºN) (Jiskoot et al., 2003). This is an area with very sparse glaciological research (Weidick, 1995). The region is highly sensitive to climate change, as: 1) About 2/3 of the glaciated area drains through tidewater glaciers (Jiskoot, 2002; Jiskoot et al., 2003), 2) 30-70% of the glaciers are of surge-type (Weidick, 1988; Jiskoot et al., 2003) and can suddenly speed up and cause extreme calving events (Jiskoot et al., 2001; Pritchard et al., 2005), and 3) Many Scoresby Sund glaciers are polythermal (Kirchner, 1963; and from icings) and climate change might affect their thermal regime and ice dynamic behaviour. Only 5 of the 71 glaciers with strong surge evidence in the region have a described surge, and an additional 108 glaciers have surge evidence that is equivocal (Jiskoot et al., 2003). One of these, an unnamed valley glacier (70°7'39"N, 26°56'42"W) that drains from the Geikie Plateau into Scoresby Sund, advanced 2.8 km during a recent surge (Fig 1). The glacier is relatively small (~21 km^2) and has at least one bulging tributary (Class III, R_a>1: Kargel et al., 2005). We unofficially name the glacier *Sermeq Peqippoq*, Greenlandic for "glacier that bends".

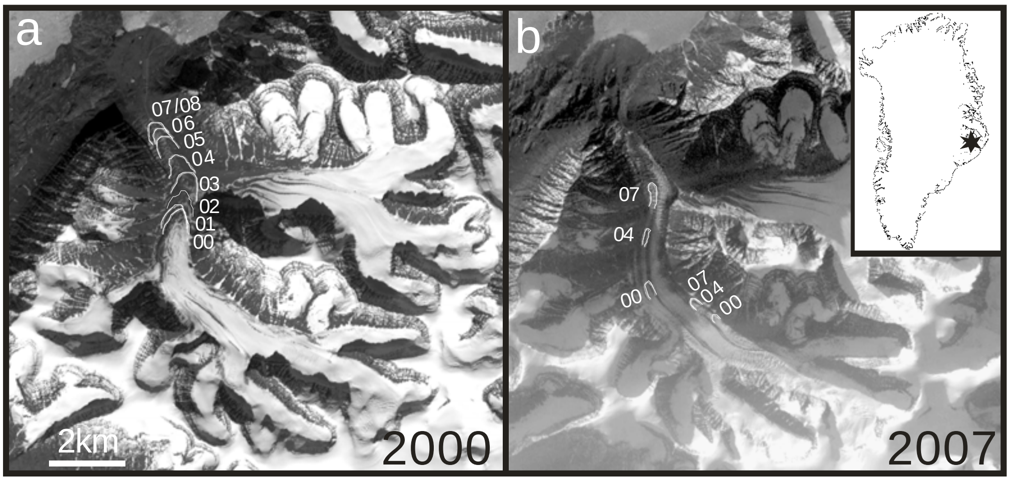

Figure 1. *a) Pre-surge ASTER image of 25/6/2000 with terminus positions between 2000 and 2008. b) Late-surge ASTER image of 19/6/2007 with glacier extending to its full advance of 2.8km. The positions of the 2000, 2004 and 2007 moraine loops are superimposed. Note that the E moraine loop is virtually stagnant after 2004.*

Methodology

We reconstructed frontal positions, elongation of moraine loops and surface characteristics between 1981 and 2008 from 13 ASTER and Landsat 7 images covering the period June 2000-July 2008, and Kort og Matrikelstyrelsen airphotos from 1981 and 1987. Danish Meteorological Institute weather data (http://www.dmi.dk/dmi/index/gronland/vejrarkiv-gl.htm) for were obtained for the nearest weather station, Ittoqqortoormiit (190km east of the glacier), in order to constrain the start of the surge. Change in marginal position was measured in ENVI 4.5 using 3 focal points along the margin. Feature tracking was performed on the furthest downglacier extent of elongated moraine loops and on the propagation of a surge bulge. Advance rates and surge velocities were reconstructed from summer images with approximately one year intervals. The 2000-2008 ice margin and location of two moraine loops are indicated on Fig 1. Quiescent phase velocity is assumed to be equal to creep velocity and is 0.01 md^{-1}, using the method of Murray et al. (2003) and an average ice thickness of 100-125 m. This is given as baseline velocity in Fig 2.

Results, discussion and conclusions

The surge proper started between 2 June and 12 July 2001, and most likely around 20 June (Table 1 and Fig 2). However, the surge initiation probably occurred over months to a year which is reflected by the increasing displacement rate of the surge bulge from 0.07-0.90md^{-1} between August 2000 and July 2001. Maximum flow rates and terminus advance are 5-15md^{-1} and rapidly slow down to <2md^{-1} after a year and then gradually down to <1md^{-1} in 2007 and quiescent phase velocities in 2008. Assuming that the previous surge extended as far as the current it is assumed that the previous surge occurred around 1941. Sermeq Peqippoq therefore has a quiescent phase of ~60 years and an active phase of 6-7 years.

Image Date	Glacier Length km	Displacement terminus m d^{-1}	Displacement W moraine loop m d^{-1}	Error m d^{-1}
4/8/1981	10.5	-0.11	0.01	-
1/1/1987	10.5	0.00	0.01	-
25/6/2000	10.0	-0.11	0.01	-
28/8/2000	10.0	0	0	-
4/5/2001	10.0	0	0	-
1/6/2001*	10.0	0	0	-
		3.43 (start 9/6/01)	6.96 (start 2/6/01)	1.40
		7.56 (start 27/6/01)	12.65 (start 20/6/01)	2.55
12/7/2001**	10.1	56.67 (start 10/7/01)	139.2 (start 10/7/01)	28.0
24/6/2002	10.5	1.01	1.57	0.16
19/6/2003	11.0	1.31	1.29	0.08
18/6/2004	11.7	0.91	0.55	0.15
7/6/2005	12.3	0.66	1.30	0.16
26/7/2006	12.6	0.36	1.03	0.14
19/6/2007	12.8	0.30	0.82	0.17
10/7/2008	12.8	0.01	0.01	0.14

Table 1: *Glacier length, frontal advance and moraine loop displacement rates for time intervals between the images. Three possible start dates are: 2 June = 1 day after the last pre-surge image, 20 June = after a period of melt and rain, 10 July = a day before the first surge image. This last date is unlikely given the large displacement rates. Errors are based on a maximum 2 pixel manual location error and a 2 pixel geocorrection error (4x14.5m) divided by the image interval in days. * last pre-surge image, ** first image in surge.*

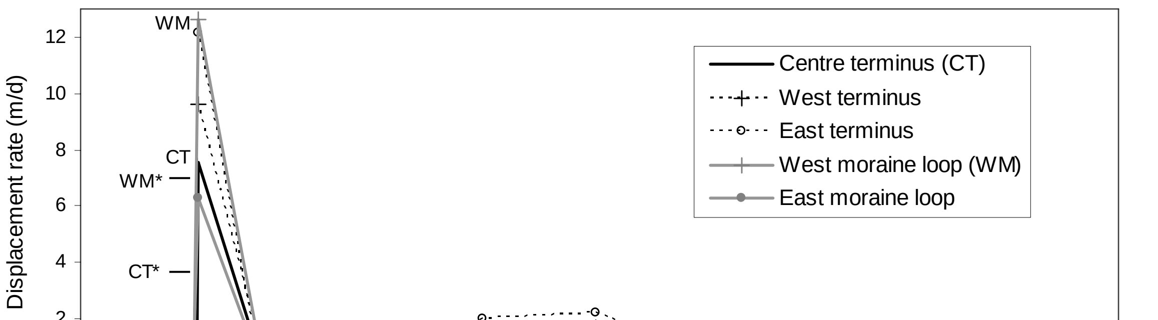

Figure 2. *Displacement rates from frontal advance and feature tracking of two elongated moraine loops (Fig 1). The E moraine loop speed of 0.9md-1 (2000-2001) is based on the surge bulge propagation speed. Displacement rates for the Centre terminus and West moraine loop are indicated for a 20/6/2001 (WM and WT) surge initiation in the line graph and for a possible 2/6/2001 surge initiation as horizontal markers (CT* and WM*). Displacement rates and errors for the Centre terminus and West moraine loop are given in Table 1.*

Morphological changes during surge includethe propagation of a surge bulge, extensive crevassing, frontal steepening and advance, drawdown of the upper glacier observed from trimlines left on the valley walls, elongation of tributary moraine loops and displacement of medial moraines, and the formation of a large supraglacial lake in the first years of the surge.

Regional surge characteristics and dynamics of the 1992-94 surge of Sortebræ have been used to suggest a hydrologically controlled surge mechanism, with surge behaviour that is more Alaskan-type than Svalbard-type in this region (Jiskoot et al. 2001; Murray et al., 2002; Jiskoot et al., 2003; Murray et al., 2003; Pritchard et al., 2005). These surges are characterised by a sudden initiation, sustained high velocities, and a rapid termination that is often accompanied by an outburst flood (Murray et al., 2003). In contrast, the surge behaviour of Sermeq Peqippoq clearly indicates a multi-phase surge, with an acceleration phase over several months to a year, a surge phase with gradual deceleration to quiescence over 6-7 years, maximum surge velocities $<15 md^{-1}$ and prolonged surge velocities of $1\text{-}2 md^{-1}$, a quiescent phase of 60 years, and no sign of rapid discharge events. These are all surge characteristics typical of Svalbard surges (Murray et al., 2003). Surge phases for other East Greenland glaciers are 5-15 years and quiescent phases 70-200 years (Jiskoot et al., 2001), indicating that Sortebræ's surge behaviour was possibly an exception for this region

We postulate that the observation of both surge behaviours in this region might be related to a mixed thermal regime of the glaciers in East Greenland. The occurrence of both surge behaviours in this region coinciding with a transition in thermal regime, the slow development of a surge bulge on Sermeq Peqippoq and the absence of a discharge event, suggest that the two surge mechanisms suggested by Murray et al. (2003) are indeed thermally differentiated, rather than substrate, climate or glacier morphology differentiated.

Note: *Most of the information presented here is in press as a Correspondence paper in the Journal of Glaciology*.

Acknowledgements

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Extreme pollution events as markers in snow cover stratification

Krzysztof Migała[1], Bartłomiej Luks[2], Piotr Głowacki[2] & Dariusz Puczko[2]

[1]Institute of Geography and Regional Development, University of Wrocław
[2]Institute of Geophysics, Polish Academy of Sciences

The chemical properties of precipitation and snow cover have been monitored at the Polish Polar Station In Hornsund (Svalbard) since 1988. Apart from the standard measurements and chemical analyses of daily totals of precipitation, the chemistry of fresh snowfall episodes and the properties of snow cover are monitored in the altitudinal profile of the Hansbreen. Each winter events with extreme load of acidifying pollutants or high content of total dissolved solids occur. Snowfalls with very specific chemical composition might be used as markers in snow cover stratification. Such conclusion is based on results from spring 2006 fieldwork.

The 2005/2006 winter season was the warmest on record for many years: January (-1.7 oC), April (-0.3 oC) and May (-0.2 oC) were extremely warm, and the number of thaw days (117, i.e. 48 % of the October-May period) was exceptionally high. In January, the thaw often reached the highest sections of the glacier. The distinctive feature of this season was the very large amount of precipitation recorded at the Polish Polar Station (330.7 mm) and, even more remarkable, the largest accumulation of snow on the Hans Glacier during the 1988–2006 period (1.30 m w.e. on average). Weather conditions in winter (multiple thaws with precipitation, water infiltration, freeze-thaw cycles) caused intensive snow metamorphosis, an increase in its density and water equivalent as well as the formation of numerous layers of ice and snow crust within the snow cover.

At the end of April 2006, the physical properties of the snowpack were examined at two sites on the Hans Glacier (at 200 and 450 m a.s.l.). Snow samples were collected for hydrochemical analysis (pH, SEC, ion chromatography). In the lower part of the glacier, the snow cover exhibited higher density (0.530 g cm-1), higher stratification and a significant number of ice/snow crust inlays as well as higher acidity (a pH of 3.67). Values of pH and electrolytic conductivity (SEC) indicated significant variation in the hypsometric profile of the glacier. In both snow pits analysis of physical properties of snow and snowpack structure allowed to match several snow layers to exact weather episodes.

Due to often melt-freeze periods and snowdrift such method don't give full information on snow stratigraphy. Solution might be identification of extremely polluted events in snow cover structure, which are often well preserved and easy to read in snowpack structure

On the 15th April 2006 a highly acidified snowfall event occurred, in which a pH 3.57 was recorded at the Polish Polar Station Hornsund. The same day fresh snow samples were collected from the surface of the Hans Glacier in altitudinal profile (Figure 1). This episode took place two weeks before an Arctic haze event observed in Ny Ålesund in the NW part of Svalbard with high concentrations of pollutants (2nd May 2006). The pH of fresh snow ranged from 3.35 to 3.61 with the mean value of pH = 3.48. Acidified precipitation formed a layer, identified in the both snow pits structure 10 days later, i.e. 24-26 April (Figure 2). Event identification was supported by ionic composition of fresh snowfall and snow from identified layer.

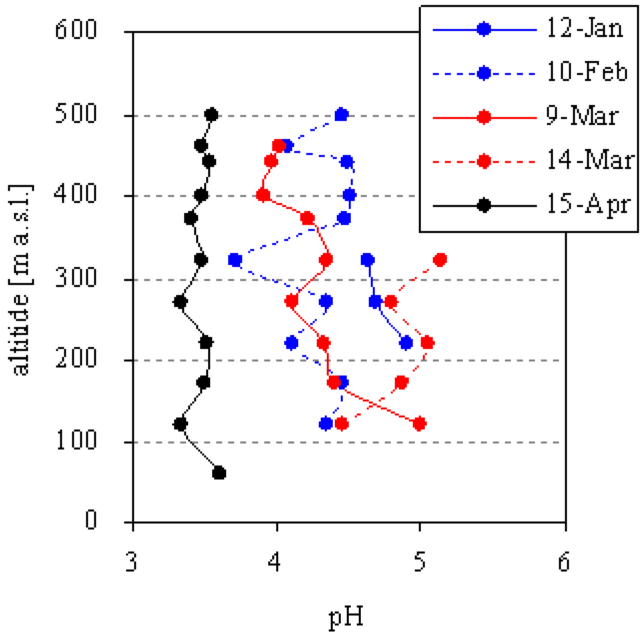

Figure 1. *Fresh snowfall acidity in the altitudinal profile of the Hans Glacier, 2006.*

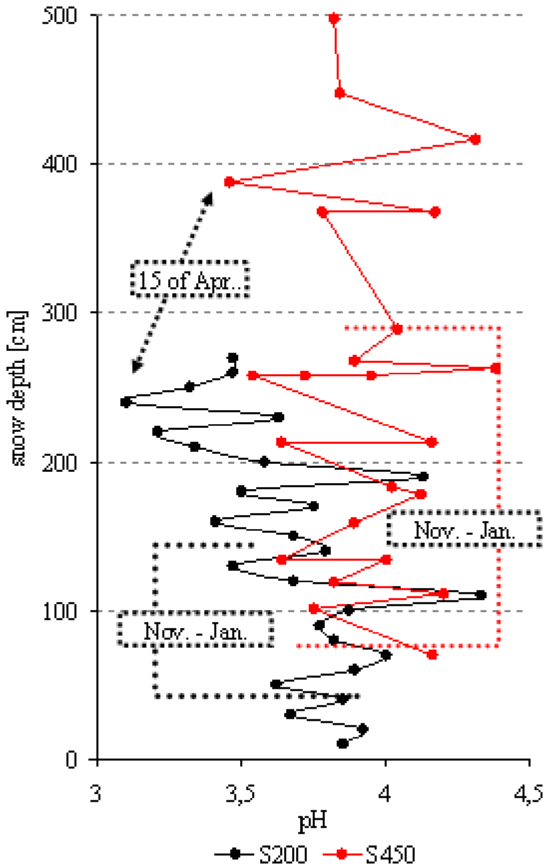

Figure 2. *Snowpack acidity at the 200m and 450m a.s.l., Hans Glacier, 26 of April 2006.*

On the hydrothermal state of Ariebreen, Spitsbergen

F.J. Navarro[1], P. Glowacki[2], J.J. Lapazaran[1], M. Petlicki[2] & F. Machío[3]

[1]Dept. Matemática Aplicada, ETSI de Telecomunicación, Universidad Politécnica de Madrid, Av. Complutense, 30, 28040 Madrid, Spain, fnv@mat.upm.es, jlapazaran@mat.upm.es
[2]Institute of Geophysics, Polish Academy of Sciences,
ul. Księcia Janusza 64, Warsaw, Poland, glowacki@igf.edu.pl, petlicki@igf.edu.pl
[3]Dept. Electrónica y Comunicaciones, Facultad de Ingeniería Informática, Universidad Pontificia de Salamanca, Pº Juan XXIII, 3, 28040 Madrid, Spain,
francisco.machio@upsam.net

Introduction

Ariebreen (77º 01' N, 15º 29' E) is a small valley glacier (ca. 0.36 km^2 in August 2007) terminating on land, located at Hornsund, Spitsbergen, Svalbard, ca. 2.5 km to the west of Hornsund Polish Polar Station. Like many other Svalbard glaciers, it has experienced a significant recession at least since the 1930s, and most likely since the end of the Little Ice Age in the early 20th century, which has accelerated from the 1990s. In the case of Ariebreen, this recession has been manifested as retreating of the ice front and thinning of most of the glacier, except for a small zone in the uppermost area, where the ice thickness has remained rather stable.

Ice-penetrating radar studies performed in summer 2006 and spring 2007 (Navarro et al., 2008) have suggested that Ariebreen is a cold glacier, in contrast with the polythermal structure of the neighbouring Hansbreen and Werenskioldbreen. This is mainly based on the following evidences: 1) nearly absence of endoglacial diffractions for both high (200 MHz) and low (25 MHz) frequency radar data; and 2) absence of any internal reflector which could be interpreted as an interface between cold and temperate ice layers, again true for both high and low frequencies. These characteristics are illustrated by the sample radar profile shown in figure 1.

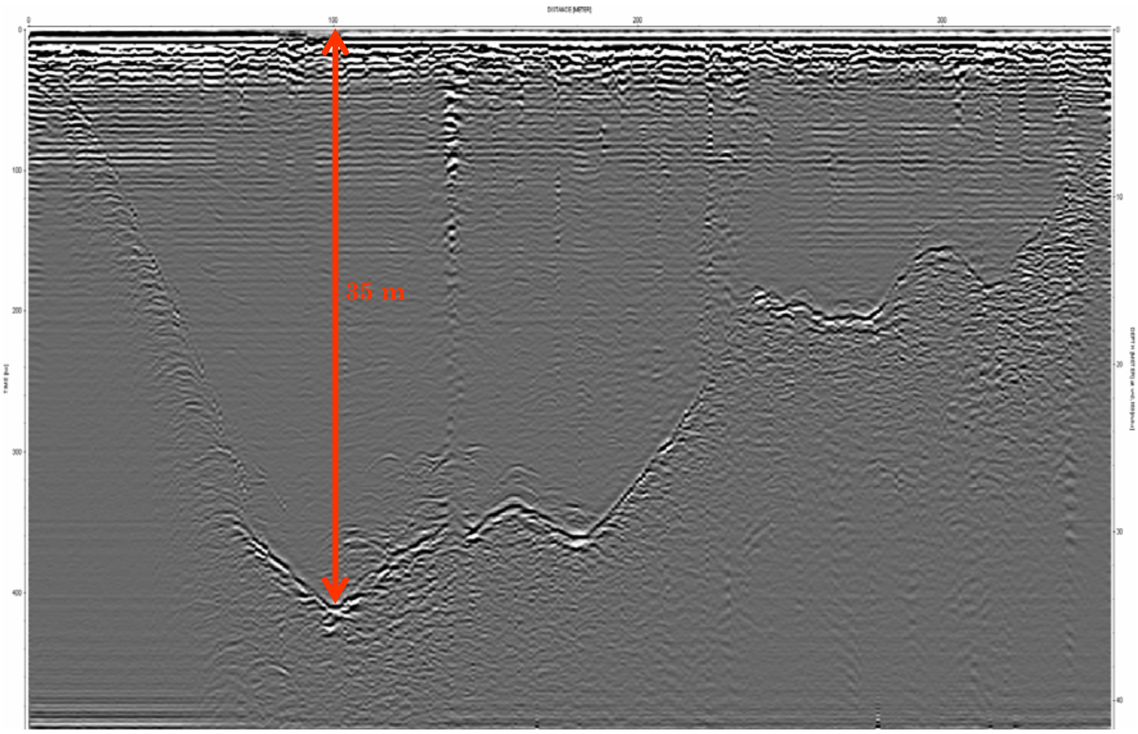

Figure 1. *Sample transverse radar profile in the ablation area of Ariebreen (200 MHz radar).*

In this contribution, we explore the reasons that could explain the present cold hydrothermal state of Ariebreen, and a hypothetical transition from polythermal to cold structure even under a climate warming environment such as that observed in this region during the last century and, in particular, the most recent decades.

Polythermal to cold evolution of Ariebreen during the recent past

We hypothesize an evolution of Ariebreen from polythermal to cold structure in the recent past. Some models have been presented for explaining the migration of the cold-temperate transition surface (CTS), e.g.

- Macheret et al. (1992) estimated the time needed, under different scenarios of water content in temperate ice and surface temperature cooling, for a thickening of the cold layer of -Fridtjovbreen, Svalbard, leading to the total disappearance of the temperate ice layer, thus becoming a fully cold glacier.
- Pettersson et al. (2007) have attributed the thinning of the cold ice layer of Storglaciären, Sweden, during the most recent decades, to an increase in winter air temperature since the mid-1980s.

But, why should a polythermal glacier evolve to a cold one under a climate warming scenario such as that experienced in Svalbard since the end of the LIA?

First, let us present some physics of the migration of the CTS.

-The thickness of the cold layer is a function of

  - the surface energy flux,
  - the net ablation at the surface,

- the water content in the temperate ice next to the cold layer,
  - the velocity field.
- The CTS can migrate according to the balance between
  - the amount of latent heat released when the top of the temperate layer freezes, which depends on the liquid water content,
  - the capability to transport this heat away from the CTS, which depends on the temperature gradient through the cold ice layer.
- The thickness of the cold ice layer depends on the balance between
  - the migration of the CTS,
  - the ablation of ice at the glacier surface.

But, where does the heat causing the CTS migration come from?
- Fowler (1984) rightly pointed out that only internal heat generation can lead to an internal "mushy" zone where two phases of ice and water coexist.
- A heat source at a boundary (upper surface, bed) merely leads to a classical Stefan problem at a melting surface.
- The typical heat sources at the glacier boundaries, which are
  - the radiative/turbulent energy fluxes at the glacier surface,
  - the geothermal heat flux and frictional heat at the base,

do not contribute per se to the development of temperate ice.
- The relevant heat sources for the latter are
  - the strain heating, which is most effective in the basal ice and favoured by a thicker ice,
  - the percolation of meltwater in firn,
  - to a minor degree, the radiation penetration in a rather thin surface layer.

Figure 2 shows a sketch of the above mentioned heat sources.

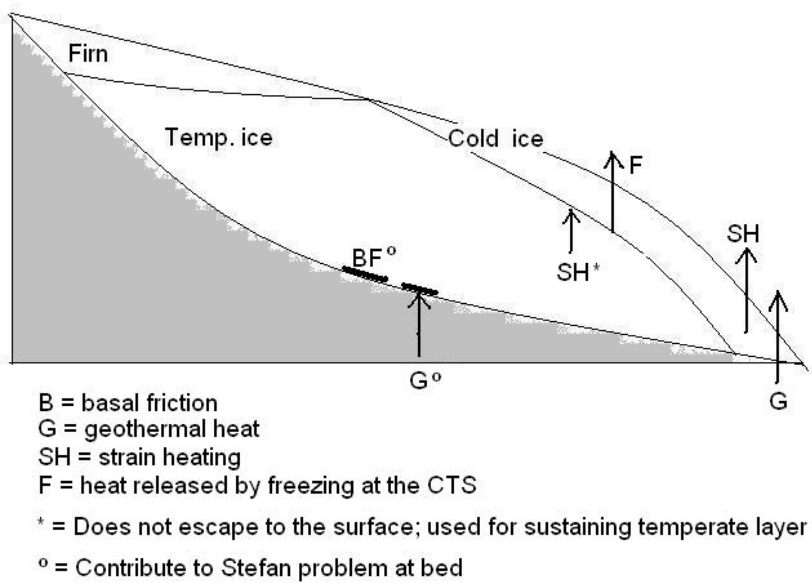

Figure 2. *Sketch of the heat sources relevant and not relevant to the CTS migration.*

Let us now present our hypothesis concerning the polythermal to cold evolution of Ariebreen during the recent past.

- We are hypothesising a downward migration of the freezing front under the warming conditions experienced in Svalbard during the last century.
- However, even under such warming scenario, the strong thinning experienced by the ablation area of Ariebreen during the last century, mostly attributed to increased melting at the surface, could dominate the opposite effect of climate warming.
- As discussed by Blatter and Hutter (1991), the thinning can contribute to the downward migration of the freezing front through two mechanisms:

    First mechanism

    - Along the freezing part of the CTS, heat is released and removed from the CTS, which is possible due to the generally negative temperature gradient of the cold ice layer. A thin cold layer makes easier the transmission to the surface of the latent heat released at the CTS when the moisture at the temperate ice at the CTS freezes.
    - Our hypothesis for Ariebreen: During the last century, there was not an increase in the thickness of the cold layer but a downward migration of the CTS simultaneous with an increased melting at the surface, so as to approximately maintaining the thickness of the cold layer, while gradually decreasing the thickness of the temperate layer, until its disappearance. In other words, the thinning of Ariebreen would have been mainly accomplished through the thinning –and final vanishing– of its temperate layer.

    Second mechanism

    - The decrease in strain heating in the basal ice resulting from the thinning of the glacier. This influence should not be underestimated, as shown by the earlier discussion on the comments by Fowler (1984).
    - Our hypothesis for Ariebreen: The thinning of Ariebreen lead to a decrease of strain heating and, consequently, a reduction of the heat necessary for maintaining its internal temperate ice layer.

Outlook

We plan the following fieldwork in order to confirm the cold structure of Ariebreen:

- Radio Wave Velocity (RWV) measurement by the Common MidPoint (CMP) method. Planned April 2009.
- Temperature measurements (thermistor chain) at a “deep” borehole.

Concerning the modelling of the polythermal to cold transition, we plan to develop a 1D thermomechanical model similar to that in Pettersson et al. (2007), i.e. a 1D time-dependent heat diffusion and advection model, for which the boundary conditions for the cold ice layer would be the surface temperature and melting temperature at the CTS. This would be complemented by a 1D transition condition at the CTS. The changes in the vertical position of the CTS would be found from the latter, provided that the temperature gradient at the CTS, the water content and the vertical velocity are prescribed. The temperature gradient at the CTS would be found by integrating equation heat diffusion-advection equation, but the water content, vertical velocity and surface temperature should be assumed or taken from measurements. This model would be used to check whether a polythermal to cold transition is likely taking into account measured (e.g. past local temperatures), or computed (e.g. velocities from modelling for past glacier geometry), or assumed (e.g. by similarity with nearby glaciers) parameter values.

Concluding summary

- Radar records suggest a cold thermal structure for Ariebreen, different from the usual polythermal structure of neighbouring glaciers.
- Numerical modelling experiments can support whether or not Ariebreen has experienced a transition from polythermal to cold structure during the recent decades.
- Common midpoint method measurements are an easy tool to determine whether the radio-wave velocity in Ariebreen is typical of purely cold ice.
- Deep borehole temperature measurements (thermistor chain) would of course be key to confirm the thermal structure.

Acknowledgements

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Coupling Glacial Hydrology into a High-Order Numerical Ice Model

Sam Pimentel Gwenn Flowers
Department of Earth Sciences, Simon Fraser University

Introduction

A new high-order ow-band model with coupled subglacial hydrology is used to explore the drainage of supraglacially-stored water through englacial fractures and assess the inuence of this water on glacier dynamics. This work forms part of the Canadian contribution to IPY project GLACIODYN. The focus of this project is a study of the dynamics of the Belcher glacier located in the Canadian high arctic. The Belcher Glacier system is the major outlet of the Devon Island Ice Cap. This is a large, fast-flowing, tidewater glacier sharing similarities with many Greenland outlet glaciers.

Model Description

Our model is two dimensional being composed of one horizontal dimension, in the direction of the ice flow, and one vertical dimension through the ice column. We also include a flow-band adaption by having a parameterization of the width across the flowline.
The model incorporates a multilayer longitudinal stress scheme following [2] and [5]. The longitudinal stresses are expected to have an essential role at the Belcher especially near the terminus and where basal sliding is considerable.
As a means of validating our model we have conducted experiments that form part of the Ice-Sheet Model Intercomparison Project - High-Order Models (ISMIP-HOM) [6]. Results from one particular example, Experiment E: Haut Glacier d'Arolla, are presented here. This experiment provides a test for the velocity/stress solution of the non-linear force-balance equations. It uses the fixed geometry of Haut Glacier d'Arolla under isothermal conditions and simple basal boundary conditions. Results for surface horizontal velocity and basal shear stress are shown to be within the benchmark range.
The model has an evolving free surface by considering conservation of mass and can be thermomechanically coupled by solving an advective-diffusive heat equation. The coupling occurs through the temperature dependent flow-law parameter, as well as from internal friction from deformational heating. There is also the framework here for potential coupling to the hydrological system.
As this model is intended for a tidewater glacier a suite of basic options for calving has been installed. These include the water-depth relation [4], the flotation criterion [8], and a more recent rule based on crevasse formation [1].
The hydrology aspect of the model incorporates vertical fracture propagation and a subglacial drainage system. The drainage system will ultimately comprise ‘slow’/distributed and ‘fast’/channelized drainage networks. The hydrology model is taken from the subglacial water drainage component of [3]. This considers conservation of mass in a subglacial water sheet

$$\frac{\partial h^s}{\partial t} + \frac{\partial Q}{\partial x} = \frac{Q_G + u_b\tau_b}{\rho L} + M_b, \qquad Q = -\frac{Kh^s}{\rho_w g}\frac{\partial \phi}{\partial x},$$

where h_s denotes the water sheet thickness, Q is a Darcian water flux $\phi = P_w + \rho_w gb$ reprsents the fluid potential, and K = K(hs) the hydraulic conductivity. From this we determine the water pressure $P_w = P_i\ (h^s/h^s{}_c)^{7/2}$, where $h^s{}_c$ is some critical water sheet thickness. In order to mimic vertical englacial fracture propagation we employ the linear elastic fracture mechanics of [9].

The hydrology interacts with the ice dynamics through the basal interface. To model the basal sliding we use a Coulomb friction law as proposed and advocated by [7]. This law relates the basal drag, τ b, to the basal velocity, μb, as follows

$$\tau_b = C\left(\frac{u_b}{u_b + N^n\Lambda}\right)^{1/n} N, \qquad \Lambda = \frac{\lambda_{max} A}{m_{max}}$$

where N = P_i - P_w is the effective pressure, λ_{max} a dimensional wavelength for the dominant bedrock bumps, m_{max} a typical bed slope, A and n are Glen's flow law parameters and C is a constant. This is a non-linear Robin-type boundary condition which cannot be solved independently but forms part of the solution to the ice-flow problem. This type of sliding law is favoured over the more standard power law formations which allow large basal stresses to develop at the bed for an arbitrary effective pressure.

Using a two-dimensional model we assume that ice flows in an infinite plane; however, glacier flow is confined in a channel and thus is effected by lateral drag. We parameterize the lateral drag at the side walls and include this stress term in our force-balance equations.

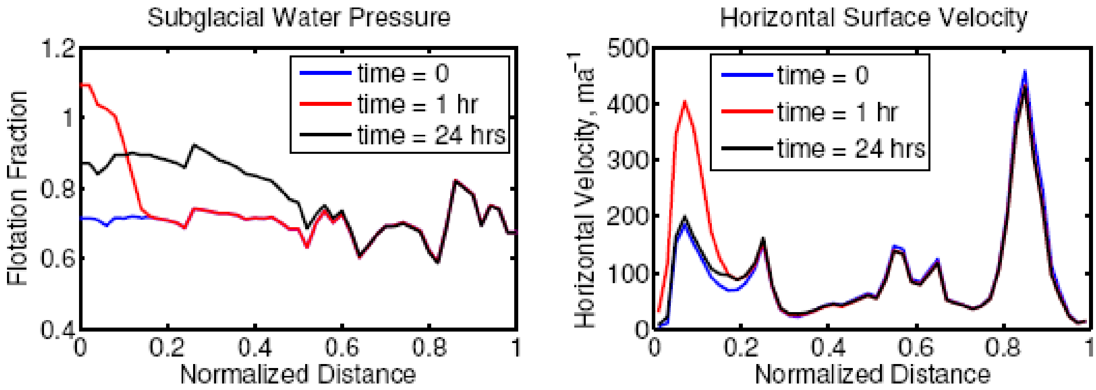
$$\sigma_{xy} \approx -\frac{\nu(u - u_L)}{W} = C\left(\frac{u_L m_{max}}{u_L m_{max} + N_L^n \lambda_{max} A)}\right)^{1/n} N_L,$$

Where u_L is the lateral sliding along the side walls. This sliding is computed by again applying the Coulomb friction law with the vertical distribution of effective pressure along the side walls, N_L.

A Drainage Scenario

An experiment to mimic the drainage of a supraglacial lake in an effort to understand the coupling between hydrology and ice dynamics is conducted. In this scenario we envisage supraglacial drainage into a crevasse forcing englacial fracture propagation, with high enough tensile stresses and water injection rates the fracture reaches the bed providing a surface to bed connection. Rapid drainage of a meltwater pond through this fracture occurs providing local water injection into the subglacial drainage system.

We then monitor the transient response of water pressure and glacier flow speed, as seen in this figure.



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Spatial Variability of Snow Accumulation Patterns across the Belcher Glacier, Devon Island, Nunavut, Canada

Tyler Sylvestre[1], Luke Copland[1], Michael Demuth[2], John Sekerka[2]

[1]Department of Geography, University of Ottawa, Ottawa, Ontario K1N 6N5, Canada
[2]Natural Resources Canada, National Glaciology Group, 562 Booth Street, Ottawa, Ontario K1A 0E4, Canada

The tidewater Belcher Glacier (75°N, 82°"W), in the northeast sector of Devon Ice Cap, covers 1180 km2 and is the largest and fastest flowing (up to 300 m yr-1) outlet of the ice cap. The basin faces the North Open Water (NOW) polynya at the head of Baffin Bay and its flow is largely dominated by basal sliding and bed deformation as indicated by flow stripes on the glacier surface (Burgess and Sharp, 2004). Over the past 40 years, the glacier has experienced thinning along its entire extent and the calving front has retreated ~1.3 km (Burgess and Sharp, 2004). Belcher Glacier accounts for approximately half of the iceberg calving loss and ~15% of the total mass loss of the entire Devon Ice Cap (Burgess and Sharp, 2004).

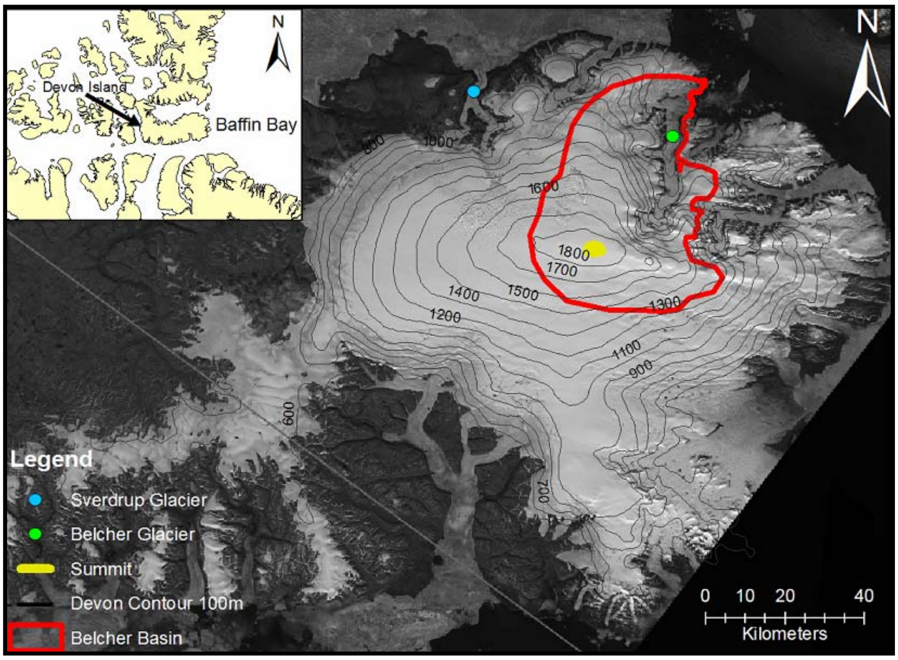

Figure 1. *Devon Ice Cap location Map (left) with inset map of QEI (Base Image Landsat 7 ETM+ July 7, 2000).*

A series of field measurements have been conducted on Devon Ice Cap since the 1960s to assess surface mass balance changes, including shallow and deep ice cores (Koerner, 1977), snow pits (Bell et al. 2008), gamma-spectrometry measurements (Colgan and Sharp, 2008; Mair et al. 2005), automatic weather stations, mass balance stakes, and remote sensing studies (Burgess and Sharp, 2004). Few of these measurements have been focused in the NE sector, however, but rather concentrated in the NW (Sverdrup Basin), summit (1920 m a.s.l) and SW regions. Mass balance patterns and changes in the Belcher Glacier basin have been largely interpolated from these remote point measurements, resulting in uncertainties concerning the ability of these measurements to truly represent the conditions within the basin. This study presents the first high frequency (500 MHz) GPR data collected in the Belcher basin over ~306 km of track in May 2008. The main goal of this research is to quantify annual and long-term basin wide spatial variability in snow accumulation across the glacier. GPS data was recorded simultaneously with GPR surveys for position, elevation, and length of survey line.

Figure 2. *GPR profiles for 2008 Field season (Base Image Landsat 7 ETM+ July 7, 2000), PulseEKKO Pro 500 MHz Transmitter and Receiver housed in a sled and pulled by skidoo.*

The main moisture source region for Belcher Glacier and Devon Ice Cap is Baffin Bay and the NOW polynya. Maps of winter mass balance suggest there is a SE - NW accumulation gradient, with higher accumulation rates on eastward flowing basins (up to 500 kg m-2 a-1), compared with the west (200 kg m-2 a-1) (Burgess and Sharp, 2004). Mair and others (2005) point out that locally, accumulation reduces dramatically from the summit (~ 1920 m a.s.l.) to 1600 m a.s.l. in the NW and gradually decreases towards 1300 m a.s.l. at the equilibrium line in this region.

Near surface snow density (p) for the Belcher was measured by digging snow pits (< 1 m) to last summers surface (LSS) (2007) and recording density on the pit wall at three equi-spaced intervals using a snow scoop (250 g cm -3) and digital scale. Density measurement intervals varied depending on the snow depth to the LSS e.g. snow depth of 45 cm, density intervals at 15 cm, 30 cm, and 45 cm. The LSS is a distinct hard / dense layer which forms at the end of the summer melt season and is identified because it contrasts from the lighter/less dense annual snow removed above this layer. The LSS is also dirty, due to the incorporation of dust /dirt redistribution by wind during the summer which subsequently freezes at the end of the melt season. This layer forms a continuous lense across the basin. The depth of snow above the LSS represents the annual accumulation for the pit site. The three p-values of the annual snow pack are averaged for the pit to give a near surface density profile of the 2007/08 annual accumulation, and averaged < 400 kg m3 for all pits combined. Measurements were made in May, prior to the onset of melt, which enabled identification of shallow layers and annual isochrones using high frequency GPR over long (5 – 10 km) distances. GPR provides information with regards to the spatial continuity of point measurements and also enhances the understanding of controls on snow accumulation patterns (e.g. topography, elevation, and distance from moisture sources).

Pulse Ekko Pro software and filters were used to post-process the GPR data. Preliminary results from 500 MHz surveys (Figure 3) across a 9 km transect shows that snow layers are detectable to depths > 10 m. Layering in GPR traces is related to dielectric contrasts in the snowpack which are caused by density variations between snow (0.3 – 0.4 kg m3), firn (0.5 – 0.7 kg m3), and ice (0.8 – 0.9 kg m3). Annual layers e.g. LSS, can be tracked and identified in GPR traces across long distances (>10 km) because of density contrasts between snow, ice and firn, and also because the LSS is dirty, all which produce high GPR reflections. Additionally, the depth to the LSS expressed in GPR traces is ensured via in situ field measurements from snow-pits and avalanche probe so that the correct tracked layer is associated with the LSS.

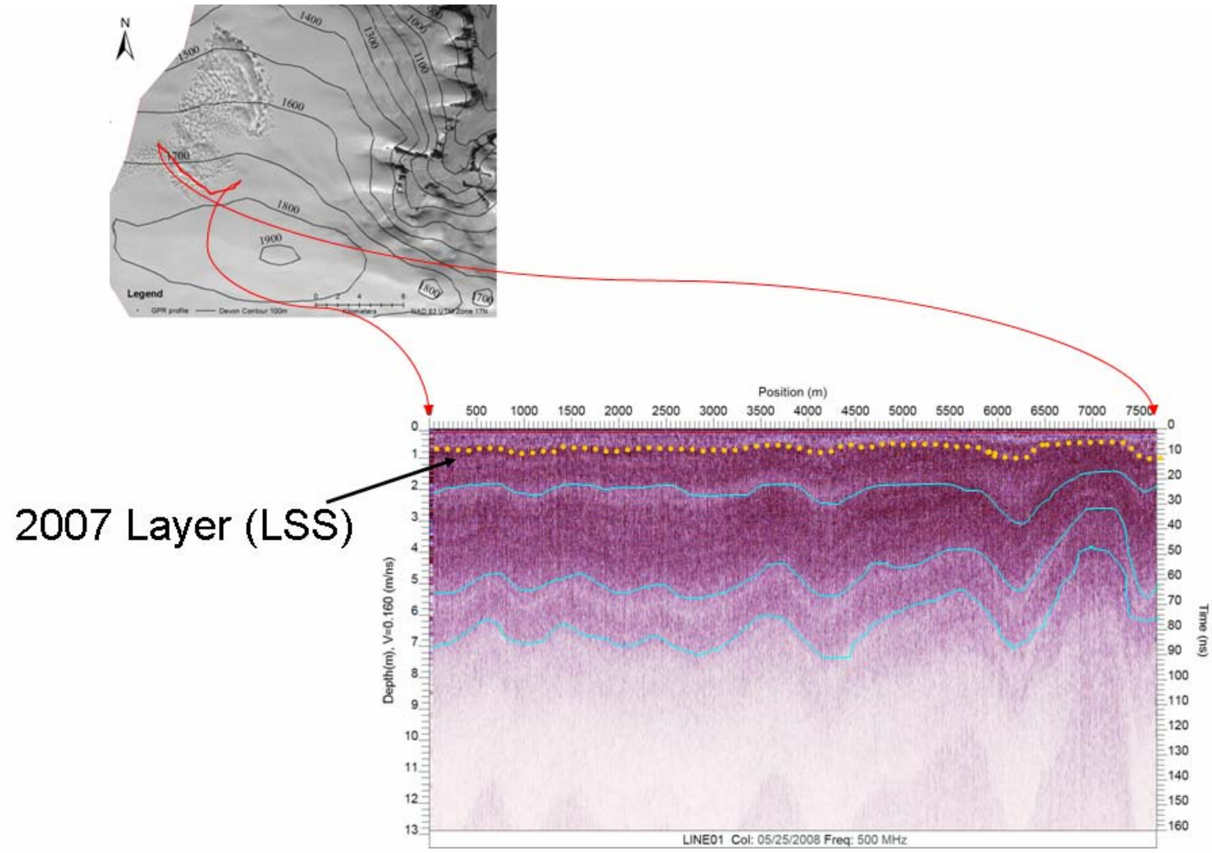

Figure 3. *500 MHz GPR data recorded on May 25, 2008 (~ 9 km by 13 m deep) taken in the accumulation zone (1696 – 1795 m a.s.l); inset map showing location of GPR profile (red line) (Base Image: Landsat 7, June 7, 2000).*

Belcher Glacier shows large spatial variation of snow accumulation on the surface. Locally snow varies on horizontal length scales of 1 – 10 m due to wind redistribution forming sastrugi, and on the order of 100 m due to topographic inundations such as gullies where snow can collect. Broad-scale (basin wide) patterns in snow accumulation via GPR data suggests that the spatial variability in snow thickness is largely related to changing altitude.

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Continuous flow observations on Svalbard glaciers

Carleen Reijmer[1], Marianne den Ouden[1], Veijo Pohjola[2], Jack Kohler[3], Monica Sund[4], Jon Ove Hagen[5], Thorben Dunse[5] and Johannes Oerlemans[1]

[1] Institute for Marine and Atmospheric Research, Utrecht University, The Netherlands
[2] Department of Earth Sciences, Uppsala University, Sweden.
[3] Norwegian Polar Institute, Tromsø, Norway
[4] University Centre in Svalbard, Longyearbyen, Norway
[5] Department of Geosciences, Faculty of Mathematics and Natural Sciences, University of Oslo, Norway

In recent years several papers have been published discussing the relation between melt water and glacier dynamics (e.g. Zwally et al., 2002). The simplest explanation of this relation is that increasing amounts of melt water lead to an increase in glacier velocity through improved lubrication at the bed. However, more recent publications have shown that although there seems to be a direct relation between glacier velocity and melt water at the start of the melt season, on longer time scales this relation is less clear (Van de Wal et al., 2008). In the framework of the IPY-GLACIODYN project *'Meltwater input, flow and calving of Arctic glaciers'* several glaciers on Svalbard have been selected to study this relation between ice velocity and melt water production (Figure 1a, Table 1).

The ice velocity measurements are performed using relatively low cost stand alone single frequency GPSes (Figure 1b). The GPSes store hourly data locally except for the GPSes on Kronebreen, which are equipped with an Argos transmitter broadcasting new data every six days. From the hourly stored positions flow velocities are calculated. The mass balance is determined by traditional stake measurements and by means of a sonic height ranger.

First results on velocities and mass balance are reported in Reijmer et al. (2008). Here we briefly summarize the main results for two years of velocity measurements on Nordenskiöldbreen, and the first results of velocity observations on Kronebreen. First data for the other glaciers will become available in May 2009.

a) 

b) 

Figure 1. *a) The Svalbard archipelago with the locations of the target glaciers studied with our GPS system marked in red circles. Numbers refer to the glaciers in table 1. b) GPS mounted on a mass balance stake on Nordenskiöldbreen.*

Table 1. *Information on location of the GPS systems on Svalbard (see also Figure 1a).*

	Glacier	No. GPSs	Start	Remarks
1	Nordenskiöldbreen	10	March 2006	
2	Kronebreen	5	September 2008	Argos transmitting
3	Vestfonna	2	April 2008	
4	Duvebreen, Austfonna	5	April 2008	
5	Basin no. 3, Austfonna	5	April 2008	

Nordenskiöldbreen

For Nordenskiöldbreen 2 years of data are presently available (April 2006 - April 2008). Of the 10 GPSes, one is placed on a nunatak serving as a reference site that can be used to asses the uncertainty in the observations. The reference site shows a maximum displacement from the mean location over the two-year period of 5.8 m. The standard deviation in the mean location, a measure for the annual accuracy, is 1.7 m. DGPS observations of the reference site correspond to within one standard deviation. Frequency analysis was performed to determine the presence of system related periodicities. This analysis only shows significant peaks for periods related to the diurnal cycle in solar radiation.

On the glacier, observed annual mean velocities range from 45 to 55 m/yr on the main flow line. These velocities correspond reasonably well with DGPS determined velocities. The observations furthermore show the expected velocity peak (values ranging from 80 - 105 m/yr) in early summer related to the onset of melt. Probably due to the short distances of several km between the different sites there is no significant difference in timing of the peak between the different sites. An interesting feature in the observations is the velocity fluctuations in the winter on time scales of about 20 days. These fluctuations have typical amplitudes of about 40% of the annual mean velocities, are do not show up in the frequency analysis of the reference site, and are not significantly related to temperature or mass balance.

Kronebreen

For Kronebreen data since September 2008 are available. The observations show the expected increase in velocity towards the front of the glacier from ± 100 m/yr to 460 m/yr. More interesting is the increase in velocity with 30 to 60 m/yr for the lower locations towards the winter season (September to January). Another interesting feature is the observed variations for all sites with a typical amplitude of 40% of the period mean velocity, on time scales of about 20 days, as observed on Nordenskiöldbreen.

Conclusions and outlook

Velocity data of several glaciers on Svalbard are at present becoming available. First results show the expected features such as the velocity peak in the early melt season. Due to the reasonably high temporal resolution of the date, interesting variations that seem to be unrelated to mass balance or temperature are visible. These and other features will be analyzed in more detail in the near future.

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New low-frequency radio-echo soundings of Austfonna Ice Cap, Svalbard

Evgeny Vasilenko[1], Francisco Navarro[2], Thorben Dunse[3], Trond Eiken[3], Jon Ove Hagen[3] and Geir Moholdt[3]

[1]Institute of Industrial Research Akadempribor, Academy of Sciences of Uzbekistan, Tashkent, e-mail evgvasil@mail.eanetways.com
[2]Dept. Matemática Aplicada, ETSI de Telecomunicación, Universidad Politécnica de Madrid, Spain, e-mail fnv@mat.upm.es
[3]Department of Geosciences, University of Oslo, Norway, e-mail joh@geo.uio.no

Introduction

Austfonna (79.5°N, 25°E), ca. 8120 km^2, is by far the largest ice cap in Svalbard and one of the largest in the Arctic. Within the framework of the IPY project GLACIODYN and CryoSat calibration and validation activities, annual field visits to Austfonna started in April 2004.

Ice thickness and bedrock maps inferred by airborne radio-echo sounding (RES) are available for the entire ice cap, but date back in time to the mid 1980s. Here we present additional new bedrock information acquired using a ground-based 20 MHz system.

Radar equipment and profiles

The low-frequency radar data were acquired using a ground-penetrating radar VIRL-6, with centre frequency of 20 MHz. The transmitter generates pulses of 25 ns duration at a pulse repetition frequency of 20 kHz, and the receiver uses a variable-gain linear amplifier. The antennae were resistively loaded half-wave dipoles of 5.8 m in length, which were arranged collinearly, following the profiling direction. The digital recording system (DRS) counts on a high stacking capability (user-selectable, from 256 to 16384 traces summed in real time) aimed at suppressing external electro-magnetic noise. The DRS records 4000 samples for each trace, with a 2.5 ns sampling interval. Navigation information from a GPS Garmin II Plus receiver was recorded simultaneously with radar data. For profiling, the transmitter and the receiver+DRS were placed on separate pulka sledges towed by a snowmobile and the antennae deployed along the rope joining the convoy elements. Figure 1 shows the detail for the receiver+DRS pulka.

Figure 1. *Pulka with receiver, DRS, GPS and odometer*

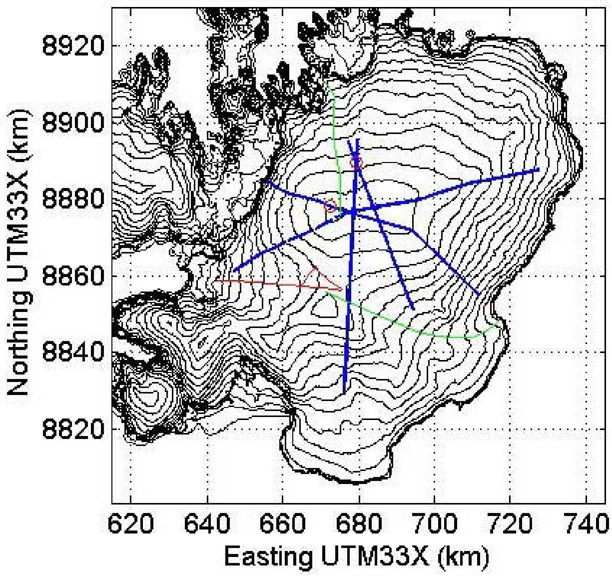

Figure 2. *Layout of the radar profiles.*

Data was collected along several profiles across the ice cap and along/across the flow lines of two fast flowing outlets (see Figure 2), with the aim of complementing the existing airborne data and to provide an improved input geometry for numerical studies of the flow dynamics of Austfonna. Furthermore, data was collected along some parts of the flight lines from 1983/1986 (Dowdeswell, 1986; Dowdeswell et al., 1986) and 2007 (Kristensen et al., 2008) in order to validate the quality of the airborne data.

Total length of VIRL-6 profiles made on Austfonna was close to 800 km. The bed return could be seen in all profiles and also internal reflection horizons could be followed for long distances. A sample radargram is shown in Figure 3. For time to depth conversion, a radio-wave velocity of 168 m/μs, typical of cold ice, was used. Maximum ice thickness of 590 m was found at the central part of the ice cap.

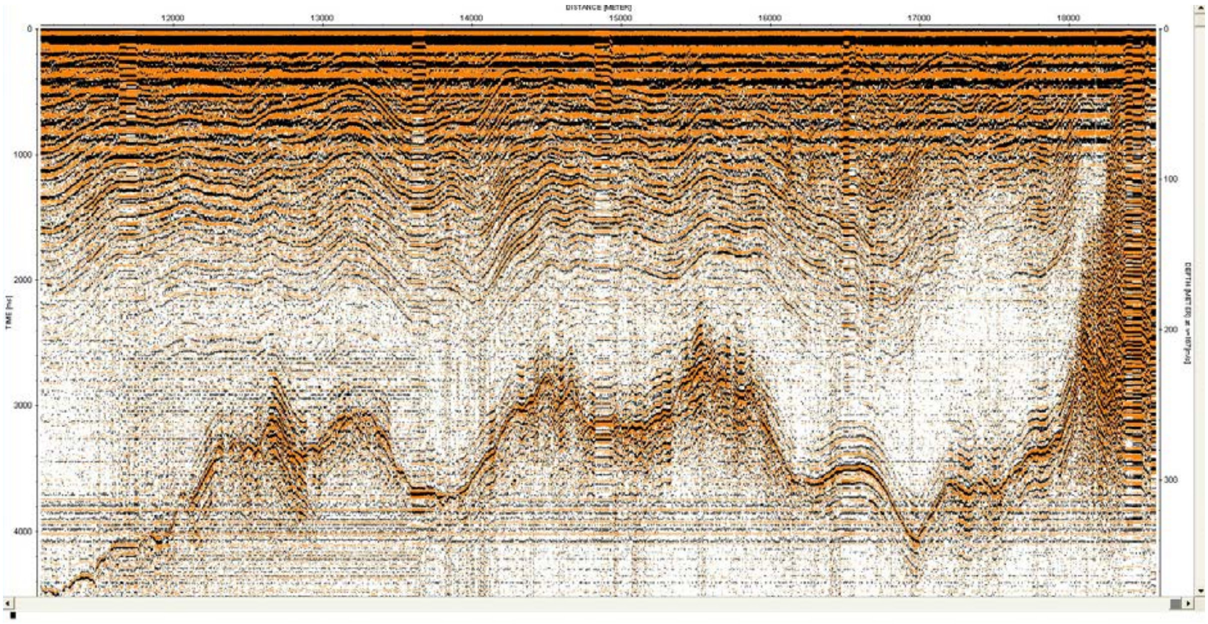

Figure 3. *Sample radargram, corresponding to the 18.5 lowermost km of the south-eastern profile in Figure 1.*

Some preliminary results from radar data processing

1) Internal reflection horizons are visible down to a depth of about 200 m and are related to both accumulation (surface mass balance) and topography.

2) An important aim of the radar work was to map the thickness of the thin temperate layer (say 20 m) that was assumed to exist next to the bed, at least at certain areas of the ice cap, as inferred from the 566 m long ice core retrieved from the summit in 1987 (Kotlyakov et al., 2004). However, our radar data (at least in its present stage of processing) does not allow to identify such layer. Below a depth of 200 m, the ice is nearly absent of internal reflections or diffractions, and the only reflection below 200 m is that from the bedrock. The temperate layer could indeed exist, but be too thin to be properly resolved from the bedrock reflection, or perhaps the reflection from it be so faint (due to the low amount of radar energy reaching such depth) that it could not be detected by our radar. An exception concerning the absence of temperate ice is Duvebreen (northernmost profile in Figure 2), where temperate ice is visible in the radargrams. Here, the ice is channeled into a narrow fjord/valley and strain heating might be significant. Therefore, further data analysis, combined with modelling work, is required in order to get a better understanding of the thermal regime of Austfonna ice cap.

3) The radar data is also useful in order to map both individual crevasses and crevasse fields.

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Inter-annual variability in snow/ice facies distribution on Arctic land ice

Gabriel Wolken and Martin Sharp

Department of Earth and Atmospheric Sciences
University of Alberta, Canada

Temperature increases in polar regions have exceeded those elsewhere on the planet, and climate model predictions consistently indicate that this trend will continue. One consequence of continued temperature increase in these regions is the widespread reduction of glaciers, ice caps, and ice sheets and subsequent rise in global sea level. Although the potential contribution to sea level rise from ice sheets is greater than that from ice caps and glaciers, the latter may be more sensitive and respond more rapidly to changes in climate (Raper and Braithwaite 2006; Meier et al 2007). Our knowledge of the mass loss of Arctic ice caps and glaciers is limited, however, due to sparse *in situ* surface mass balance measurements, which are typically of short duration and usually biased towards smaller ice masses. Snow and ice facies distributions are sensitive to changes in the annual processes of accumulation and surface melt and can be used as proxy indicators of inter-annual changes in surface mass balance. Active microwave remote sensing data can provide valuable information about the distribution of snow and ice facies over large spatial areas, and can augment our knowledge of the spatial variations in the regional surface mass balance of Arctic land ice.

Time series of enhanced resolution QuikSCAT (QS) scatterometer data were used to compute the post-freeze-up, bi-weekly averaged σ^0 for each pixel on ice caps in the Queen Elizabeth Islands (QEI), Arctic Canada, for each year (1999-2005), and the average post-freeze-up bi-weekly averaged σ^0 for the seven-year period was used to characterize the different snow and ice facies on each ice cap. ISODATA classification of the mean post-freeze-up bi-weekly average σ^0 signal for the seven-year period resulted in the delineation of four snow and ice facies (interpreted as the percolation, saturation, superimposed ice and glacier ice zones). The same classification procedure was performed for each year in the time series and changes in the spatial distribution of snow and ice facies during the 1999-2005 period can be seen in Figure 1.

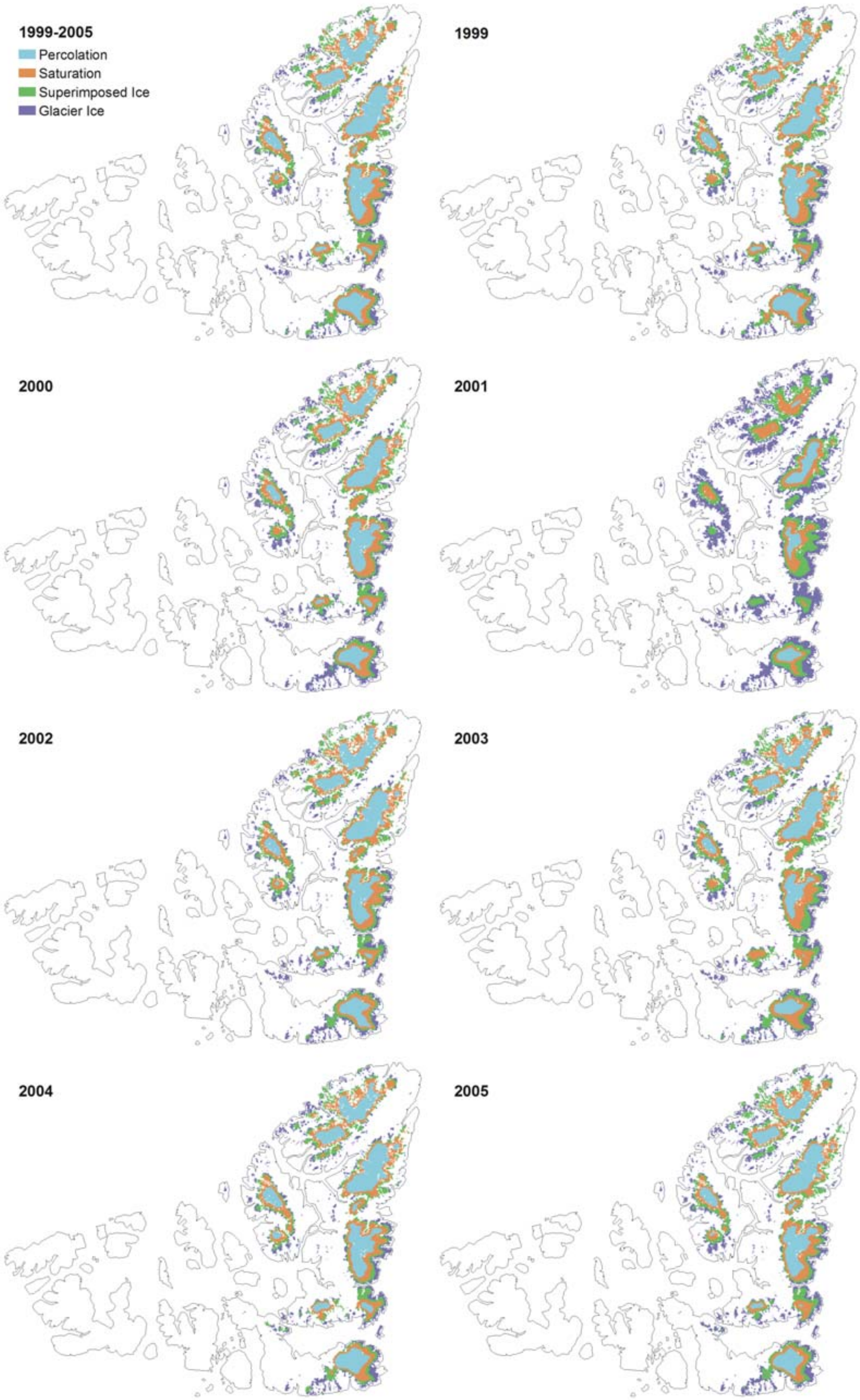

Figure 1. *Maps of snow and ice facies distribution on ice caps in the QEI, 1999-2005*.

Snow and ice facies classification results indicate that the glacier ice zone is the most extensive facies in the QEI and represents 26.7% of the total ice covered area, followed by the saturation, percolation, and superimposed ice zones (26.0%, 23.7%, 23.6%, respectively). For the QEI as a whole, variability in facies area is greatest for the glacier ice zone and least for the superimposed ice zone. Area changes in the superimposed and glacier ice zones are negatively correlated with changes in the percolation and saturation zones. Analysis with NCEP/NCAR Reanalysis reveals that changes in geopotential height in the troposphere (700, 500, and 300 hPa) and air temperature (700 hPa level) are positively (negatively) correlated with area changes in the glacier ice (percolation and saturation) zones and changes in facies boundary elevation.

The results from this study underscore the importance of understanding the complex and dynamic linkages between climatic forcings, changes in the troposphere, and ice cap surface responses. We present here, an efficient means of monitoring the inter-annual variability in snow and ice surface properties resulting from changes in melt conditions on Arctic ice caps. Mapping annual facies distributions on Arctic ice caps provides a more detailed view of seasonal melt conditions than maps of melt extent alone, and can be useful for estimating inter-annual changes in surface mass balance on ice masses in remote polar regions.

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Inventory and near-terminus velocity estimates of tidewater glaciers in Arctic Canada

Gabriel Wolken, Martin Sharp, and James Davis

Department of Earth and Atmospheric Sciences
University of Alberta, Canada

Current and future global estimates of land ice mass loss do not include mass loss from iceberg calving by tidewater glaciers. Recent studies, however, point to rapid changes in ice dynamics associated with tidewater glaciers as a causal mechanism for increases in ice sheet mass loss and contribution to sea level rise. Many of these dynamical changes are related to thinning of the terminus and retreat of tidewater glaciers, which may lead to additional changes in ice dynamics and continued retreat of tidewater glaciers. Although the data are limited, it is thought that mass loss from tidewater glaciers may account for a considerable portion of the total mass loss from major Arctic ice masses. Ice caps in Arctic Canada represent the largest ice covered area on Earth outside Greenland and Antarctica, yet it is not know what fraction of mass loss from these ice caps is attributable to tidewater glaciers.

Optical satellite remote sensing was used to determine the fraction of ice cap area that drains directly to the ocean through tidewater glaciers and to classify tidewater glacier fronts according to surface feature indicators of terminus velocity (Blaszczyk and Jania, 2008), in order to provide first order estimates of the possible mass loss from tidewater glaciers in the Canadian High Arctic (Tables 1, 2, and 3; Figures 1 and 2).

Ice Cap	TW-Count	Area (km^2)
Agassiz	20	9492.3
Axel	2	1060.0
Coburg	9	96.1
Devon	60	9423.9
Manson	55	4389.3
NElles	56	9962.6
POW	71	12857.0
Sydkap	6	607.9

Table 1. *Total area drained by tidewater glaciers and the number of tidewater glaciers for ice caps in the Canadian High Arctic.*

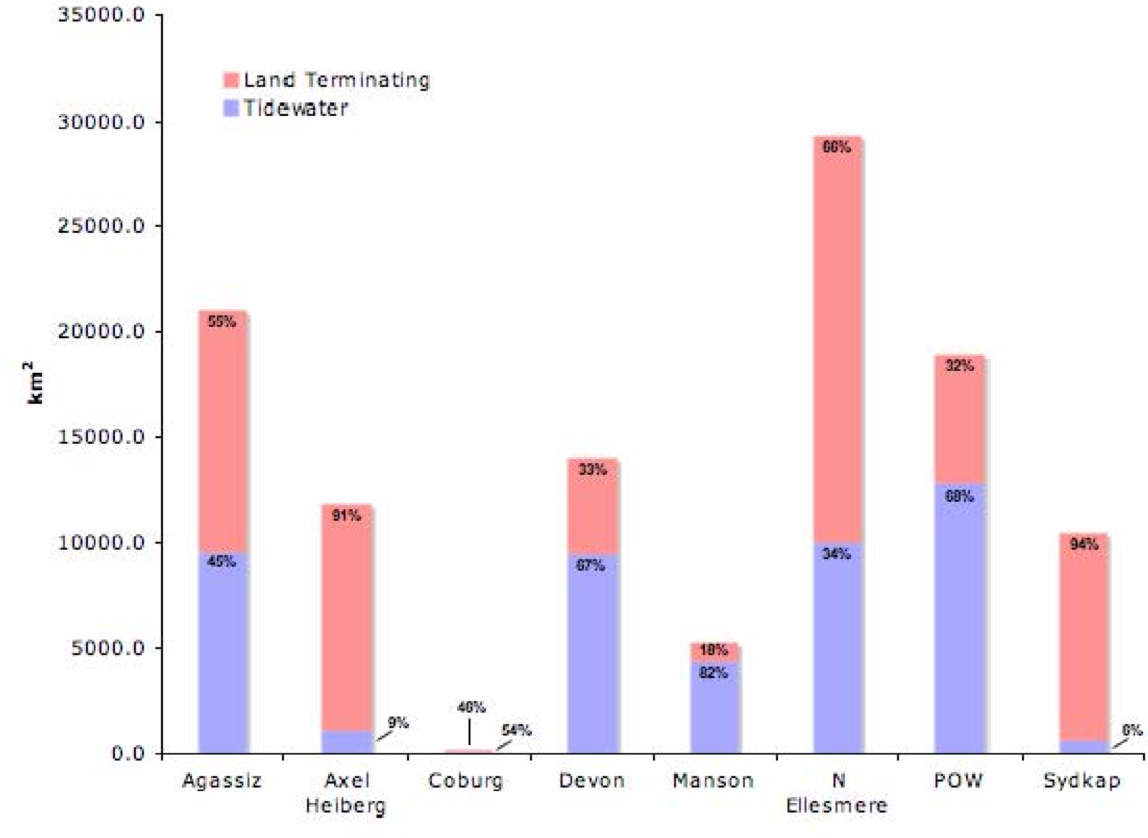

Figure 1. *Areal fractions of land-terminating and tidewater glaciers for each ice cap in the Canadian High Arctic.*

TW- Type	Description	Length of Crevassed Zone (km)	Est. Terminus Velocity (m a^{-1})
I	very slow	0 - 0.3	10 ± 5
II	slow	>0.3 - 1.0	70 ± 30
III	fast	>1.0 - 5.0	200 ± 50
IV	very fast or surging	>5.0	>300 ± ?

Table 2. *Tidewater glacier types classified according to estimated velocities. Estimated terminus velocities are based on the relationship between the length of crevassed zone and published near terminus velocities (r = 0.66; n = 42).*

TW- Type	TW-Count	Area (km^2)	% Total Area
I	136	10218.1	21.3
II	39	2456.3	5.1
III	63	8985.2	18.7
IV	41	26367.4	54.9
Total	279	48026.9	100.0

Table 3. *Total number and area of tidewater glaciers for each tidewater glacier type.*

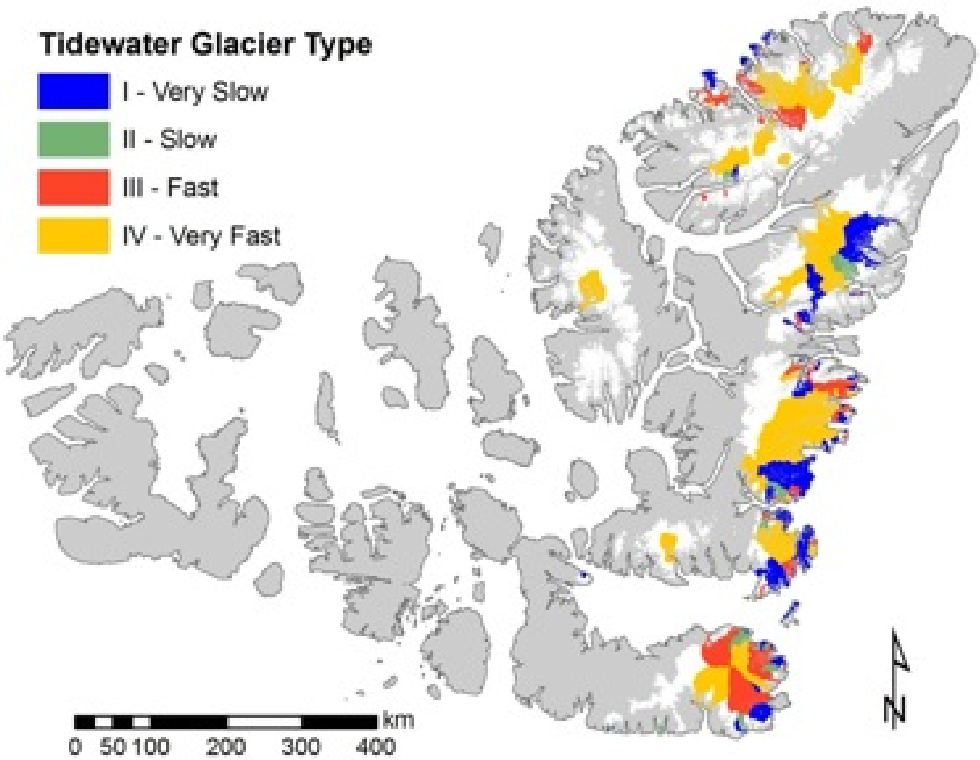

Figure 2. *Map of tidewater glaciers in the Canadian High Arctic. Tidewater glaciers are coloured according to tidewater glacier type (see Table 2).*

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