SSD
SCIENCE FOR A SUSTAINABLE DEVELOPMENT

# ANTARCTIC SUBGLACIAL PROCESSES AND INTERACTIONS: ROLE OF TRANSITION ZONES IN ICE SHEET STABILITY

## "ASPI"

T. BOEREBOOM, B. DE SMEDT, PH. HUYBRECHTS, A. HUYGHE,
F. PATTYN, L. PERICHON, J.-PH. REMY, D. SAMYN, J.-L. TISON,
H. VAN DE VYVER.

ENERGY

TRANSPORT AND MOBILITY

AGRO-FOOD

HEALTH AND ENVIRONMENT

CLIMATE

BIODIVERSITY

ATMOSPHERE AND TERRESTRIAL AND MARINE ECOSYSTEMS

TRANSVERSAL ACTIONS

.be

SCIENCE FOR A SUSTAINABLE DEVELOPMENT
(SSD)

*Antarctica*

FINAL REPORT PHASE 1

# ANTARCTIC SUBGLACIAL PROCESSES AND INTERACTIONS: ROLE OF TRANSITION ZONES IN ICE SHEET STABILITY
# "ASPI"

**SD/CA/02**


**Promotors**

**Frank Pattyn, Jean-Louis Tison**
Université Libre de Bruxelles (ULB)
Laboratoire de Glaciologie

**Philippe Huybrechts**
Vrije Universiteit Brussel (VUB)

Vakgroep Geografie – VUB

**Contributing authors**

Thierry Boereboom (ULB), Bert De Smedt (VUB),
Philippe Huybrechts (VUB), Ann Huyghe (VUB), Frank Pattyn (ULB),
Laura Perichon (ULB), Jean-Philippe Remy (ULB), Denis Samyn (ULB),
Jean-Louis Tison (ULB), Hans Van De Vyver (VUB)


BELGIAN SCIENCE POLICY

ULB

Vrije
Universiteit
Brussel

BELGIAN SCIENCE POLICY

Rue de la Science 8
Wetenschapsstraat 8
B-1000 Brussels
Belgium
Tel: +32 (0)2 238 34 11 – Fax: +32 (0)2 230 59 12
http://www.belspo.be

Contact person: Maaike Vancauwenberghe
+32 (0)2 238 36 78



T. Boereboom, B. De Smedt, Ph. Huybrechts, A. Huyghe, F. Pattyn, L. Perichon, J.-Ph. Remy, D. Samyn, J.-L. Tison, H. Van De Vyver. ***Antarctic Subglacial processes and interactions: role of transition zones in ice sheet stability "ASPI"*** Final Report Phase 1. Brussels : Belgian Science Policy 2009 – 42 p. (Research Programme Science for a Sustainable Development )

## CONTENTS

## ABSTRACT

The aim of ASPI (Antarctic Subglacial Processes and Interactions) is (i) to understand the interactions between the ice sheet and the subglacial environment and the processes that control the Antarctic ice sheet, and (ii) to quantitatively determine the stability of the ice sheet in a changing climate and the impact of climatic variations on the coastal ice sheet. A key factor in both such qualification and quantification is the existence of transition zones within the ice sheet, such as grounding lines, i.e. the interface between the ice sheet and an ice shelf, between an ice sheet and a subglacial lake, as well as between an ice shelf and its pinning points. Especially grounding lines in marine ice sheets, such as the West-Antarctic Ice Sheet (WAIS), are extremely vulnerable, as small perturbations at the grounding line, such as basal melting or variations in sea level, might lead to an important grounding line retreat, as has been observed recently in the Amundsen Sea sector. ASPI therefore investigated the factors that influence the physics and mechanics of the grounding lines, such as rheology and the impact of marine ice formation in rifts at the grounding line (which could either stabilize or destabilize the ice flow), as well as how to represent the processes that are responsible for grounding line migration in ice sheet models.

On behalf of modelling and model development, we analyzed the response of a marine ice sheet to different perturbations near the grounding line using a numerical ice sheet model that takes into account longitudinal stress coupling and grounding line migration. The model is based on an existing flowline model, but extended with a novel subgrid determination of the grounding line position and migration as a function of the size of the transition zone between the ice sheet and the ice shelf. For instance, a wide transition zone is typical for an ice stream, a small transition zone typical in many areas around the East Antarctic ice sheet where the ice sheet rather suddenly changes into an ice shelf. Model results show that stress transmission or longitudinal coupling across the grounding line plays a decisive role. The grounding line migration is a function of the length scale over which the basal conditions change from frozen to the bed to floating, the “transition zone”. Perturbations at the grounding line, such as reduction in buttressing of the ice shelf, substantially thins the grounded ice sheet. Marine ice sheets with large transition zones - such as ice streams - seem highly sensitive to such perturbations, compared to ice sheets with small transition zones, such as an abrupt ice sheet/ice shelf junction.

Deformation experiments are meant to shed more light on how marine ice formation at the grounding line influences the flow properties of the ice sheet system in the transition zone. Therefore, a new analysing facility was devised and tested and is installed during ASPI Phase I at the ULB laboratory. Preliminary experiments clearly demonstrate the importance of such marine ice inclusions on the ice flow: marine ice is harder to deform than meteoric ice in a typical ice-shelf stress field (vertical uniaxial compression with lateral and longitudinal extension) and may therefore exert a stabilizing effect on the whole system. The importance of such result is in conjunction with results from inverse modelling of the transition zone. The latter technique allows for a determination of the ice viscosity properties and/or the basal characteristics in the transition zone, based on observed ice sheet configuration (ice thickness, surface topography) and observed surface velocities. Inverse modelling is therefore capable of determining the size of the transition zone, which is an important factor for predictive modelling. For instance, the application to Pine Island Glacier (WAIS) clearly allows us to determine to position of the onset of the ice stream as well as to delineate areas where stress coupling is essential.

But not only is the ocean influence an important boundary condition. Basal thermal conditions influence to a large extent the behaviour of the Antarctic ice sheet. Whenever the ice reaches pressure melting point, melt water is generated that may lead to enhanced ice flow. Also the presence of subglacial lakes is governed by the basal thermal conditions. Using a 3D thermomechanically-coupled ice sheet model, the influence of spatial variability of the geothermal heat flux on the basal temperature regime of the Antarctic ice sheet was investigated.

Results were compared with observations of known basal temperatures (e.g. ice core drill sites) as well as the spatial distribution of subglacial lakes. This way it was possible to determine what dataset is more suitable for future model experiments and what basal conditions are more likely to reign underneath the Antarctic ice sheet.

Subglacial lakes are yet another type of transition zones that are currently gaining attention. Recent observations demonstrate that subglacial lakes may drain and add significant amounts of basal water to the subglacial hydrological system. Hence, they have the potential of destabilising ice sheets through sudden lake outbursts, of which evidence exist along the coast of the East Antarctic ice sheet. In ASPI we investigate the effect of subglacial lake drainage on the stability of the ice sheet and especially how sensitive the subglacial lake system is towards drainage and flooding. Preliminary experiments show that only slight changes in surface topography might easily lead to partial drainage of such a lake.

Transition zones in the Antarctic ice sheets, whether it be grounding lines, subglacial lakes or the subglacial interface are key elements in the dynamic behaviour of the Antarctic ice sheet and its stability. Although this is the first phase of the ASPI project, we start to better understand the subglacial processes and interactions that occur at these interfaces. They seem even more important as a controlling factor as previously thought of.

# INTRODUCTION

## Context

The objectives of ASPI meet those of the third scientific support plan for a sustainable development (Global change, biodiversity and ecosystems) and the CliC (Climate and Cryosphere) project of the WCRP (World Climate Research Programme). ASPI forms part of international research programs such as the European Project on Ice Coring in Antarctica (EPICA), SCAR-SALE (Subglacial Antarctic Lake Environments, a core project of the Scientific Committee on Antarctic Research), and ISMIP-HOM (Ice Sheet Model Intercomparison Project for Higher-order ice-sheet Models), a project of the Numerical Experimentation Group of WCRP/SCAR CliC. Finally, the results from the paleo work are a direct contribution to SCAR-ACE (Antarctic Climate Evolution).

## Objectives

The aim of ASPI (Antarctic Subglacial Processes and Interactions) is (i) to understand the interactions between the ice sheet and the subglacial environment and the processes that control the Antarctic ice sheet, and (ii) to quantitatively determine the stability of the ice sheet in a changing climate and the impact of climatic variations on the coastal ice sheet.

A key factor in such quantification and impact assessment is the existence of transition zones within the ice sheet. Such transition zones are examples of specific boundary layers widely found in glaciology. Basically they are parts of the ice sheet which overlie basal transition zones where the flow is anomalous. Typical examples of such transition zones are the grounding lines, i.e. the interface between the ice sheet and an ice shelf, between an ice sheet and a subglacial lake, as well as between an ice shelf and its pinning points. These transition zones are probably among the least understood elements of ice sheets, although they determine to a large extent the processes and dynamics of lateral expansion and retreat of ice sheets as well as the stability of marine ice sheets. Apart from their role in ice dynamics and ice sheet stability, processes and interactions within basal transition zones also hamper the interpretation of the paleoclimatic signal as recorded in deep ice cores. Basal deformation is responsible for disturbing this signal and understanding the processes at the base of ice sheets enables such signal recovery. The subglacial environment opens up new frontiers in Antarctic explorations, as this dynamic and extreme interface still needs to be explored in terms of glaciological, geological, geochemical and biological research efforts.

# RESULTS

## Grounding lines and ice sheet modelling

### A new algorithm for grounding line migration in ice sheet models

Marine ice sheets rest on a bed that lies well below sea level and the drainage of the ice sheet (necessary to balance the ice accumulation) takes place through surrounding ice shelves. Marine ice sheet stability is believed to be controlled by dynamics of the grounding line, i.e. the junction between the grounded ice sheet and the floating ice shelf. The WAIS has therefore received particular attention because it has been the most dynamic part of the Antarctic ice sheet in the recent geological past, and because most of it is grounded below sea level - a situation that, according to early models, could lead to flow instabilities and rapid ice discharge into the ocean when the surrounding ice shelves would weaken. Such instability is governed by the physical interaction between ice streams and shelves, more precisely the feedback between grounding lines and ice flux. The potential of the WAIS to collapse in response to future climate change is still a subject of debate and controversy. If the entire WAIS would collapse, global sea level would rise by 5 to 6 m. It has been suggested that present observed rapid thinning of the WAIS is due to increased melting under ice shelves caused by a gradual ocean warming (Shepherd et al., 2004). Payne et al. (2004) showed that such melting could lead to an acceleration of grounded ice flow.

Grounding line migration is a key element in assessing the stability of marine ice sheets, and the WAIS in particular. We believe that stress transfer or longitudinal stress coupling across the grounding line plays a decisive role herein and is determined by the length of the transition zone between ice sheet and ice shelf. Such transition can be small, i.e. an abrupt change from ice sheet to ice shelf, or large when an ice stream makes the linkage between both systems. Several theories exist with respect to the behavior of marine ice sheets, in which some even neglect the presence of an ice shelf. For this project we analyzed the response of a marine ice sheet to different perturbations near the grounding line using a numerical ice sheet model that takes into account longitudinal stress coupling and grounding line migration. The model is based on the flowline models developed by Pattyn (2002), but extended with a novel subgrid determination of the grounding line position and migration as a function of the size of the transition zone between the ice sheet and the ice shelf (Pattyn et al., 2006). Consider the function based on flotation criterion (with $h$ ice thickness, $z_{sl}$ sea level and $\rho_i$ and $\rho_w$ the density of ice and sea water, respectively:

$$f = \frac{(z_{sl} - b)\rho_w}{\rho_i h}$$

It follows that $f = 1$ at the grounding line, $f < 1$ in the grounded ice sheet, and $f > 1$ in the ice shelf. The exact position of the grounding line $x_g$ (km) is then defined by linear interpolation between the last grounded and the first floating grid point:

$$x_g = \frac{1 - f_j + \nabla f \, x_j}{\nabla f}$$

where $\nabla f = \left(f_{j+1} - f_j\right)/\Delta x$. This scheme has the advantage that it can be easily translated to the plane. The difference between the ice sheet and shelf arises because the shelf is floating and because it experiences vanishing tangential traction on the bottom as well as the upper surface (Hindmarsh, 1993). The length of the transition zone is then determined by the distance over which this vanishing traction occurs. Vanishing traction towards the grounding line is introduced through a friction parameter $\beta^2$ that relates the basal velocity to the basal drag, $\tau_b = u_b \beta^2$. For $\beta^2 = 0$, basal drag equals zero (no traction) which is a situation similar to an ice shelf; for $\beta^2 = +\infty$, the basal velocity equals zero (ice frozen to the bedrock) which is valid for the ice sheet; for $0 < \beta^2 < +\infty$ a transition zone exists. We shall not consider the physical mechanisms that are responsible for the reduction in traction (i.e. sliding) but suppose basal friction to be of the general form: $\beta^2 = \exp\left[\beta_0\left(x_g - x\right)\right]$, where $\beta_0$ controls the length of the transition zone. The exponential form of the basal friction stems from an analysis of the longitudinal profile of PIG (Pattyn et al, 2006). Linking the subgrid determination of the position of the grounding line directly with the stress field within the ice sheet is a novel approach. There is no need to *a priori* prescribe what is ice sheet, ice stream or ice shelf and the use of these relationships is independent of the chosen grid size, which make the basal conditions independent of grid size as well. Values of $\beta_0$ and their relation to observed Antarctic transition zones can be found in Table 1 in Pattyn et al (2006). The size of the transition zone will also have an impact on the resulting ice sheet / ice shelf profile: the more concave shape is due to a wider transition zone, as can be observed in ice streams as well (Figure 1).

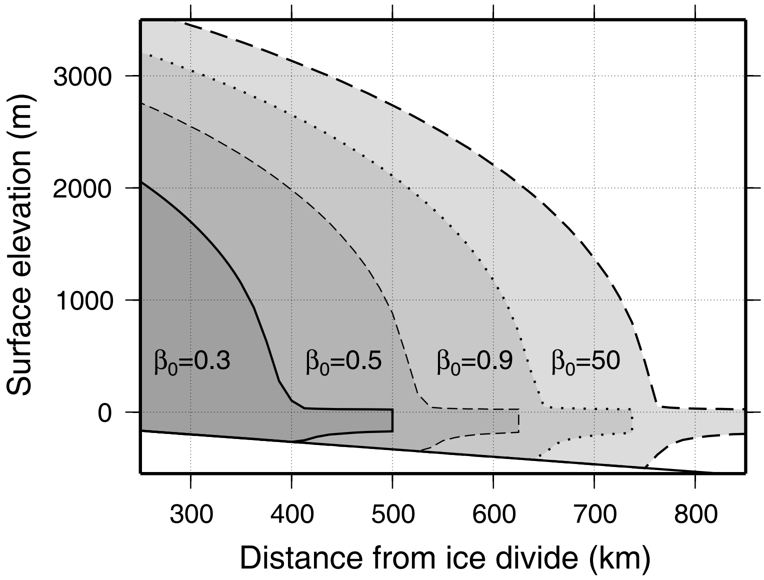


*Figure 1: Equilibrium profiles of a marine ice sheet on a sloping bedrock for different sizes of the transition zone between the ice sheet and the ice shelf. All experiments started from a fixed grounding line at x = 750 km.*

Experiments were carried out for different sizes of transition zones, which has a major impact on the shape of the longitudinal profile of the ice sheet/ice shelf system: the shape of the transition becomes more concave when the transition zone is wide, a typical feature for the ice stream/ice shelf transition. Our basic sensitivity analysis has shown that grounding line advance and retreat is a function of transition zone characteristics. The retreat is minimal for small transition zones, and the grounding line remains more or less where it is. However, the reversibility of the process increases with the length of the transition zone. These results corroborate the neutral equilibrium hypothesis of Hindmarsh (1993), i.e., that for a small transition zone multiple equilibria may exist, therefore not necessarily exhibiting reversibility. For larger transition zones the grounding line returns to its initial position, which shows the reversibility of the process, similar to the behavior of moving grid models (Vieli and Payne, 2005). Thus, according to our model results, grounding line migration for small transition zones exhibits hysteresis, while for larger transition zones the migration process is reversible. Stress transmission or longitudinal coupling across the grounding line plays also a decisive role. The grounding line migration is a function of the length scale over which the basal conditions change from frozen to the bed to floating, the "transition zone". We have demonstrated that thinning of the ice shelf due to bottom melting has a negligible effect on the grounded ice mass, at least when other parameters do not come into play. Only perturbations at the grounding line or reduction in buttressing of the ice shelf substantially thins the grounded ice sheet. Inclusion of lateral drag does not alter these results qualitatively. Marine ice sheets with large transition zones - such as ice streams - seem highly sensitive to such perturbations, compared to ice sheets with small transition zones, such as an abrupt ice sheet/ice shelf junction. Ice streams are therefore more sensitive to changes at the grounding line and form therefore a potential positive contribution to future sea-level rise. This is illustrated in Figure 2.

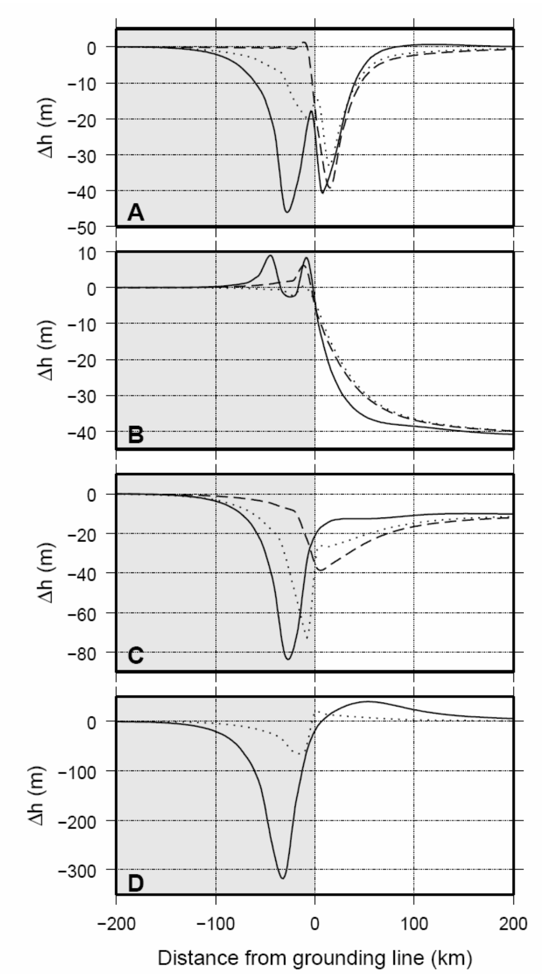


*Figure 2: Ice sheet response near the grounding line to perturbation for $\beta_0$ = 50 (dashed line), 0.9 (dotted line) and 0.3 (solid line) after 100 year. (A) Melting of 1 m/a at the grounding line; (B) Melting of 0.5 m/a along the whole length of the ice shelf; (C) Reduced ice shelf buttressing (see text for details); (D) Decrease in $\beta_0$, from $\beta_0$ = 0.9 to 0.7 and from $\beta_0$ = 0.3 to 0.2. The grounded ice part is shaded.*

Validating stress coupling across the grounding line

Although predictions of sea-level rise due to global warming have become more precise over the last decade, as highlighted by the recent report from the Intergovernmental Panel on Climate Change (IPCC), the contribution that could arise from rapidly changing flow in ice sheets, especially in Greenland and Antarctica, remains excluded (Vaughan and Arthern, 2007). Furthermore, it is apparent that the late 20th- and early 21st-century ice sheets at least are dominated by regional behaviors that are not captured in the models on which IPCC predictions have depended (Shepherd and Wingham, 2007). The viscous flow of ice is rather well understood on a theoretical level, but the difficulties in modeling the ice sheet system arises from the specification of stress boundary conditions at the base and the seaward margin. Here, several stress components come into play in regions of high variability in basal topography and/or basal slipperiness.

Despite the lack of comprehensive predictive ice-sheet modeling, the ice-sheet modeling community has evolved considerably over the last decade. Increasing computational power has led to the development of more complex ice sheet models, with varying degrees in approximations to the Stokes system. Although only few models solve the full Stokes problem in three dimensions, there is a need for validating these so-called higher-order models (i.e. models that incorporate further mechanical effects, principally longitudinal stress gradients) as analytical solutions are not always available. Such a benchmark exercise was done with large-scale ice sheet models in the 1990s (e.g. Huybrechts et al., 1996). Similar benchmark experiments are carried out now and validated by $>22$ numerical ice-sheet models of varying degrees of complexity (Pattyn et al., 2008). Besides benchmarking, the experiments also allowed to discern under which conditions the approximations to the Stokes system are viable and whether numerical issues play a role in the result.

Implementation of grounding line migration in a 3D ice sheet model

A preparatory literature survey was undertaken to explore ways to implement an improved grounding line treatment in a 3-D Antarctic ice sheet model (Huybrechts, 2002). The mechanical problem has long been stated from the fundamentally different stress balance in floating and grounded ice. While the grounded ice is dominated by vertical shear, ice-shelf flow is a buoyancy-driven flow dominated by longitudinal stretching and lateral shearing (e.g. Van der Veen, 1999). This implies a transition zone across the grounding zone which dimensions are crucially controlled by the importance of basal sliding in the adjacent ice sheet (e.g. Mayer and Huybrechts, 1999; Pattyn et al., 2006). So far grounding line migration in existing 3-D Antarctic ice sheet models is only implemented in the models operated in Grenoble (Ritz et al., 2001) and in Brussels/ Bremerhaven (Huybrechts, 1992; Huybrechts and de Wolde, 1999; Huybrechts, 2002). In both models, the position of the grounding-line is diagnosed from a floatation criterion. Ritz et al. (2001) introduced the concept of a 'dragging ice shelf' to incorporate ice-stream dynamics, which is particularly important for the Siple Coast area of the West Antarctic ice sheet. In their model, inland ice is differentiated from an ice-stream zone by the magnitude of basal drag. This is based on the observation that ice-stream zones are characterized by low surface slopes, and thus low driving stresses, but have fast sliding. Ritz et al. (2001) treat these zones as semi-grounded ice shelves, and replace the shallow-ice approximation by a set of equations for ice-shelf flow to which basal drag is added. In the Huybrechts models two sets of equations are solved independently for grounded and floating ice. The transition zone at the grounding line is assumed to consist of one grid point only, located directly adjacent to the ice shelf. Here, longitudinal stresses are taken into account in the effective stress term of the flow law, which additional stress terms are found by iteratively solving three coupled equations for depth-averaged horizontal stress deviators. Gross model behaviour of both models turns out to be quite similar, except that the Grenoble model has a lower surface slope near the grounding line in West Antarctica and that the break in the slope occurs further upstream at the place where the dragging ice shelf joins the inland ice subject to the shallow-ice approximation.

One consequence is that grounding-line retreat in the Grenoble model occurs more readily in response to rising sea levels.

More recent work has predominantly concentrated on numerical issues, mostly for schematic geometries along a 2-dimensional flow line. In Hindmarsh and Le Meur (2001), the grid resolution problem was overcome by using a moving grid model, which allows the grounding line to be followed continuously using an expression for the grounding line migration rate to compute the grounding line motion. The grounding line migration rate was found from a total differentiation of the floatation condition following Hindmarsh (1996). In their approach there is no need for an ice shelf, as the position of the grounding line is solely determined from the geometry at the grounding line, the local ice flux and accumulation rate. The influence of the ice shelf on the inland ice, such as from buttressing, is ignored.

Nearly 10 years after the EISMINT experiments (Huybrechts et al., 1998), Vieli and Payne (2005) reconsidered a comparative study of the ability of numerical ice sheet models to simulate grounding line migration. They found that the dynamics of the grounding line are mainly controlled by the discretization scheme used and that the physics (such as longitudinal coupling) incorporated into a particular model appears to have only secondary importance. Moreover, moving grid models perform fundamentally different from fixed grid models, the latter allowing for larger changes in response to an imposed forcing. Another thing which became very clear already from the EISMINT tests is that grid resolution of the transition zone is also of crucial importance in any properly functioning scheme.

Pattyn et al. (2006) offer a trade-off between fixed grid models and moving grid models (see supra). By modelling a flow line in a higher-order model, vertical shear and longitudinal stress gradients are both included in the transition zone. This idea is much related to the work of Herterich (1987) where the vertical gradient was taken to be a linearly increasing function of the distance upstream from the grounding line. In principle these concepts are generalisable to 3 dimensions. The length of the transition zone and the location of the grounding line then take the form of a curved plane.

Schoof (2007a) applied boundary layer theory in order to obtain analytical approximations for the ice flux in the transition zone. In Schoof (2007b), three different models are employed based on a moving grid. Grid refinement in the transition zone is again identified as a critical component in obtaining reliable numerical results. Mathematically spoken, the models are free boundary value problems which means that the moving boundary is also a part of the problem. Schoof remarks that parabolic free boundary value problems, such as the Stefan problem, require two boundary conditions. This is in contrast to the moving grid models of Hindmarsh (1996), Hindmarsh and Le Meur (2001) and Vieli and Payne (2005) who use only one boundary condition. Based on this observation, and on a suggestion for the second boundary condition, Schoof (2007b) uses an analytical approximation to describe steady states, stability and hysteresis. His work supports several earlier theories predicting instability, cf. Weertman (1974).

Based on the work by Schoof, Pollard and DeConto (2007) embedded a relation for the flux across the grounding line in a higher-order model, but using an equidistant 10 km numerical grid. This approach produced very similar results to another experiment, in which a very fine 0.1 km grid was used around the grounding line, but not the boundary condition on the grounding line flux. From these experiments, it may be concluded that the reasons for unrealistic behaviour produced by earlier models (such as those included in EISMINT) at the transition zone is not the resolution problem itself, but some parallel effect. Large grid sizes cause an inaccurate flux at the grounding line, which can only be overcome by using a numerically-demanding high resolution grid or by imposing the flux itself. The disadvantage of the Schoof approach, however, is that it can not incorporate buttressing effects. Nor is it straightforward to generalize it to three dimensions as the flux across the grounding line can not be established in an evident way.

However, Schoof's approach makes it possible to test models, and any new model that is going to be built should be conform theoretical results. For instance, Schoof (2007b) found that marine ice sheets may undergo dramatic and irreversible changes - hysteresis - under changes in physical properties (sliding, ice viscosity) or external forcing (accumulation rates, sea levels). The basic questions that arise from these papers and earlier work are the following:

- Do marine ice sheets have one or more distinct equilibrium shapes, or do they not exhibit `neutral equilibrium': will a perturbation in grounding line position away from steady state result in the grounding line either returning to the (stable) steady state position or migrating away from the (unstable) steady state position to a stable steady state?
- Do stable steady states have to have their grounding lines in a region of downward-sloping bed?
- How do equilibria depend on bed geometry and the physics of sliding at the bed, ice viscosity and gravitational forces as well accumulation rates?
- Is hysteresis under changes in forcings and internal physical properties possible when the bed is overdeepened?
- To what extent is high grid resolution, especially near the grounding line, necessary to obtain reliable results? Is this particularly important when modelling transients?

Experiments with the above described flowline model have been carried out to test the hypothesis of stable grounding line equilibria on a downward sloping bed as well as on an overdeepened bed that should exhibit hysteresis (combination of stable and unstable equilibria). This also allows us to test whether the stress balance matters in the transition zone or not (i.e. can we just use the shallow-ice approximation (SIA) to solve the stress field upstream of the grounding line or do we explicitly have to take into account longitudinal stress gradients?). The impact of higher-order stress gradients is shown in Figure 3, which displays the evolution of steady state grounding line positions for different values of the ice flow factor A. Starting from an initial ice sheet configuration the parameter A was changed and the ice sheet/ice shelf system evolved until a new steady state was reached. For decreasing values of A, the grounding line moves forward (as depicted by the blue dots in Figure 3). However, using a SIA model – shallow ice approximation – increasing values of A do not lead to a grounding line retreat and the grounding line remains more or less at its initial position (Figure 3a). Using the higher-order model with the grounding line interpolation scheme as described above, leads to a reversible grounding line migration (Figure 3b), conform the theoretical considerations described by Schoof (2007b).

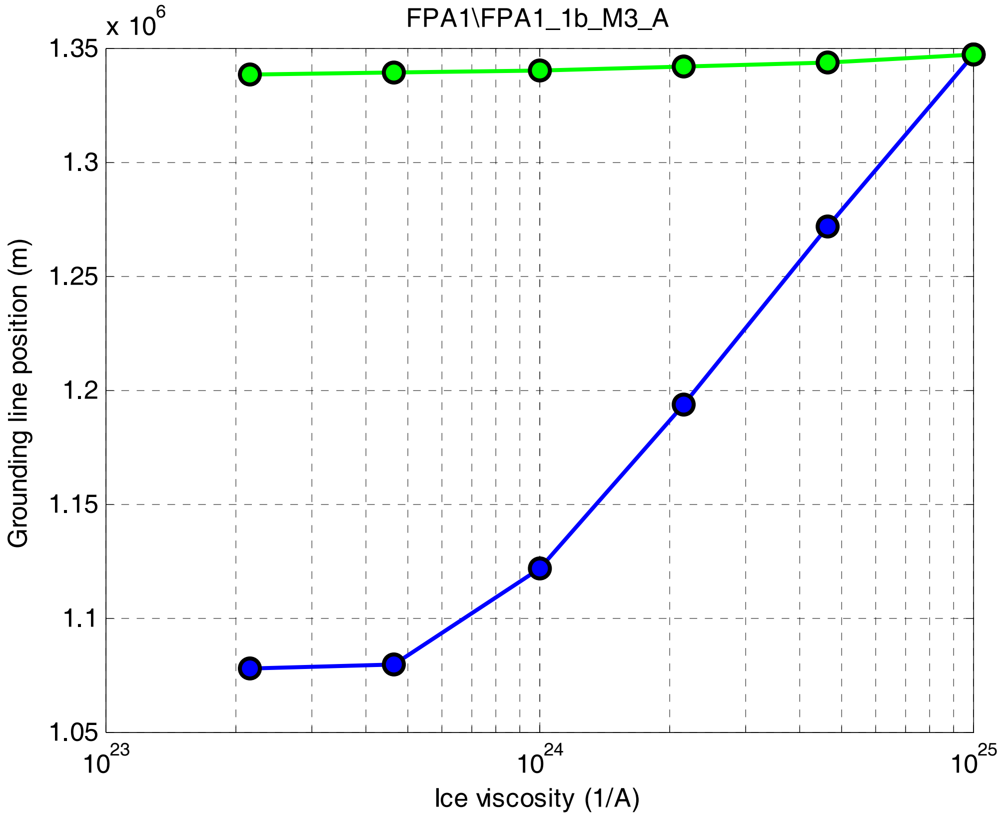


*Figure 3a: steady state grounding line positions for different values of the ice flow parameter A. The blue curve shows the evolution of stepwise decreasing values of A. the green curve the reverse evolution. Results are shown for a SIA model with coupled ice shelf model.*

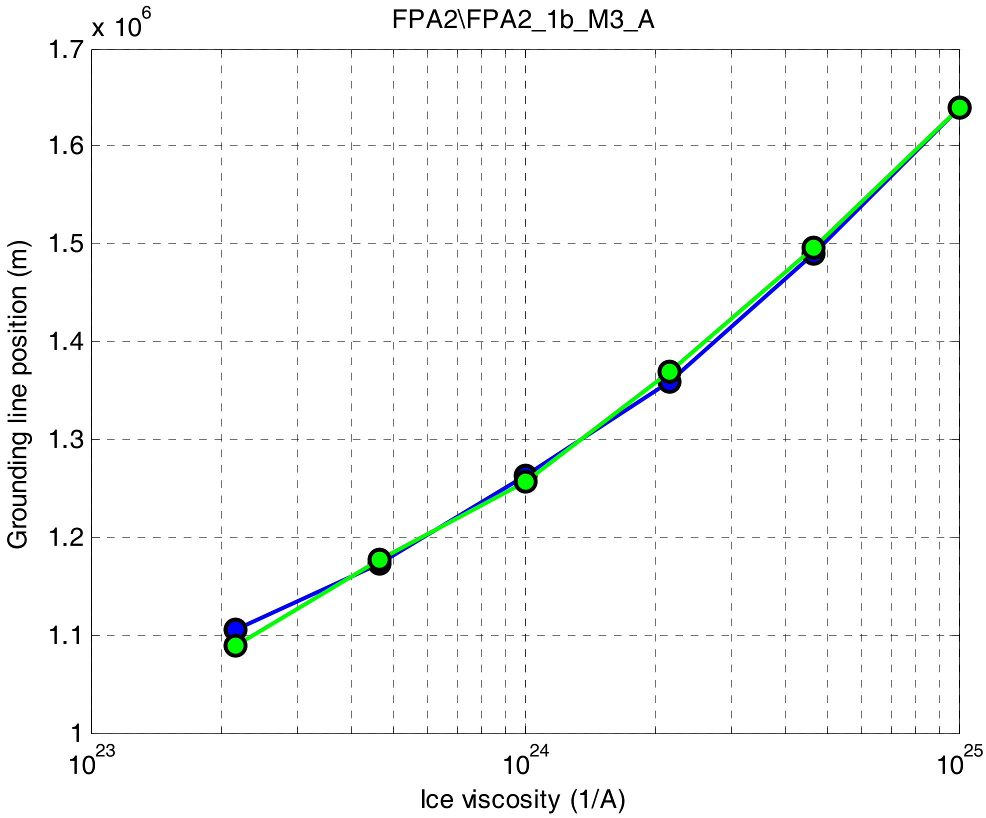


*Figure 3b: as Figure 3a, but with the higher-order model. Note that increasing and decreasing values of A result in similar steady state ice sheet configurations with the grounding line at more or less the same position.*

Building further on these studies, we have tested various approaches to implement a new scheme for grounding-line migration in three dimensions. This new scheme is expected to combine a high-resolution fixed grid with a higher order stress solution in the transition zone using a nested modeling approach similar to the one adopted recently for dating and interpreting the EPICA Dronning Maud Land ice core (Huybrechts et al., 2007).

Impact of grounding line evolution

It is generally expected that the amount of basal sliding plays a crucial role in the evolution of the Antarctic ice sheet. In the three-dimensional thermomechanical ice-sheet model (Huybrechts, 2002) basal sliding is restricted to basal areas which are within 1°C of the pressure melting point. Basal sliding is considered to be of Weertman-type, relating basal sliding to basal shear stress to the third power and inversely proportional to the reduced normal load due to the pressure of subglacial water. In a first series of experiments, we tested the influence of the multiplyer in the basal sliding law while keeping the value of the exponent unchanged. Changing the basal sliding parameter by an order of magnitude is found to have a critical impact on the simulated evolution over the last 30000 years (Figure 4). Considering a basal sliding parameter 10 times lower than in the reference experiment, all other things being equal, increases overall ice volume by 15% and moreover impedes grounding-line retreat in response to postglacial rising sea levels. The main reason is a steeper and thicker margin, especially in West Antarctica, which directly impacts on the height of buoyancy of the inland ice. Conversely, increasing the basal sliding parameter by a factor of 10 considerably thins the overall ice sheet and results in a smaller grounded ice sheet area. It can be concluded that the strength of basal sliding crucially impacts on the process of grounding-line retreat.

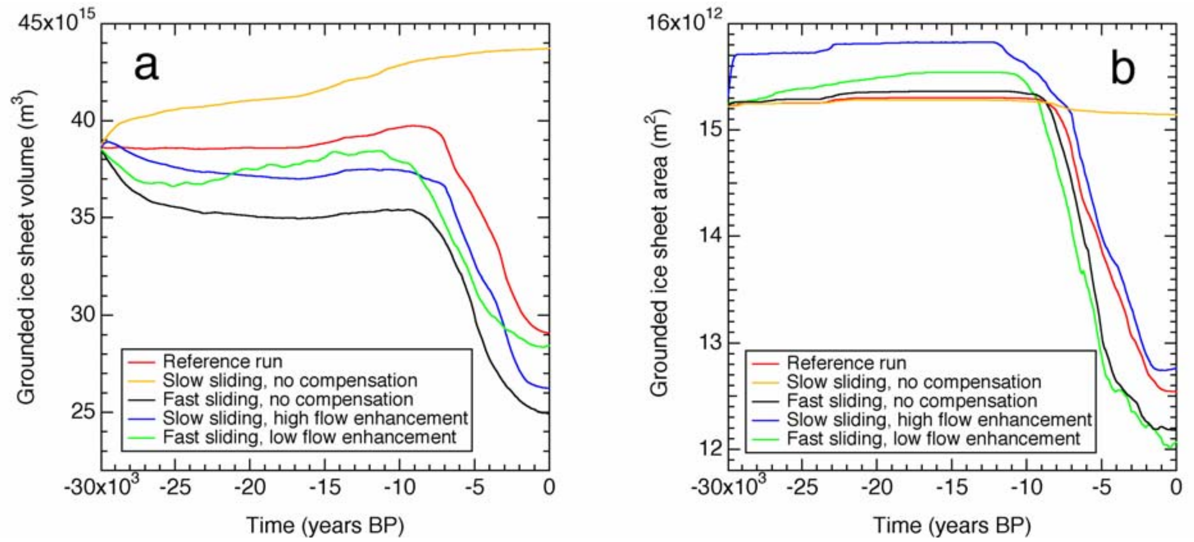


*Figure 4: Evolution of the Antarctic ice sheet over the last 30000 years in experiments with varying amounts of basal sliding that is/ or not compensated by commensurate changes in the rate factor for ice deformation. (a) grounded ice sheet volume; (b) grounded ice sheet area.*

To compensate for the overall volume change from changes in basal sliding, one can modify the stiffness of the ice, and hence, the rate factor for ice deformation in Glen's flow law. In the slow sliding case, a five-fold flow enhancement yields a rather similar ice volume and ice sheet area as in the reference run, and again produces grounding-line retreat in West-Antarctica. Fast sliding can be compensated by a flow enhancement with a factor 0.1 (making the ice 10 times harder), again producing a larger overall ice volume. As can be seen on Figure 5, this results in quite different surface geometries. In the fast sliding case one remarks a clear thinning for wet-based parts of the East Antarctic ice sheet and a thickening in those central dome locations frozen to bedrock. The effect is also clearly visible in West Antarctica, where the high-sliding case produces a surface elevation profile closer to observations than in the reference run. It must be noted, however, that owing to the long Antarctic response times, the geometries in these preliminary sensitivity tests have not yet fully adjusted to the parameter changes at 30 kyr BP, and are therefore out of equilibrium not only with the environmental forcing, but also with the internal model physics. We infer from these experiments the need to further investigate the effect of the sliding law exponent and the physical mechanisms controlling the sliding law multiplier, such as basal roughness and basal water depth.

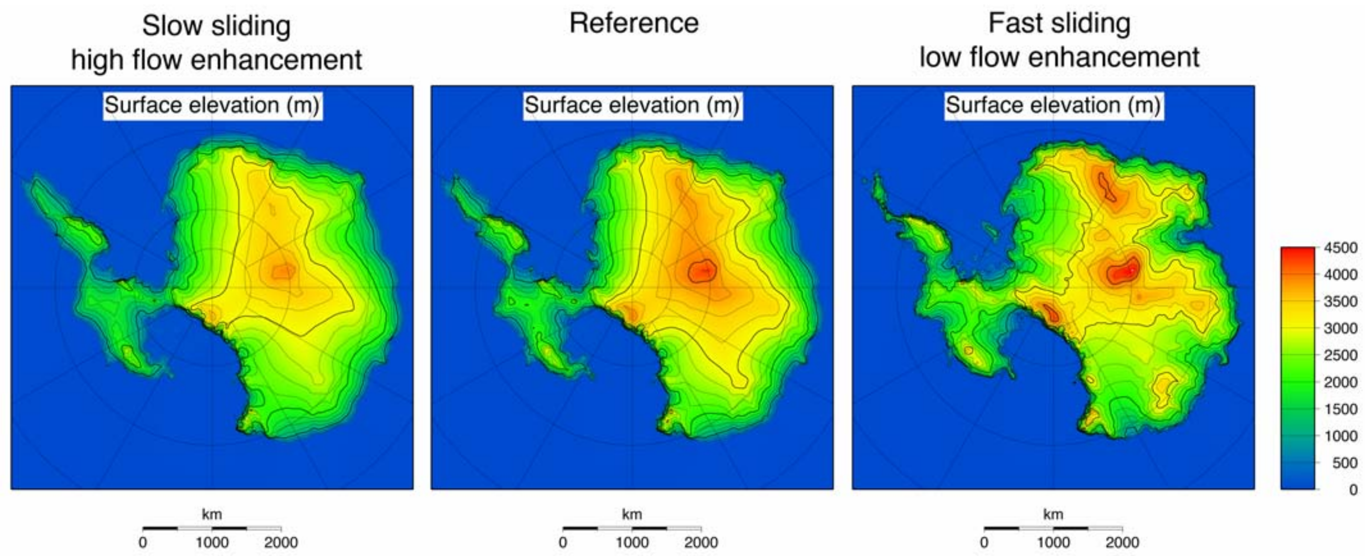


*Figure 5: Present-day modelled Antarctic ice sheet geometry in three experiments which varied the basal sliding parameter by an order of magnitude with respect to the reference run (middle panel). At the left is the resulting ice-sheet elevation with a ten-fold reduction of the sliding parameter, compensated by a five-fold increase of the rate factor for ice deformation. At the right is the result of an experiment with a ten-fold increase in the basal sliding parameter together with a ten-fold reduction of the rate factor.*

## Marine ice and its influence on ice shelf flow

### Set-up of an Ice Deformation Laboratory and fabric analyser

To ensure fast and low-cost build-up of the deformation unit, we opted for a compression rig operated by pneumatic drives. Design of the machine was done with advice from engineers from *Festo*, a Belgian company specialized in industrial pneumatic systems. The shell of the rig was built by *Lichtert Industries*, an independent Belgian workshop, and the pneumatic compression system was assembled and embedded in the shell by *Festo*. There has been a few months' delay in the assemblage of the machine due to the novel aspect of the design (load in ice deformation rigs is indeed generally provided in direct way, i.e. with weights) and also due to security issues (mostly as a result of the high impact potential of the driving units).

The compression rig allows double independent, simultaneous strain experiments. It has been designed for performing uniaxial compression tests at constant load and under unconfined conditions. It is however pre-conditioned for use with rotational shear configuration in the future. Load is provided by pneumatic drives with integrated guide rods. Operating pressure ranges from about 0.5 to 10 bars.

Sample temperature is controlled by immersing the sample in a liquid bath connected to an external refrigerating system. Silicone oil of low viscosity and low volatility is used as a cooling liquid in the bath, which has the advantage to protect the ice from ablation, to be highly stable at experimental conditions and to help keeping constant the ice temperature, a drastic pre-requisite for good quality data.

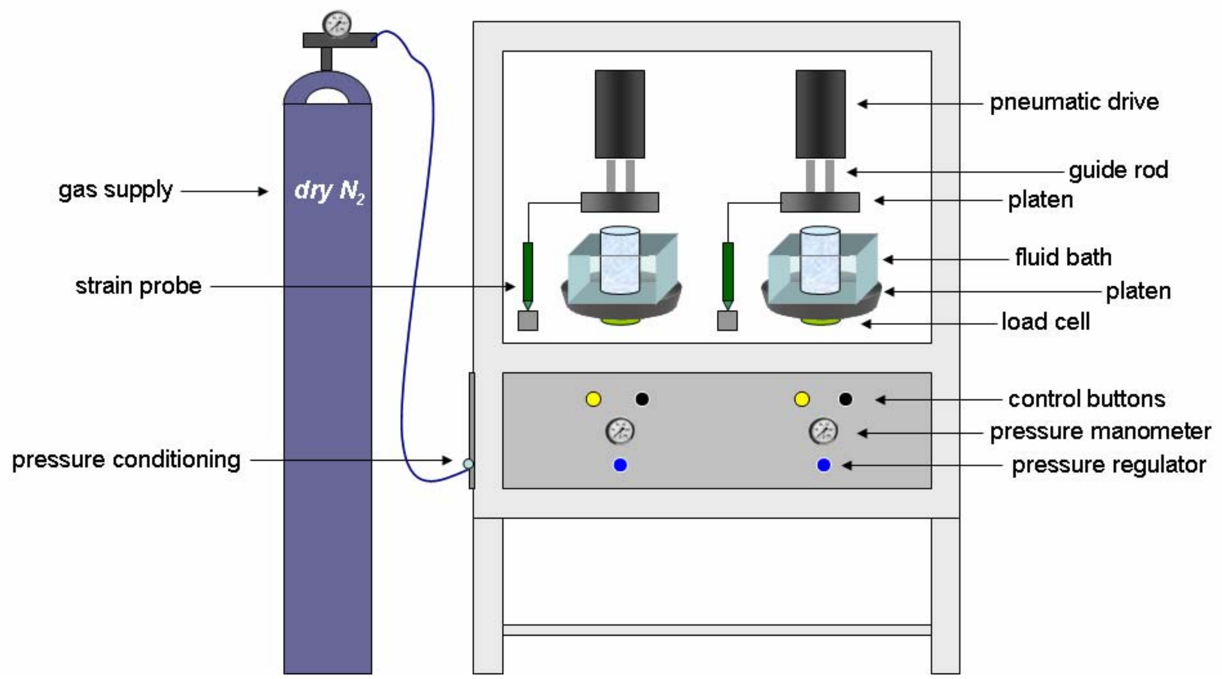


*Figure 6: Pneumatic compression rig from the ULB-Glaciology Lab*

Ice deformation is measured by displacement transducers with digital electronic output. Several digital probes can be used together in any combination by means of a network system connected to a PC via USB. Compression load is monitored by analogue load cells that are connected to the digital network through analogue input modules. Precision temperature probes are similarly connected to the network. Readings from the various sensors have been designed to be acquired directly into an Excel™ spreadsheet, which does not implies the use of specialized and expensive software. The digital network allows for monitoring of the experiments in process from remote places, which may be convenient knowing that strain experiments may last up to several weeks or months. This digital network represents another novel step in the ice deformation field. The whole system is depicted in Figure 6.

We also acquired an automated ice fabric analyzer, which makes it possible to study ice textures and fabrics from the whole polycrystal to the sub-millimetric scale in a fully automatic way. Such device, a G50, has been delivered at ULB by D. Russell-Head (DRH) in May 2007. There has been a serious delay in the delivery of the machine (about one year), but we finally managed to get one operational in the laboratory by insisting in DRH coming to work directly in our lab for nearly a month. This initiative also allowed 3 other European glaciology labs (LGGE, AWI, Aberystwyth) to obtain their machine that had been ordered for more than 3 years.

The G50 works with two types of experimental settings: one with a reduction tube, providing relatively fast scanning of the thin section at a large angle of view, and another (without the reduction tube) providing slower scanning but at finer resolution. This device is also suitable for studying other uniaxial materials like quartz or apatite. Tests on polycrystalline, debris-rich ice as well as on quartz and apatite thin sections are under way.

Deformation experiments on marine ice and rift ice-mélange samples: determination of ice rheological properties.

Three creep experiments have been performed until now on marine ice samples retrieved from the Nansen Ice Shelf (Terra Nova Bay, Antarctica - Khazendar, 2000; Khazendar et al., 2001; Tison et al., 2001). These experiments have been conducted in the cold rooms of the LGGE (Grenoble - France) in the framework of our collaboration with P. Duval and M. Montagnat., following the procedure described in the ad-hoc literature (e.g. Duval, 1976). Salinity (mean sample salinity was about 3.5E-2 g $l^{-1}$) and isotopic values were typical of marine ice and all samples were devoid of bubbles. Although we aimed at samples as homogeneous as possible, salinity and mean crystal size vary by a factor of 2. Fabric (c-axes) and crystal shapes were also selected to be as isotropic as possible, though this needs to be confirmed by ongoing detailed statistics on automated fabric analyses diagrams. The samples were strained at temperature close to -10.8 °C and at normal stresses of 5 bars, with extra loads ranging between 7 and 9 bars, for periods of two to three weeks. At any stress, we obtained minimum strain rates two to four times lower than reported for isotropic ice strained under similar conditions (Jacka, 1984). Extra load experiments allowed to assess Glen's flow exponent, which varied between 2.2 (lower stress) and 3.4 (higher stress) (Figures 7 and 8). The most important conclusion at this stage of the work is that, at conditions given here, marine ice from NIS deforms significantly slower than isotropic ice under pure shear deformation, emphasizing thereby the potential stabilizing effect of marine ice on continental flow.

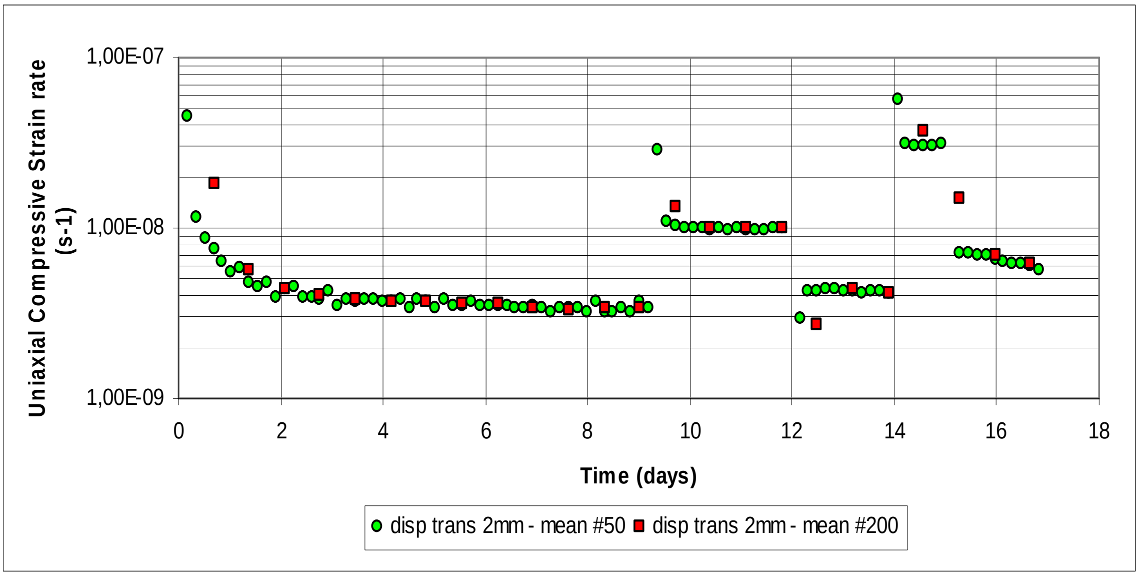


*Figure 7: Example of acquisition curve for the deformation of a marine ice sample (NIS2 72c). the direct displacement measurements have been converted into an uniaxial compressive strain rate and plotted against time. The driving stress has been increased twice in the course of the experiment. "disp trans 2mm" is abbreviation for displacement transducer with course of 2 millimeter. Green dots correspond to means calculated with integration values of 50. Red dots correspond to means calculated with integration values of 200.*

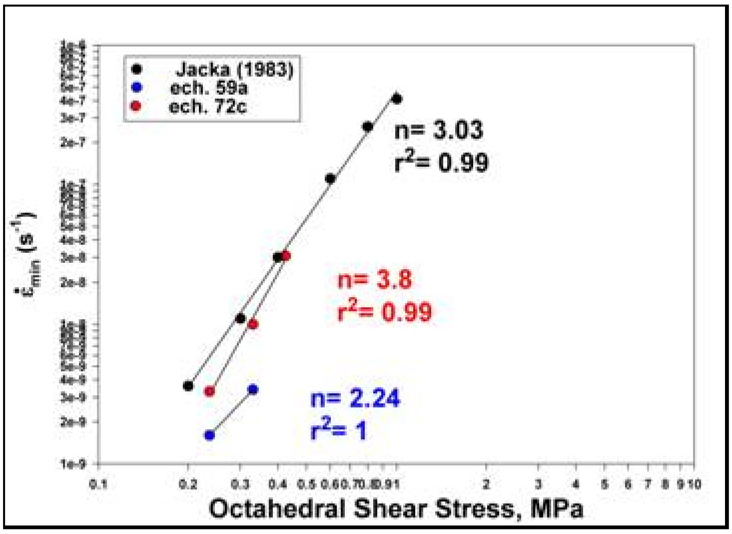


*Figure 8: Preliminary results from our marine ice deformation experiments compared to those of Jacka (1983) on artificial isotropic bubbly ice.*

More experiments, at varying experimental conditions and with varying physico-chemical parameters will help us to refine the values given above and to gain a better understanding of the sensitivity of marine ice to changing stress configurations.

Marine ice and ice-mélange properties in rifts

Thin sections have been prepared along sections of a 50 meters deep ice core retrieved from a rift in the Brunt Ice Shelf, Antarctica. A continuous section was obtained from the well coherent bottom 8 meters of the core. Thin section preparation in the upper 12 m of firn revealed itself to be rather tedious, due to its very low cohesion. About 20 of the ice thin sections have been analyzed using the Danish (Niels Bohr Institute) Automatic Fabric Analyzer in Nov. 2006, but data have not been processed yet. Conductivity data have been gathered from both the firn and the marine ice and have been converted into salinity. Salinity from the firn and ice zones are strongly contrasted, with values of about 0.035 per mil in the former and between 0.055 and 6.10 per mil in the latter (mean: 3.54 per mil). The transition from firn to ice is sharp on the salinity profile: values increase from 0.0026 per mil at the base of the firn (10.32m depth) to 3.70 per mil at 11.77m depth, in the upper ice zone (a value similar to multiyear sea ice salinity). This suggests that firn and ice found their origin in contrasting processes. Firn would result from compaction of local snow deposits formed at the bottom of the rift, on top of the pre-existing ice cover that resulted from the freezing of sea water. Bulk salinities are indeed probably too high for a marine ice origin, but that should be confirmed by the detailed textural analyses to come.

Preliminary co-isotopic ($\delta D$-$\delta^{18}O$) results have been obtained from various sampling sites at Minna Bluff (McMurdo Ice Shelf, Antarctica). Samples from firn, marine ice, debris-rich ice and melt ponds were pre-processed at Scott Base facility, on return from the field, and were analysed at University of Alberta, Canada, by Prof. M. Sharp. On a co-isotopic diagram, all samples (N = 38) align with a good fit (> 0.994) on a slope of 8.41, which is compatible with a "Local Meteoric Slope", although this could also potentially reflect some mixing processes (Figure 9). Indeed, end members on this slope are the firn samples (most depleted in heavy isotopes) and the marine ice samples (forming a cluster of positive values, a few per mil above SMOW composition). The composition of debris-rich ice samples, which were retrieved at the contact between the ice shelf and shore moraines, is trending toward that of marine ice, whereas the composition of the melt ponds is scattered between the firn and the debris-laden samples. This setting reflects some recycling between the various reservoirs on site. Rough mixing ratio calculations between firn and marine ice suggest a contribution of firn to ~ 70% of the melt ponds composition. A very first scenario would be that meltwater from hanging glaciers at Minna Bluff produces melt ponds in topographic depressions which can eventually reach the shore vicinity. Transgression phases of the ice shelf would then allow forming push-up moraines on shore, thereby mixing debris-rich ice from the base of the ice shelf, of partially marine ice origin, with local melt ponds. This scenario is supported by sea organism skeletons found within debris-rich ice, shore moraines and melt ponds. Deformation studies on these marine ice samples, with a potentially large range of debris content and salinities, should provide us an ideal tool to study the control of marine ice properties on its rheology.

Ice textures and fabrics from the Nansen Ice Shelf, initially described by Khazendar et al. (2001), have been resampled locally to study strong deformation features with the help of the Danish automatic analyzer. Considerations from the local ice shelf geometry setting, the c-axes orientation and the crystal elongation clearly indicate the development of longitudinal compressional folding under lateral constraints, an unusual situation for ice shelves where only vertical compression is usually considered. This, together with the development of strongly oriented fabrics, suggests a totally different stress/strain anisotropy that has never been considered before in ice shelf modelling.

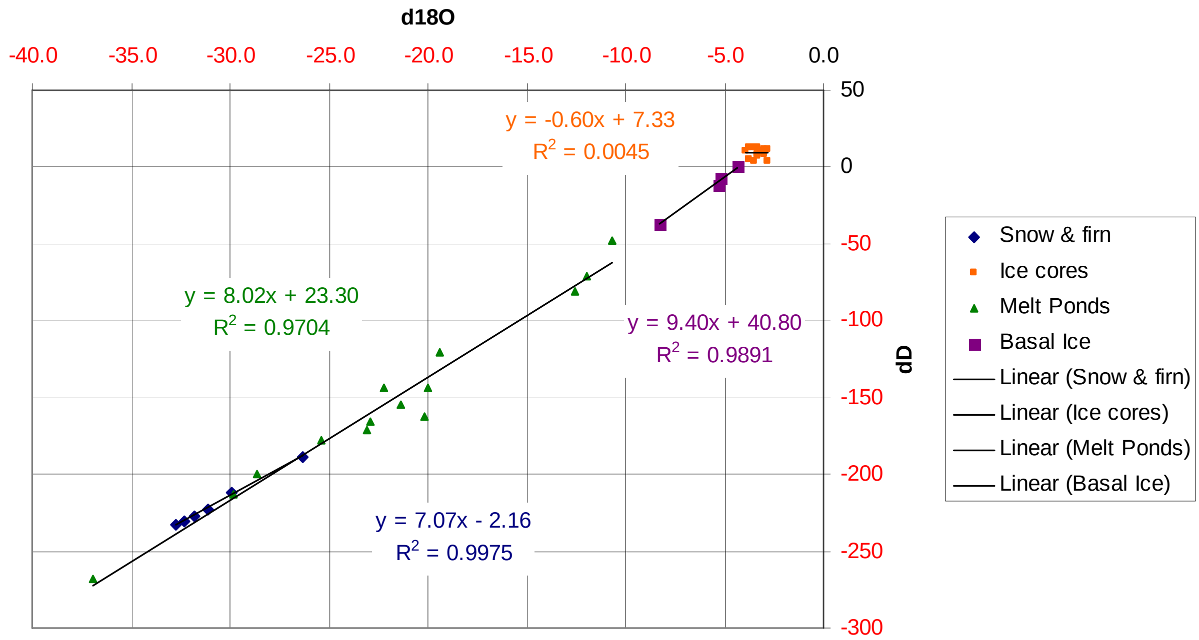


*Figure 9: Co-isotopic diagram for various sample sets collected at Minna Bluff (McMurdo Sound, Antarctica) in Jan.-Feb. 2007. 'Snow & firn' samples were collected at the surface of a glacier close to the shore; 'Melt Ponds' samples were collected between this glacier and the shore; 'Basal Ice' samples were collected within debris-rich ice layers overlying marine ice in the shore vicinity; 'Marine Ice' samples on a marine ice ridge ~ 1km offshore. This diagram has been produced in collaboration with Profs. M. Sharp and S. Fitzsimons)*

## Modelling of the transition zone between ice sheet and ice shelf

Basal processes and ice shelf viscosity play an important role in marine ice sheet dynamics. A thorough understanding of this role is hampered by the difficulty of collecting data about basal processes and ice shelf viscosity. This lack of knowledge hampers the credibility of current ice-flow modelling efforts. On the other hand, for many grounding zones, there is a good spatial coverage of surface velocity, surface elevation and ice thickness. With an inverse model, we can use this data to derive information about basal processes and ice shelf viscosity. This is done by tuning the necessary input values for basal processes and ice shelf viscosity so that, when used in an ice dynamics model, the modelled surface velocity equals the observed surface velocity.

For these inversions, we developed a robust 2D time-independent isothermal higher-order flowline model (De Smedt et al., 2007). The physical model is based on Pattyn (2002) and takes into account both vertical shear and longitudinal stress gradient aspects of ice flow, which makes it suited for both valley glaciers, ice sheets and ice streams. We use a Dirichlet boundary condition at the base in the form of a basal velocity. This way, we avoid the hypothesis of a sliding law which may not hold in practice. The numerical model is based on finite elements. We use triangular elements and linear basis functions to build a piecewise linear approximation of the velocity solution. This leads to a set of equations that is very similar to a finite difference discretisation on a staggered grid. This results in a significantly better stability in the iterative solution of the physical model. The numerical solution is accurate and robust for all sets of input data, as is confirmed by ample validation experiments. Part of these validation experiments follows the ISMIP-HOM benchmarks (Pattyn et al., 2008). From these benchmark experiments, it follows that our model produces results in line with other robust higher-order models. A second improvement is a simpler and more efficient relaxation scheme for the iterative solution of the physical model. An adapted preconditioned conjugate gradient solver assures a fast solution of the ice velocity for each iteration. These virtues of our numerical model allow for a fast and robust inversion of basal parameters.

Recent satellite data shows significant changes in the dynamics of a few fast-flowing coastal glaciers. For Pine Island Glacier (PIG), that has the largest discharge of all West-Antarctic ice streams, a grounding line retreat of about 1 m $a^{-1}$ has been observed. This retreat has been accompanied by a strong increase in surface velocity and a thinning that extends far inland. To learn more about the role of basal processes in these changes, we applied the newly developed 2D higher-order flowline model (De Smedt et al., 2007) along a 250-km-long flightline over PIG, extending from inland to the grounding line. Along this flightline, data is available on surface elevation, surface velocity and ice thickness at a resolution of 5 km (Figure 10a and 10b). Up to km 100 of the flightline, the observed surface velocity is approximately 100 m $a^{-1}$. From km 100 towards the grounding line, there is a linear increase in surface velocity up to 2000 m $a^{-1}$. Using the model in combination with a fixed point optimisation, we have derived the basal velocity pattern necessary to match the observed surface velocity. We performed this inversion for 2 values for the overall ice temperature: 0°C and -10°C. The actual deformation behaviour of PIG should lie between these two extremes. Since the basal ice is at pressure melting point along the whole of the profile and most of the ice deformation occurs at the bed, the 0°C scenario seems more realistic.

For both inversions, the derived basal velocity equals the observed surface velocity for most of the flightline (Figure 10b). This shows that the flow of PIG is strongly dominated by basal processes. Only in 2 small zones in the inland part (around km 50 and at km 100) and a larger zone near the grounding line (km 180-230), there is a significant relative contribution of ice deformation to the horizontal velocity. The deformation spot at km 100 is caused by a local peak in basal drag (Figure 10d). The two other deformation zones are due to a steeper surface slope. The mean deformation velocity in the grounding zone should lie between 100 km $a^{-1}$ (-10°C scenario) and 500 km $a^{-1}$ (0°C scenario). For both scenario's, the inverted basal drag shows a distinct drop from km 100 to km 105 (Figure 10d), where we have the onset of fast flow. This drop in basal drag coincides with a strong change from compression flow to extension flow and this extension flow continues throughout the rest of the flight-line (Figure 10d). The latter is due to the steady increase of sliding velocity towards the grounding line and is responsible for the inland propagation of grounding line effects. This is supported by another modelling study on PIG (Payne et al., 2004). Payne et al. showed that a 50% drop of basal friction at the grounding line has a thinning effect extending far inland. We also find that the low basal drag zones around km 70, 110 and 170 coincide with bed depressions. This suggests a lower friction in these zones, presumably due to basal hydrology.

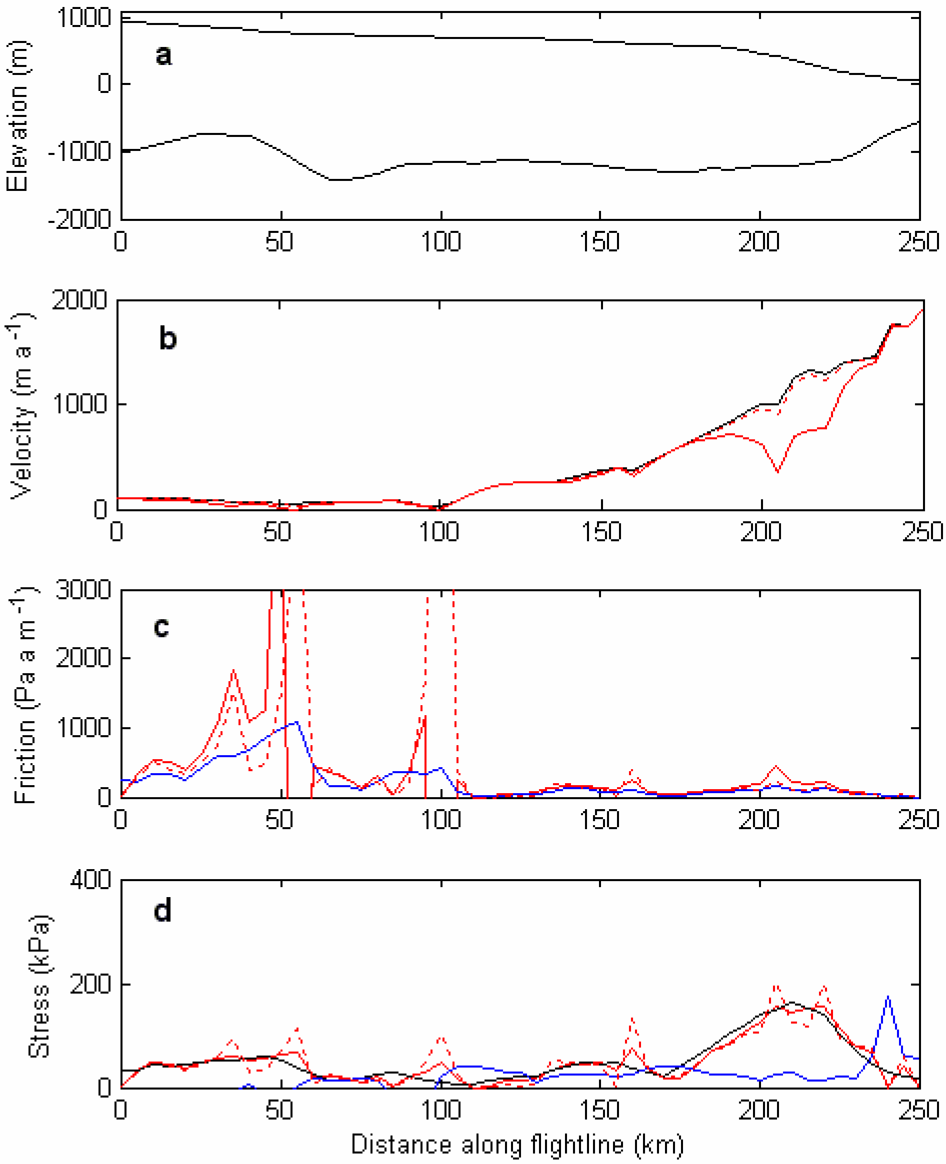


*Figure 10: Observations (Vieli and Payne, 2003) and model results along a flightline over Pine Island Glacier, West-Antarctica. Model results are shown for an overall ice temperature of 0°C (solid line) and -10°C (dotted line). (a) Observed geometry. (b) Observed surface velocity (black) and derived basal velocity (red). (c) Derived friction under the assumption of a linear sliding law for this study (red) and from Vieli and Payne (2003) (blue). (d) Observed driving stress (black) and derived basal drag (red) and mean longitudinal deviatoric stress (blue).*

Under the assumption of a linear sliding law $\tau_b = \beta^2 u_b$ with $\tau_b$ the basal drag, $\beta^2$ the basal friction and $u_b$ the basal velocity, we can derive basal friction from our model results. Just as basal drag, the resulting basal friction shows a pronounced drop around km 100 (Figure 10c). We also find a gradual rise of basal friction towards km 55, where we also have a strong drop. Throughout the rest of the profile, basal friction is very low, which is another indication of the high sensitivity of PIG to changes in the grounding zone. It is interesting to compare this friction pattern with the pattern derived by Vieli and Payne (2003), also under the assumption of a linear sliding law. In their study, Vieli and Payne use an ice-stream model, hereby assuming vertical shearing is negligible. Our results show that this assumption is justified over most of the profile, but probably too simplistic at the inland part and the grounding zone. As a result, for these two zones, Vieli and Payne underestimate the actual friction. The general friction pattern by Vieli and Payne , however, is very similar to our reconstruction. The flow of Pine Island Glacier (PIG), West-Antarctica, is strongly dominated by basal processes. As a result, basal drag and friction are very low over most of the grounding zone. This explains the observed sensitivity of the inland flow to changes in the grounding zone. The onset of fast flow coincides with a strong drop in basal drag. Low basal drag zones coincide with bed depressions. This suggests an enhanced activity of basal processes in bed depressions. These results are in line with studies by Vieli and Payne (2003) and Payne et al. (2004).

## Subglacial processes

### Stability of subglacial lakes

More than 145 subglacial lakes have been identified underneath the Antarctic ice sheet, mainly by analyzing airborne radio-echo sounding profiles. Subglacial lakes are characterized by a strong basal reflector, constant echo strength, corroborating a smooth surface, and a flat surface compared to the surrounding with a slope which is around ten times, and in the opposite direction of, the surface slope (Siegert et al. 2005). Hence, the ice column above a subglacial lake is in hydrostatic equilibrium. Most lakes lie under a thick ice cover of >3500m and are therefore situated close to ice divides. The volume of water in known Antarctic subglacial lakes is approximately 25% of the water world-wide in surface lakes (SALE Workshop Report, 2007). This is equivalent to a uniform sheet of water 1m thick if spread out underneath the whole Antarctic ice sheet.

Subglacial lakes are supposed to be relatively closed and stable environments with long residence times and slow circulations (Kapitsa et al., 1996). Therefore, such huge amount of subglacial meltwater would pose no threat to the stability of the ice sheet if it were not moving around through a hydrological network. This view has recently been challenged through the observation of rapid subglacial lake drainage events (Gray et al., 2005; Wingham et al, 2006; Fricker et al, 2007). The observations by Wingham et al (2006) suggest the rapid transfer of subglacial lake water and periodically flushing of subglacial lakes connected with other lakes that consequently fill through a hydrological network. For the draining lake (Adventure Trench Lake), they observe a surface lowering of 4m, corresponding to a water discharge of 1.8km$^3$ over a period of 16 months at a rate of 50m$^3$s$^{-1}$. These observations stem from rapid changes at the surface from which the drainage events are inferred. At present, direct observation of such subglacial events is lacking, and therefore knowledge on the mechanisms that trigger them as well. One way, however, to investigate possible mechanisms that lead to rapid subglacial drainage is through ice-sheet modelling in which ice flow across and interactions with a subglacial lake are taken into account. We therefore developed a full Stokes numerical ice-sheet model which is coupled to the subglacial lake environment to investigate the stability of subglacial lakes and their drainage through a sensitivity analysis. A full description of the model can be found in Pattyn (2003) and Pattyn (2008). Developing a full Stokes model was necessary, as bridging effects in the force balance equations (vertical resistive stresses) cannot be neglected, especially with respect to rapid variations in lake volume. In the model, basal hydrology is represented in terms of the subglacial water flux. The water flow algorithm is invoked to check whether subglacial water captured in the lake cavity is transported outside this cavity or not and therefore serves to detect whether the hydrostatic seal is intact or broken. Full basal hydraulics, such as tunnel formation and closure and associated water discharge (Nye, 1976), are not taken into account. Besides, as small floods creating small surface depressions are considered here (not the cauldron dimensions associated with Jökulhlaups), only flotation criterion for the ice over the lake as well as grounding line migration due to lake volume variations, is considered.

Starting from the initial conditions, the model ran forward until a steady state was reached. The general characteristics of the resulting ice sheet geometry are typical for those of a slippery spot, as is a subglacial lake, i.e. a flattened surface of the ice/air interface across the lake and the tilted lake ceiling in the opposite direction of the surface slope, due to hydrostatic equilibrium. The tilt of this surface in the direction of the ice flow will determine the stability of the lake, since the hydraulic gradient is dominated by surface slopes and therefore the flatter this air/ice surface the easier water is kept inside the lake cavity. Changes in the surface flatness will therefore be crucial in subglacial lake stability. Therefore a sensitivity analysis was carried out from which it follows that lake stability is prevalent for large lakes, shallow ice, small ice surface slope and fast ice flow. However, the most determining parameter here is general surface slope: the flatter the surrounding area the more stable lakes are, which corroborates the presence of Antarctic subglacial lakes close to ice divides.

Our hypothesis is that (partial) lake drainage is a common feature of the Antarctic subglacial lakes and that only small perturbations are necessary to lead to a sudden outburst. Therefore, three perturbation experiments were carried out, i.e. (i) an experiment where the surface of the ice sheet is slightly perturbed, (ii) one in which subglacial water is gradually added to the lake system, and a third one where 20% of the lake water volume is removed at once. The first two experiments are regarded as common gradual changes in the glacial environment that on the time scales of decades are hardly noticeable, hence forming part of the natural variability of the system. Once the drainage condition is fulfilled – i.e. the hydrostatic seal is broken – the lake is drained at a rate of $50 m^3 s^{-1}$ for a period of 16 months (see above). This essentially means that enough energy is released to keep the subglacial tunnel open for a while and a siphon-effect can take place. It is not clear whether this effect is only temporary or ends when the lake is completely emptied as suggested by Wingham et al. (2006). However, the drainage event of Lake Engelhardt suggests that water is still present in the lake after the event, as the surface aspect (flatness) has not changed over this period of time. Furthermore, Evatt et al. (2006) suggest that the drainage event of the Adventure Trench Lake is due to channelized flow over soft sediments at low effective pressure, rather than the emptying of a very shallow lake. Not all experiments exhibit lake drainage. Certain parameter settings, such as small ice thickness, render lakes very stable. When drainage occurs, episodic events take place, even though the initial geometric conditions are not met. These events occur at a higher frequency at the beginning, i.e. when more water is present in the lake, than later on and occur with frequencies of less than a decade. Most experiments show stable conditions after 200 years of integration, either because the drainage conditions are not fulfilled anymore, either because the cavity is devoid of water or a complete drainage has taken place. The episodic drainage results in variations in the surface elevation across the lake of rapid lowering, followed by a gradual increase until the initial level is more or less reached. This surface rise is clearly nonlinear, and the rate of uplift decreases with time. The frequency decreases with time as well as less water is present in the system, hence hampering drainage (Figure 11). Interesting to no note, however, is that the surface rise is not caused by an influx of subglacial water entering from upstream. On the contrary, less water is subsequently present in the whole subglacial system. Surface increase is due to an increased ice flux, filling up the surface depression created by the sudden lake drainage.

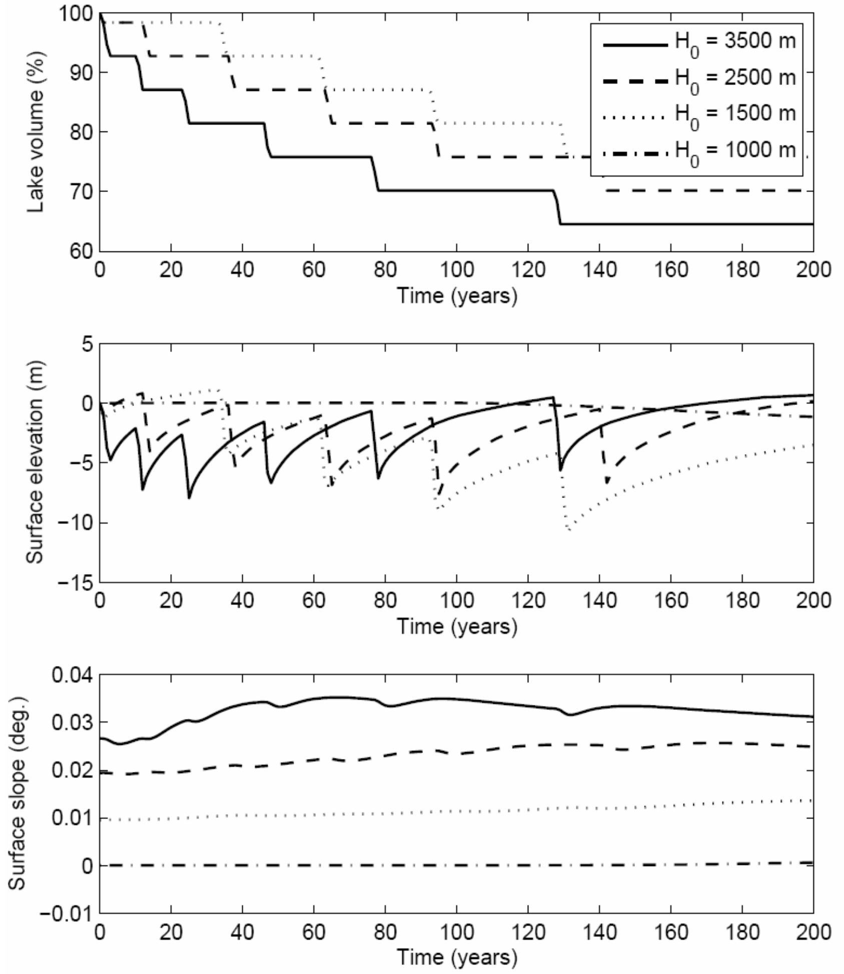


*Figure 11: Time series of the perturbation experiment according to a change in surface elevation. The three panels show the time evolution of lake water volume (upper), ice surface elevation on top of the lake (middle) and ice surface slope across the lake (bottom) for the ice thickness sensitivity. No lake drainage is observed for small ice thicknesses ($H_0$ = 1000m).*

Most subglacial lakes are lying close to ice divides, hence in areas of low surface slopes and high hydraulic fluid potential. Therefore, a fundamental question arises (Priscu et al., 2007): does the ice sheet control the location of subglacial lakes or does the lithospheric character necessary for lake formation, such as geothermal heat flux and the presence of basal cavities and sub-ice aquifers, constrain the evolution of these catchments? According to the model simulations, geometric effects, such as ice thickness, mean surface slope, ice viscosity (hence ice temperature), and lake size play an important role in the stability of subglacial lakes. They favour the existence of subglacial lakes near ice divides in areas of low viscosity and low surface slopes. Lake filling might therefore happen when the ice sheet is in full expansion during sustained cold periods and when surface slopes are very low, such as a glacial period. During interglacial periods, the smaller and warmer ice sheet characterized by slightly higher surface slopes in the interior facilitates lake drainage, a point that will be investigated more closer in the future and that might as well explain why underneath the Greenland ice sheet subglacial lakes are not observed.

The above described experiments have shown that the developed model is capable of exhibiting the major characteristics of drainage due to changes in ice sheet and/or subglacial lake geometry. Results demonstrate that only small changes are necessary to provoke episodic lake drainage. These events only lead to a partial drainage of the lake, after which the ice sheet adapts to the changed conditions and ice flow `fills' the surface depression after drainage. Therefore, lake drainage on a decadal time scale can be regarded as a common feature of the subglacial hydrological system and may influence to a large extent the behaviour of large ice sheets through such jökulhlaups (Evatt et al., 2006). This has potential implications for ice stream dynamics, especially in areas that are underlain by sedimentary material. Unless this water escapes to the margin via well defined channels, it will act to increase pore-water pressures and so reduce the strength of sediments, which will increase ice velocity (Siegert et al., 2007). Finally, the experiments show that big lakes under thick ice in slow moving regions drain easy, hence a feature that should be common for subglacial Lake Vostok.

Basal properties of the Antarctic ice sheet and subglacial lake presence

Physical properties at the base of the Antarctic Ice Sheet are hardly known. The basal thermal regime is a main player in the ice sheet dynamics: if the ice is frozen to its underlying bedrock, the only velocity component will be ice deformation. But when the pressure melting point is reached at the ice base, the ice mass will start to slide over its base, reaching velocities of up to more than 1000 meters per year. Knowing whether the basal ice is melting or not is also important for the correct interpretation of paleoclimatic signals of ice cores. And it controls whether erosion and landscape modification will take place.

We used a three-dimensional thermomechanical ice sheet model (Huybrechts, 2002) to calculate the temperature distribution within the Antarctic ice sheet in an effort to help identifying regions which reach pressure melting. Two boundary conditions are needed in the temperature calculation: observed surface temperature and the basal temperature gradient. The latter is determined by basal drag, the basal sliding velocity and the geothermal heat flux from the earth.

The geothermal heat flux is the largest unknown in the calculation. Some authors use a constant value, e.g. 54.6 mW/m² (Huybrechts, 1992), others prefer to divide the Antarctic continent into different geological provinces with a high geothermal heat flux in West-Antarctica, which is composed of younger oceanic crust. East-Antarctica has a lower heat flux because it consists of an old continental craton (e.g. Llubes et al., 2006). Two other approaches to calculate the geothermal heat flux for the Antarctic continent are based on extrapolation of a seismic model (Shapiro and Ritzwoller, 2004) or on the use of satellite magnetic data in combination with a model of crustal thickness to obtain a good estimate of the flux values (Fox Maule et al., 2005).

In the experiments, we studied how the basal thermal conditions of the ice sheet are affected by the different heat flux values. Experiments with a constant value (54.6 mW/m² and 70 mW/m²) are compared with runs where the data of Shapiro and Ritzwoller (2004) or Fox Maule et al. (2005) are used.

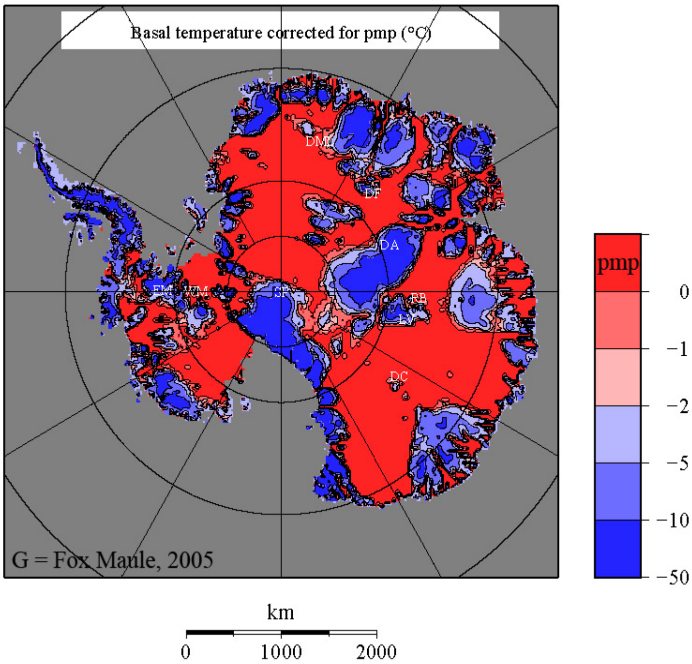


Figure 12: Modelled basal temperatures, using the data from Fox Maule et al. (2005) for geothermal heat flux values.

Using the data of Fox Maule et al. (2005), pressure melting is reached at most of the coastal areas and in the inland regions (Figure 12). Dome C, Ridge B and Lake Vostok as well as the Siple Coast are situated in warm basal regions. Using the basal thermal regime calculated by the model, the basal melting rate can be calculated and compared with the presence of subglacial lakes (Siegert et al., 2005). In this experiment, 73 percent (106 of the 145 identified lakes) are located in a region at pressure melting point. Lakes near Dome A and Dome F are situated in cold regions whereas Lake Vostok and the lakes near Ridge B and Dome C are in pressure melting regions. The same experiments were done with values of 54.6 and 70 mW/m² and the estimate made by Shapiro and Ritzwoller (2004). The percentages of lakes that are situated in pressure melting point regions are 65.5, 96.6 and 49 percent respectively (Figure 13). Using 70 mW/m², the pressure melting point is reached almost everywhere. However, this is quite unrealistic since the East Antarctic Ice Sheet rests on an old continental craton with most likely a lower geothermal heat flux, but the experiment shows how sensitive the basal thermal regime is to the use of different values.

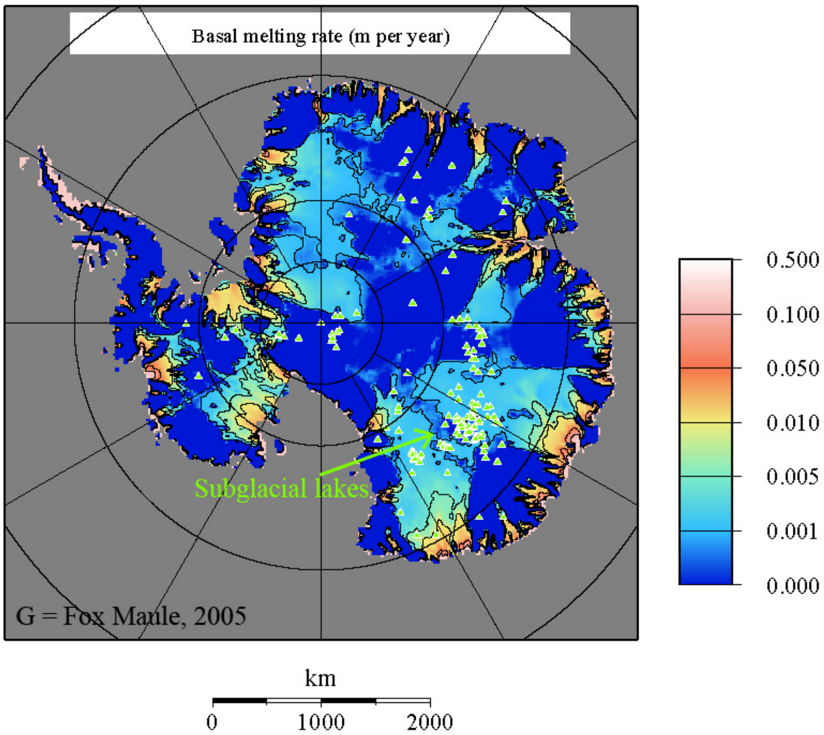


*Figure 13: Location of subglacial lakes (Siegert et al., 2005) and modelled basal melting rate, using the data from Fox Maule et al. (2005) for geothermal heat flux values.*

## Basal ice properties and processes

### Multiparametric analyses of basal ice from EPICA Dome C Deep ice

The EPICA Dome C ice core is 3259.85m long and has been stopped about 15m above the bedrock. The stable isotopes record down to 3200 m has recently allowed reconstruction of the climate back to Marine Isotopic Stage 20.2 (about 800kyr ago, Jouzel et al., 2007). The bottom 60 meter (deep ice) however shows a $\delta D_{ice}$ amplitude three times smaller than the record above, in discordance with the oceanic record of MIS 21, 22 and beyond. It also clearly shows some visible dirt inclusions up to three millimetres in size. There is therefore some doubt on its paleoclimatic validity. During the last two years, we have gathered information about this EPICA Deep ice, measuring a suite of variables, generally at low resolution (generally 6m spacing, as opposed to the bag-55 cm resolution for the $\delta D_{ice}$), apart from the chemistry of the lower 6 meters (richer in visible inclusions) that has been measured at a 11 cm resolution. From the low resolution record (see Figure 14), several variables seem to indicate that the ice is of "glacial" origin and has kept its paleoclimatic signature (mean $\delta D$ value, slope of the co-isotopic relationship, total gas content, mean $CH_4$ and mean $CO_2$ signal, mild glacial dust content…). Some records are however still an enigma: discrepancy between the $\delta D_{ice}$ and the oceanic record, noticeable reduction of the $\delta D_{ice}$ amplitude for an uncommon length of the record, larger dispersion of the co-isotopic values along the regression slope, one low total gas value, extremely flat $\delta^{18}O_{atm}$ curve, at a value more typical of warm spells, record maximum point values of MSA as compared to the whole rest of the core. Occurrence of small scale (typically cm) disturbances of the signals has recently been confirmed by the high resolution chemistry profiles in the bottom 6 meters (Figure 15). It is important to note that these data were collected in samples that did not include visible inclusions, although we are clearly in a layer where those are frequent. All chemicals but potassium (MSA, Cl, $NO_3$, $SO_4$, Na, Mg, Ca) show similar features: a background decreasing trend from typically high "glacial" to typically low 'interglacial" values, if one refers to some of the already published data for the rest of the core, at a resolution of 2.2 meters (Wolff et al., 2006). All chemicals also show a few (5) events were they synchronously display record concentration levels for the whole ice core. This peculiar chemistry at the decimetric scale (which is also the scale for the crystal size in this deep ice) suggests concentration mechanisms and secondary "in situ" chemistry within the intercrystalline sub-liquid layer (pre-melting). Very high resolution (2 cm) to micro-scale (micro-diffraction X et XANES around solid inclusion) measurements performed on a few samples in collaboration with M. de Angelis at LGGE confirm this hypothesis. Large amounts of small calcite crystals were found in one of the particle inclusions, associated with a cloud of fine bubbles that could reflect $CO_2$ degassing on precipitation.

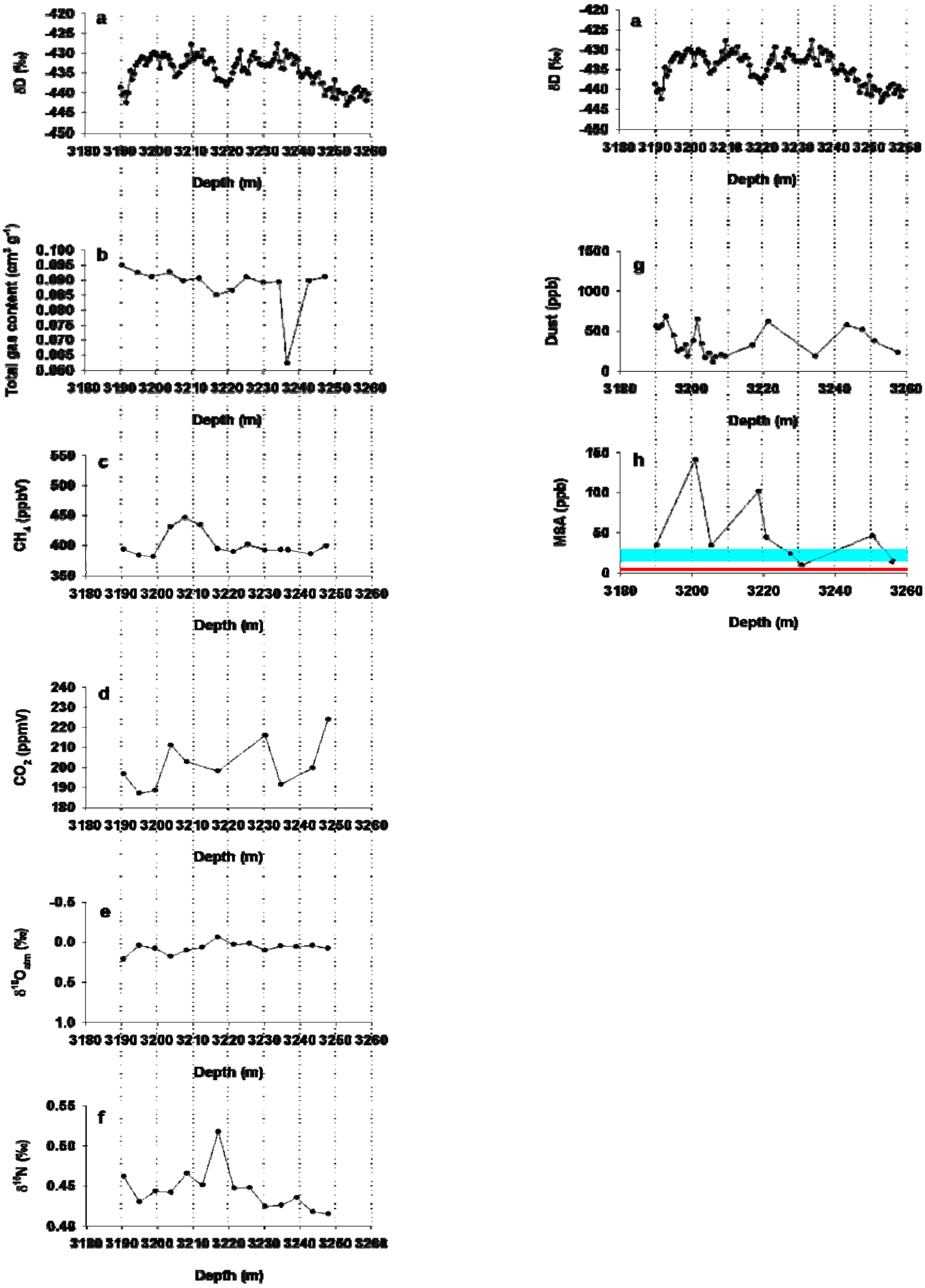


*Figure 14: Low resolution multi-parametric analyses of the bottommost 60 meter of the EPICA Dome C ice core (Deep Ice). See text for details. (Data collected in collaboration with EPICA members: J. Jouzel, D. Raynaud, J. Chapellaz, J.M. Barnola, A. Landais, J.-R. Petit, M. de Angelis, E. Wolff, G. Littot and others).*

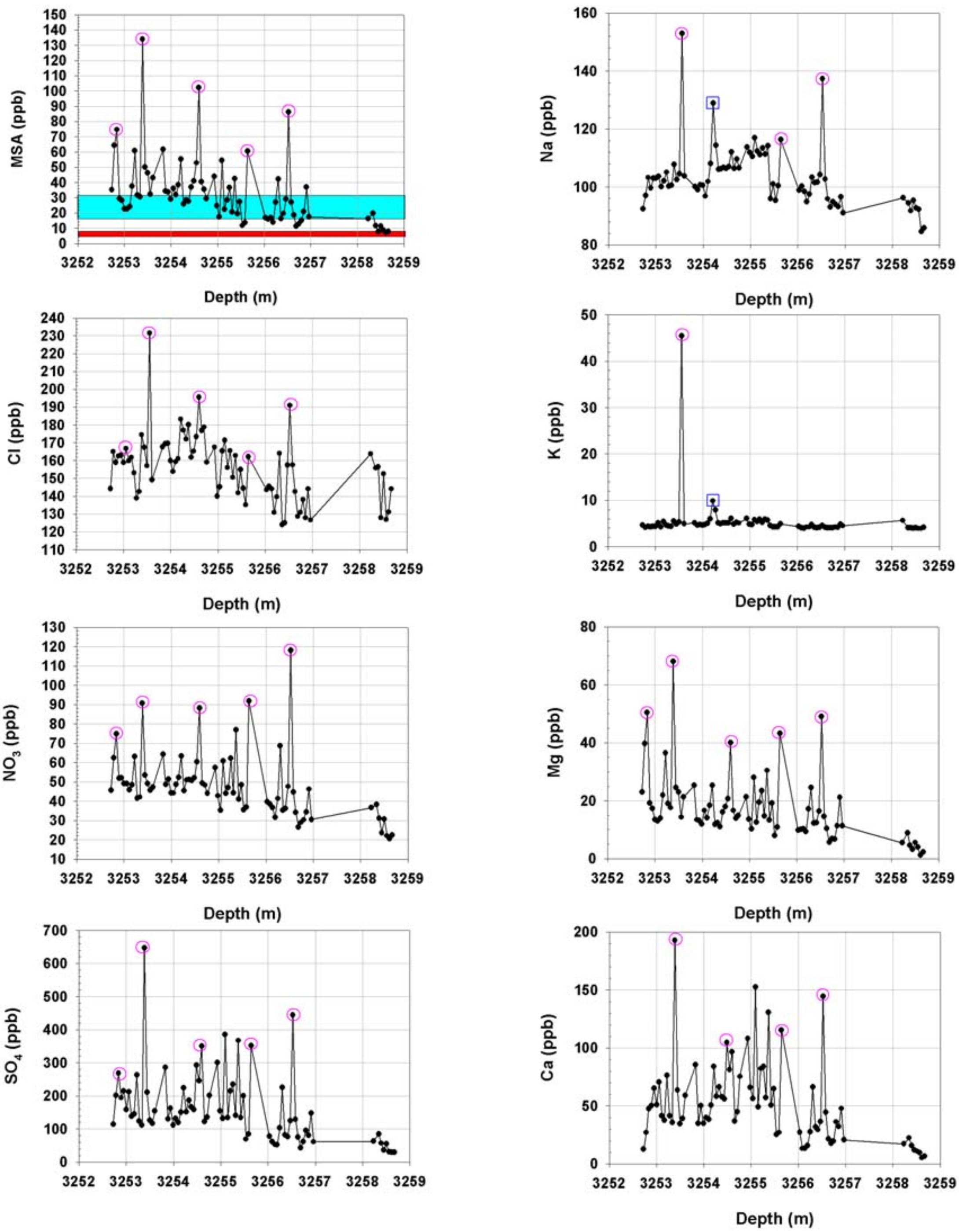


*Figure 15: High resolution (11 cm) chemical analyses of the bottommost 6 meters of the EPICA Dome C ice core. See text for details.*

Another unusual feature of this EPICA deep ice is the good correlation between MSA and sulphates (Figure 16), which doesn't show up higher up the core, due to post-depositional processes affecting MSA, especially in warm periods. The much less obvious relationship between MSA and Na indicates that adherence of MSA to primary marine aerosols cannot explain the good MSA/Sulphate relationship and suggest alternative surface conditions during the formation of this deep ice, or unknown in-situ chemical processes.

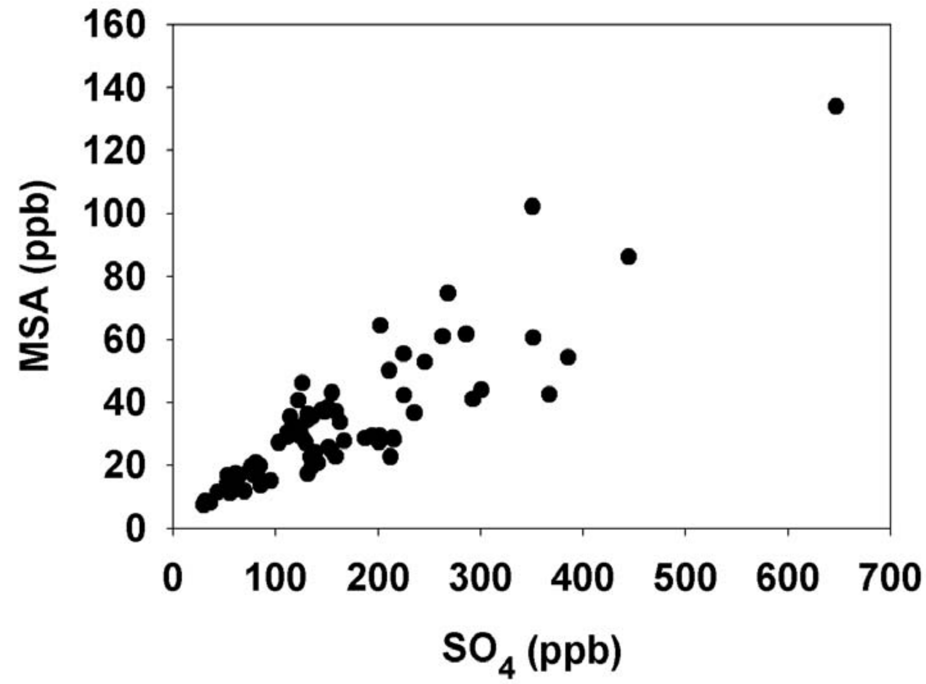


*Figure 16: MSA/$SO_4$ relationship in the bottommost 6 meters of the EPICA Dome C ice core.*

Our present state of knowledge on the EPICA Deep ice leads us to suggest the occurrence of active mechanisms linked to flow disturbances close to the ice bedrock interface. These could explain the limited variability of some of our indicators (that should otherwise show much larger amplitudes) such as the $\delta D_{ice}$ or the $\delta O^{18}_{atm}$, the presence of aggregates of solid particles (sometimes sandy) in the ice, modifications of the small scale chemical signature related to enhanced recrystallization processes due to variations of the stress field close to bedrock irregularities. Complementary measurements during the second phase of this research project should allow us to better tackle these potential processes (e.g. stretching of basal layering resulting from basal melting vs. mixing of different layers of various origin around bedrock obstacles).

Multiparametric analyses of basal ice from EPICA DML refrozen basal water

We have been solicited by our colleagues at the Alfred Wegener Institute (S. Kipfstuhl) to contribute to the study of the basal refrozen water that has been collected in the EDML drill hole, after unplanned infiltration of the subglacial water in the bottommost part of the ice sheet at that location. At the end of this deep drilling (2774.15 m), liquid water of subglacial origin swept and refroze into the bottom of the drill hole, in contact with the drilling liquid (Foran + D40) and the densifier (HCFC-141b).

The refrozen bottom water clearly differs from ice of meteoric origin and it is rather similar to chlathrate ice (Figure 17). It is "light" and appears very white and opaque. X-ray and Raman spectroscopy analyses (University of Goettingen) show that it is made of 48%$\pm$8% of hexagonal ice, 48%$\pm$8% of hydrate type II and 4% n-pentane. The latter comes from the drilling fluid and is too big to produce hydrate type II. Atmospheric gases ($N_2$, $O_2$, Ar, $CO_2$, $CH_4$) can only produce hydrate type I (Davidson, 1973, Miller, 1973). This leaves us with the densifier as a potential candidate for the chlathrates, and, indeed, its molecular structure ($CH_3CCl_2F$) is very close to a known chlathrate II compound: $CH_3CClF_2$. Further, chromatographic elution of the gas extracted from the ice (Figure 18) clearly shows a peak associated with elution of pure densifier vapour. Clearly, the ice that formed at the time of subglacial water infiltration, did incorporate portions of the densifier and of the drilling fluid (emulsion).

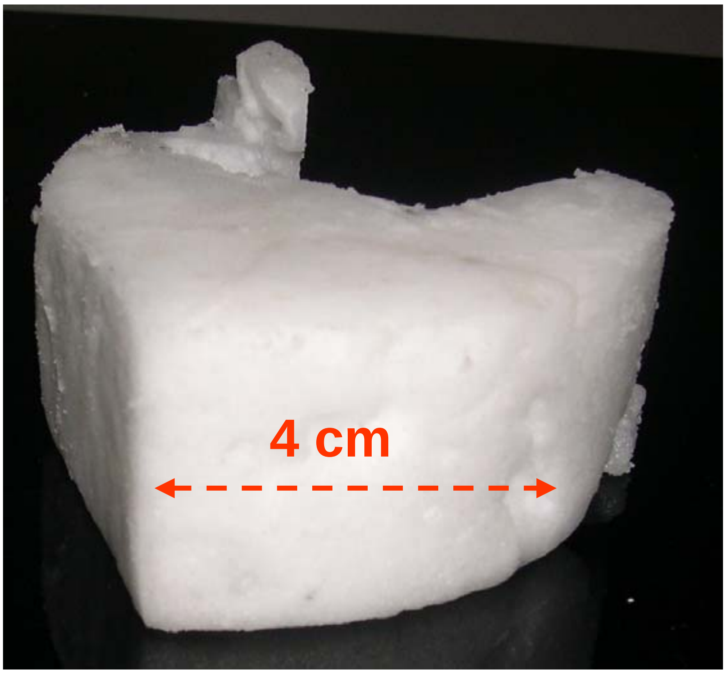


*Figure 17: EPICA Dronning Maud Land basal refrozen water sample. Note the similarity to Chlathrate ice.*

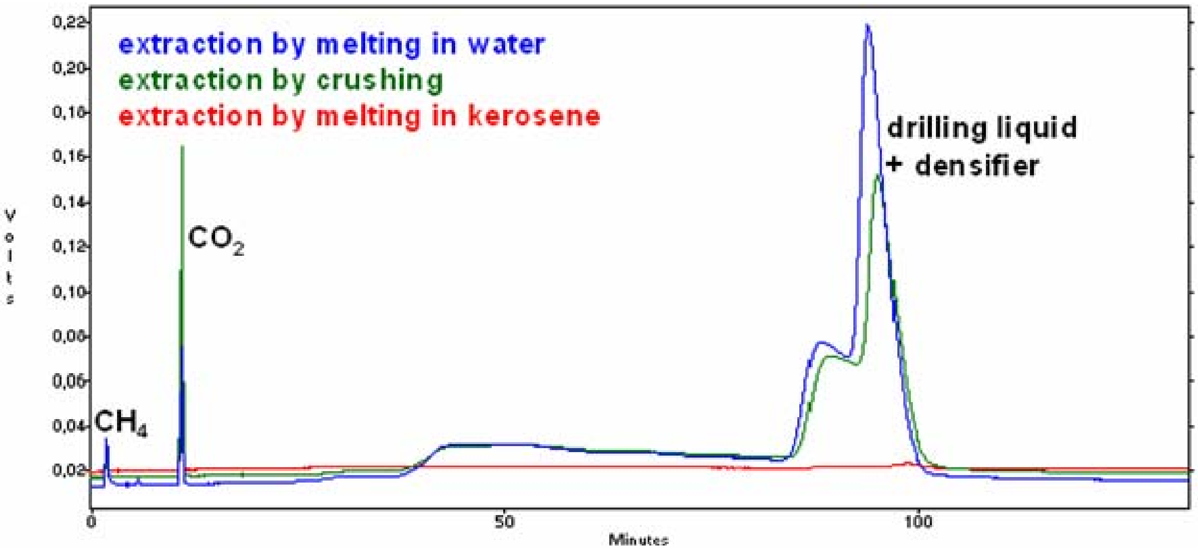


*Figure 18: Combined chromatograph for the elution of three samples of gas extracted from the EDML refrozen water using three different techniques. The drilling liquid and densifier peak is clearly seen for the samples extracted by melting under water and crushing. It is however absent from the extraction method using melting in kerosene. This is explained by the fact that the density of the densifier and drilling liquid is higher than that of kerosene. These compounds therefore remain at the bottom of the melting.*

We are however mainly interested in the “atmospheric” component of the ice, to see if we can learn something from it, in terms of subglacial water sources and properties. To collect this fraction of the gas content we tried various techniques that proved to be more or less successful. Our usual technique to extract the total amount of gases from the ice, using a Toepler pump system with ice melted and refrozen in an evacuated vessel did not work. It simply continuously evaporated the densifier which has a very high partial vapour pressure (40x water value!), resulting in the collection of ridiculously large amounts of gas. We therefore opted for the old method used by Langway (1958), where the ice block is melted at room temperature in a liquid (water or kerosene), under an inverted funnel connected to a burette filled with the liquid. As the ice melts, the gases collect at the top of the graduated burette, where the total gas content is read at the end of the experiment. Further corrections need to be applied to subtract the vapour partial pressure contribution of water, kerosene, drilling liquid and densifier, in order to obtain the actual contribution of the atmospheric gases to the total.

Gas composition ($O_2$, $N_2$, $CO_2$, $CH_4$) was obtained using the traditional ice crushing extraction technique at low temperature and under vacuum, with Gas Chromatography measurements on the same sample, using an Interscience Trace GC. The table here below summarizes the total gas content and gas composition measurements we have made on several samples of refrozen water (about 20), the stratigraphy of which however remaining uncertain.

| | Total Gas (l kg$^{-1}$) | | $O_2$ (%) | | $N_2$ (%) | | $CO_2$ (ppmV) | | $CH_4$ (ppmV) | | $O_2/N_2$ | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | **Min** | **Max** | **Min** | **Max** | **Min** | **Max** | **Min** | **Max** | **Min** | **Max** | **Min** | **Max** |
| **samples** | 0.38 | 4.07 | 21.26 | 30.65 | 69.04 | 78.38 | 1100 | 3600 | 2.7 | 11 | 0.27 | 0.46 |
| **Air bubbles in ice** | 0.09 | | 20.95 | | 78.08 | | 200-370 | | 0.6 | | 0.27 | |
| **Solubility in water (0°C under 2700 m ice)** | 6.26 | | 37.8 | | 62.19 | | 22400 | | 117.54 | | 0.59 | |

A first obvious feature is the large range of total gas content in the ice samples analyzed (380 ml to 4 litres per kg of ice). Another striking feature is that this is, in all cases, much higher than the usual mean value of about 90 ml per kg of ice normally encountered in deep ice cores. It should be noted also that the maximum value is however still below the expected saturation value for water at about 245 bars, i.e. below 2700 meters of ice (around 6 litres per kg of ice). Gas compositions also indicate that the ice was globally undersaturated as compared to water solubility, at least for carbon dioxide, methane and oxygen. $O_2$ to $N_2$ ratios range between atmospheric value (characteristic of the bubbles in the ice) and saturation solubility value in fresh water. These characteristics give us some hints on the processes occurring at the ice-bedrock interface. If, as one might hypothesize, the only input of gases to the melted pure crystals of ice at the ice-bedrock interface, are those initially occluded in the bubbles at the surface (closed system), and if no fractionation occurs on melting, then the dissolved gases in the subglacial waters should display an $O_2/N_2$ ratio close to atmospheric, until saturation occurs for the less soluble gases. However, if the hypothesis is true that the only source of gases are the bubbles in the ice, it is impossible to understand total gas content values of up to 4 litres per kg of ice, when the mean value in meteoric ice is only 0.09. We have therefore to think of potential processes that could be responsible for the gas enrichment in the subglacial water. Until now we can think of three possible mechanisms depicted in the next series of figures:

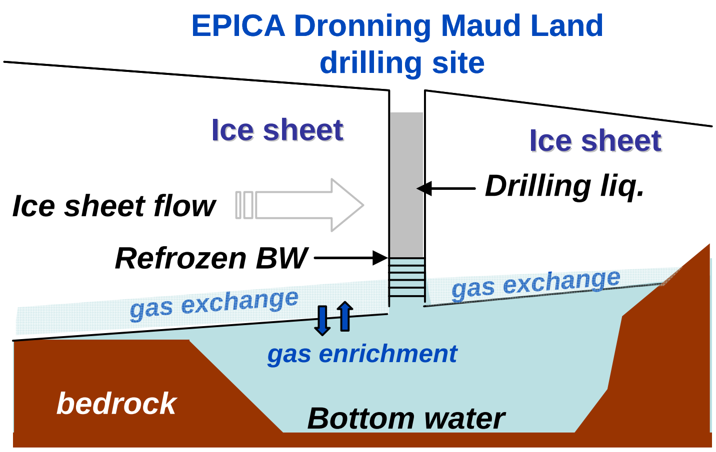


*Figure 19a: An ice sheet moves over a water body of constant volume, with small scale melting-refreezing processes affecting the intercrystalline sub-liquid layer during recrystallization close to the pressure melting point within the bottom*

*layers. This process has been identified under ice shelves at the meteoric ice – marine ice interface (J.-L. Tison, pers. com.), and proven to be physically possible at the metric scale by Rempel et al. (2001). This should then result in the gas impoverishment of the bottom meteoric ice. Preliminary results indeed show very low conductivity and total gas content in the basal few meters of ice at EDML (S. Kipfstuhl, pers. com.).*

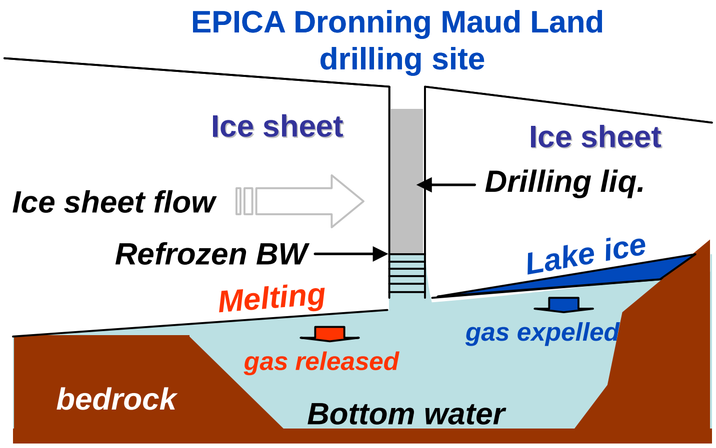


*Figure 19b: An ice sheet moves over a water body of constant volume with large scale melting-refreezing: the Vostok case. The gas enrichment results here from both the melting of meteoric ice and refreezing of lake ice (in a general cell of thermohaline circulation), which occurs with a near total expulsion of gases in the remaining liquid. However, no large body of water appears to exist in the vicinity of EDML.*

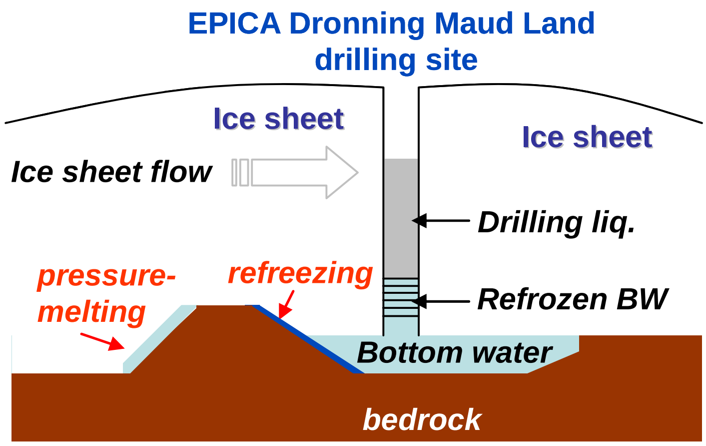


*Figure 19c: An ice sheet moves around bedrock obstacles, with bottom ice at the pressure-melting point, and with a potential water film or layer at the interface. Melting on stoss side of bedrock bumps delivers air bubbles to the interface water, and refreezing on the lee side rejects and concentrates gases in the solution.*

Note that all three processes could involve selective diffusion of gases ahead of the freezing interface, which would favour the faster diffusing oxygen and increase the $O_2/N_2$ ratio in the subglacial water. It should be emphasized, however, that part of the interpretation of this data set is depending on unknown processes of gas exchanges and nucleation occurring at the time of solidification of the infiltrating subglacial water within the drill hole. A large range of consolidation rates is plausible, which probably - at least partly - explains the wide range of total gas content values observed in the different samples.

Crystallographic investigations at Taylor Glacier

In an attempt to formalize basal ice processes at subfreezing temperature and to assess the potential role played by debris during recrystallization, crystallographic and structural analyses have been conducted in cold basal ice from the snout of Taylor Glacier, Antarctica (Samyn et al., 2007). Fabric analysis shows a clear increase in grain boundary and nucleation kinetics at the interface between debris-rich and debris-free ice layers of the basal ice sequence, which denotes strain localization at these interfaces during basal ice genesis. Analogies with bottom ice from deep polar ice sheets where temperature is currently higher than at the studied site are highlighted and recrystallization scenarios are proposed for the development of the fabrics observed. It is shown that by controlling the repartition of stress and strain energy within basal ice, the rheology of debris-rich ice layers plays a decisive role in recrystallization dynamics at structural interfaces, which in turn controls the dynamics of the whole ice body. It is also shown that the same recrystallization regimes may occur in cold glaciers and temperate ice sheets, provided that strain accumulation has been high enough in the former. This challenges the common belief that migration fabrics observed in bottom ice from deep ice-sheets are exclusive to stagnant, annealed ice.

## CONCLUSIONS AND OUTLOOK

ASPI aims to understand the interactions between the ice sheet and the subglacial environment and the processes that control the Antarctic ice sheet, and to determine the stability of the ice sheet in a changing climate and the impact of climatic variations on the coastal ice sheet. A key factor is the existence of transition zones within the ice sheet, such as grounding lines, i.e. the interface between the ice sheet and an ice shelf, between an ice sheet and a subglacial lake, as well as between an ice shelf and its pinning points. Especially grounding lines in marine ice sheets, such as the WAIS, are extremely vulnerable, as small perturbations at the grounding line, such as basal melting or variations in sea level, might lead to an important grounding line retreat, as has been observed recently in the Amundsen Sea sector. Therefore, a major task by ASPI is to investigate the factors that influence the physics and mechanics of the grounding lines, such as rheology and the impact of marine ice formation in rifts at the grounding line (which could either stabilize or destabilize the ice flow), as well as how to represent the processes that are responsible for grounding line migration in ice sheet models.

During the first two years of this project a number of these tasks have been addressed. On behalf of modelling and model development, we analyzed the response of a marine ice sheet to different perturbations near the grounding line using a numerical ice sheet model that takes into account longitudinal stress coupling and grounding line migration. The model is based on an existing flowline model, but extended with a novel subgrid determination of the grounding line position and migration as a function of the size of the transition zone between the ice sheet and the ice shelf. A wide transition zone is typical for an ice stream, a small transition zone typical in many areas around the East Antarctic ice sheet where the ice sheet rather suddenly changes into an ice shelf. Model results show that stress transmission or longitudinal coupling across the grounding line plays a decisive role. The grounding line migration is a function of the length scale over which the basal conditions change from frozen to the bed to floating, the "transition zone". We demonstrated that thinning of the ice shelf due to bottom melting has a negligible effect on the grounded ice mass. Only perturbations at the grounding line or reduction in buttressing of the ice shelf substantially thins the grounded ice sheet. Marine ice sheets with large transition zones - such as ice streams - seem highly sensitive to such perturbations, compared to ice sheets with small transition zones, such as an abrupt ice sheet/ice shelf junction.

A major ASPI task is to include the mechanisms of stress transfer and grounding line migration into a thermomechanically coupled 3D SIA model of the Antarctic ice sheet, in order to refine the response of the ice sheet to climate changes. The stress transmission algorithm has successfully been included in this model and ways of implementing grounding line migration on subgrid level are currently under investigation. Besides the outer boundary condition, special care is taken to ameliorate the subglacial hydrology and basal sliding algorithms. Basal ice temperature sensitivity was investigated for different geothermal heat data sets and validated against current knowledge on subglacial lake distribution. This way it was possible to determine what dataset is more suitable for future model experiments and what basal conditions are more likely to reign underneath the Antarctic ice sheet.

Deformation experiments are meant to shed more light on how marine ice formation at the grounding line influences the flow properties of the ice sheet system in the transition zone. Therefore, a novel deformation device, driven by pneumatic units and equipped with a digital network for data acquisition, has been installed. Once tested and calibrated, this facility will enable us to better understand marine ice rheology. Preliminary experiments indicate a contrast in the behaviour of marine versus isotropic ice. At equivalent deviatoric stress conditions, minimal strain rates for marine ice are indeed lower and appear sooner than for artificial isotropic ice, underlining the potential for marine ice to stabilize continental meteoric ice flow. The second phase of the project should decipher the range of variability of these rheological properties, depending on fabric pre-conditioning, solid and dissolved impurity content, bubble density, … These rheological properties should certainly strongly be contrasted with those from other types of ice-mélange such as the one sampled in the rift of Brunt Ice Shelf.

The importance of such result is in conjunction with results from inverse modelling of the transition zone. The latter technique allows for a determination of the ice viscosity properties and/or the basal characteristics in the transition zone, based on observed ice sheet configuration (ice thickness, surface topography) and observed surface velocities. Inverse modelling is therefore capable of determining the size of the transition zone, which is an important factor for predictive modelling. For instance, the application to Pine Island Glacier (WAIS) clearly allows us to determine to position of the onset of the ice stream as well as to delineate areas where stress coupling is essential. 2D plane inverse model techniques based on genetic algorithms are currently looked in to.

Recently, the latest version of an Automatic Ice Fabric Analyzer (IFA) was acquired. Preliminary tests show significant resolution improvements as compared to previous versions and promises a robust characterization of marine and basal ice physical properties, combined with numerical processing with the Matlab Texture Toolbox. Thin sections have been prepared in firn and ice from Brunt Ice Shelf. Some of these have been analyzed with the IFA, and data processing is scheduled for late 2007. Conductivity analysis of firn and ice suggest contrasting origins. The firn zone would result from classical snow compaction, whereas the ice zone would result from regelation processes involved in sea ice build-up. Ice fabric, Ion composition and co-isotopic ($\delta D$-$\delta^{18}O$) measurements will help us to reconstruct the origin and evolution of both zones as well as to detect the potential contribution of marine ice to the bottom accretion in the rift. These measurements are scheduled for late 2007.

Subglacial lakes are yet another type of transition zones that are currently gaining attention. Recent observations demonstrate that subglacial lakes may drain and add significant amounts of basal water to the subglacial hydrological system. Hence, they have the potential of destabilizing ice sheets through sudden lake outbursts, of which evidence exist along the coast of the East Antarctic ice sheet. In the first phase of ASPI we investigated the effect of subglacial lake drainage on the stability of the ice sheet and especially how sensitive the subglacial lake system is towards drainage and flooding. It is shown that only slight changes in environmental conditions (surface slope, basal melting) easily lead to episodic subglacial lake drainage. Furthermore, these events seem to be very common underneath the Antarctic ice sheet. Whether such drainage applies to subglacial Lake Vostok is currently under investigation.

Turning to the ice core analyses, low resolution multi-parametric measurements in the bottom 60 meters of the EPICA Dome C ice core show pro's and con's as whether it should be trusted as a paleoclimatic sequence. Although several variables are coherent with a glacial stage signature, others indicate small-scale modifications potentially linked with ice deformation processes close to the bedrock. Further developments based on the comparison of high resolution data from this section and from typical glacial and interglacial sequences further up in the core should allow us to decipher the processes at work, and to assess the balance between internal chemical re-working and active interactions with the substrate. We will also aim at estimating how much of the paleoclimatic signal is lost.

Refrozen water at the bottom of EDML results from synchronous freezing of water, drilling liquid and densifier, resulting in its typical "chlathrate ice" appearance. We did recover the signature of the atmospheric gases entrapped in this ice and shed some light on the potential processes occurring at the base of large ice sheets. Subglacial water shows a high level of dissolved gases that can only be explained by melting-refreezing and associated concentration processes at the ice bedrock interface, be it sub-ice/lake thermohaline circulation or pressure-melting around bedrock obstacles. No obvious sign of biological activity is detected in the $CO_2$ and $CH_4$ concentrations that are well below those detected in basal ice layers from the Greenland Ice Sheet, for example (GRIP, Dye-3…).

Physical properties at the base of the Antarctic Ice Sheet are hardly known. The basal thermal regime and the ice fabrics are main players in the ice sheet dynamics, but debris characteristics also hold a determining role in these dynamics. It is therefore important to decipher basal ice processes at cold and sub-zero temperatures to assess their influence on the dynamics of the whole ice body. Basal ice from Taylor Glacier, an Antarctic polythermal glacier, was studied with a multi-parametric approach to appraise the role of debris on the evolution of ice fabrics. Fabric analysis reveals a clear influence of debris arrangement on grain boundary and nucleation kinetics at rheological interfaces of the basal ice sequence. It is also shown that strain is localized at these interfaces during basal ice genesis. This has important implications for the knowledge of recrystallization dynamics at the bottom of deep polar ice sheets, with which strong analogies are found with the studied site.

Transition zones in the Antarctic ice sheets, whether it be grounding lines, subglacial lakes or the subglacial interface are key elements in the dynamic behaviour of the Antarctic ice sheet and its stability. Even at the end of the first phase of the ASPI project, we start to better understand the subglacial processes and interactions that occur at these interfaces. They seem even more important as a controlling factor as previously thought of. While this first phase was essentially focused on understanding and deciphering the processes involved in the stability of transition zones, the second phase will – according to plan – focus on the specific analysis of these processes, the implantation of these processes within existing models, application to known cases and impact assessment.

**REFERENCES**

De Smedt, B., P. de Groen and F. Pattyn. (2007): A robust 2D higher-order ice-flow model for inverse applications. Geophysical Research Abstracts 9: EGU2007-A-0083.

Davidson, D. W. (1973): Clathrate hydrates. In 'Water: a comprehensive treatise. Vol. 5', London.

Duval P. (1976): Lois du fluage transitoire de la glace polycristalline pour divers états de contrainte. Ann. Geophys., 32, (4), 335-350.

Evatt, G., A. Fowler, C. Clark, N. Hulton (2006): Subglacial floods beneath ice sheets, Phil. Trans. R. Soc. A., 364, 1769-1794.

Fox Maule, C., M. E. Purucker, N. Olsen, K. Mosegaard (2005): Heat flux anomalies in Antarctica revealed by satellite magnetic data, Science 309, 464–467.

Fricker, H., T. Scambos, R. Bindschadler, L. Padman (2007): An active subglacial water system in West Antarctica mapped from space, Science 315 (1544).

Gray, L., I. Joughin, S. Tulaczyk, V. Spikes, R. Bindschadler, K. Jezek (2005): Evidence for subglacial water transport in the West Antarctic ice sheet through three-dimensional satellite radar interferometry. Geophysical Research Letters 32 (L035001).

Herterich, K. (1987): On the flow within the transition zone between ice sheet and ice shelf, in: C. J. Van der Veen and J. Oerlemans (eds.): Dynamics of the West Antarctic Ice Sheet, D. Reidel, Dordrecht, 185-202.

Hindmarsh, R. (1993): Qualitative dynamics of marine ice sheets, in Ice in the Climate System, W. Peltier (Ed.), NATO ASI Series I (12), 67-99, Berlin, Springer-Verlag.

Hindmarsh, R. C. A. (1996): Stability of ice rises and uncoupled marine ice sheets, Ann. Glaciol., 23, 105-115.

Hindmarsh, R., C., A., Le Meur, E. (2001): Dynamical processes involved in the retreat of marine ice sheets, J. Glaciol., 47, 271-282.

Hindmarsh, R. (2004): A numerical comparison of approximations to the Stokes equations used in ice sheet and glacier modeling, J. Geophys. Res. 109 (F01012), doi:10.1029/2003JF000065.

Huybrechts, P. (1992): The Antarctic ice sheet and environmental change: a three-dimensional modelling study, Berichte zur Polarforschung, 99.

Huybrechts, P., A. Payne and the EISMINT Intercomparison Group (1996): The EISMINT benchmarks for testing ice-sheet models. Ann. Glaciol. 23, 1-12.

Huybrechts, P., Abe Ouchi, A., Marsiat, I., Pattyn, F., Payne, T., Ritz, C., Rommelaere, V. (1998): Report of the Third EISMINT Workshop on Model Intercomparison, European Science Foundation, Strasbourg, 120 p.

Huybrechts, P., De Wolde, J. (1999): The dynamic response of the Greenland and Antarctic ice sheet to multiple-century climatic warming, J. Clim., 12, 2169-2188.

Huybrechts, P. (2002): Sea-level changes at the LGM from ice-dynamic reconstructions of the Greenland and Antarctic ice sheets during the glacial cycles. Quaternary Science Reviews, 21, 1-3, 203-231.

Huybrechts, P., O. Rybak, F. Pattyn, U. Ruth, and D. Steinhage (2007): Ice thinning, upstream advection, and non-climatic biases for the upper 89% of the EDML ice core from a nested model of the Antarctic ice sheet, Climate of the Past, 3, 693-727.

Jacka, T.H. (1984): The time and strain required for development minimum strain rates in ice. Cold Regions Science and Technology, 8, 261-268.

Jouzel, J. and 31 others (2007): Orbital and Millennial Antarctic Climate Variability over the Past 800,000 Years. Science, 1141038 DOI: 10.1126.

Kapitsa, A, J. Ridley, G. Robin, M. Siegert and I. Zotikov (1996): A large deep freshwater lake beneath the ice of central East Antarctica, Nature 381, 684-686.

Khazendar A. (2000): Marine ice formation in rifts of Antarctic ice shelves – A combined laboratory study and modeling approach, Univ. Libre de Bruxelles, Ph.D. Thesis, 153 pp., unpublished.

Khazendar A., Tison J.-L., Stenni B., Dini M., Bondesan A. (2001): Significant marine-ice accumulation in the ablation zone beneath an Antarctic ice shelf. J. Glac., 47 (158), 359-367.

Langway, C. C. J. (1958): Bubbles pressures in Greenland glacier ice. In 'Inter. Union Geodesy Geophys. Symposium of Chamonix', 336-349.

Llubes, F. R. M., C. Lanseau (2006): Relations between basal condition, subglacial hydrological networks and geothermal flux in Antarctica, Earth and Plan. Sci. Lett. 241, 655–662.

Mayer, C., Huybrechts, P. (1999): Ice-dynamic conditions across the grounding zone, Ekströmisen, East Antarctica, J. Glaciol., 45, 384-398.

Miller, S. L. (1973): The Clathrate Hydrates - Their Nature and Occurrence. Physics and Chemistry of Ice. Royal Society of Canada, 42-50.

Nye, J. (1976): Water flow in glaciers: jökulhlaups, tunnels and veins, J. Glaciol 17(76), 181-207.

Pattyn, F. (2002): Transient glacier response with a higher-order numerical ice-flow model. Journal of GLaciology, 48(162), 467-477.

Pattyn, F. (2003): A new three-dimensional higher-order thermomechanical ice-sheet model: basic sensitivity, ice-stream development and ice flow across subglacial lakes. Journal of Geophysical Research (Solid Earth), 108 (B8), 2382, doi:10.1029/2002JB002329.

Pattyn, F., A. Huyghe, S. De Brabander, and B. De Smedt (2006): The role of transition zones in marine ice sheet dynamics. J. Geoph. Res. (Earth Surface). 111 (F2): No. F02004, doi:10.1029/2005JF000394.

Pattyn, F., Huyghe, A., De Brabander, S., De Smedt, B. (2006): Role of transition zones in marine ice sheet dynamics, J. Geophys. Res., 111, F02004, doi 10.1029/2005JF000394.

Pattyn, F., L. Perichon, A. Aschwanden, B. Breuer, B. de Smedt, O. Gagliardini, G. H. Gudmundsson, R. Hindmarsh, A. Hubbard, J. V. Johnson, T. Kleiner, Y. Konovalov, C. Martin, A. J. Payne, D. Pollard, S. Price, M. Rueckamp, F. Saito, O. Soucek, S. Sugiyama, and T. Zwinger (2008) Benchmark experiments for higher-order and full-Stokes ice sheet models (ISMIP-HOM). The Cryosphere 2(3), 95-108.

Pattyn, F. (2008) Investigating the stability of subglacial lakes with a full Stokes ice sheet model. J. Glaciol. 54(185): 353-361.

Payne, A.J., A. Vieli, A.P.. Shepherd, D.J. Wingham and E. Rignot. (2004): Recent dramatic thinning of largest West Antarctic ice stream triggered by oceans. Geophysical Research Letters 31, L23401, doi: 10.1029/2004GL021284.

Pollard, D., and R.M. DeConto (2007): Grounding line behavior in a heuristically coupled ice sheet-shelf model, Geophysical Research Abstracts, EGU2007-A-03103.

Priscu, J., S. Tulaczyk, M. Studinger, M. Kennicutt, B. Christner, C. Forman (2007): Antarctic subglacial water: origin, evolution and microbial ecology. In Vincent W. and J. Laybourn-Parry (eds.) Polar Limnology, Oxford University Press, UK, in press.

Rempel, A. W., Waddington, E. D., Wettlaufer, J. S. and Worster, M. G. (2001): Possible displacement of the climate signal in ancient ice by premelting and anomalous diffusion. Nature 411, 568–571.

Ritz, C., Rommelaere, V., Dumas, C. (2001): Modeling the evolution of Antarctic ice sheet over the last 420,000 years: Implications for altitude changes in the Vostok region, J. Geophys. Res., 106, 31,943-31,964.

Samyn, D., Svensson, A and Fitzsimons, S. (2007): Discontinuous recrystallization in cold basal ice from an Antarctic glacier: dynamic implications. J. Geophys. Res. (Earth Surface), under review.

SALE Workshop Report (2007): Subglacial Antarctic Lake Environments (SALE) in the International Polar Year 2007-08, Advanced Science and Technology Planning Workshop, 24-26 April 2006, Grenoble, France.

Schoof, C. (2007a): Marine ice-sheet dynamics. Part1. The case of rapid sliding, J. Fluid. Mech., 573, 27-55.

Schoof, C. (2007b): Ice sheet grounding line dynamics: steady states, stability and hysteresis, J. Geophys. Res., in press.

Shapiro, N., M. Ritzwoller (2004): Inferring surface heat flux distributuions guided by a local seismic model: particular application to Antarctica, Earth and Planet.Sci. Lett. 223, 213–234.

Shepherd, A., D. Wingham, and E. Rignot (2004): Warm ocean is eroding West Antarctic Ice Sheet, Geophys. Res. Letters, 31 (L23402), doi:10.1029/2004GL021106.

Shepherd, A., And D. Wingham (2007): Recent sea-level contributions of the Antarctic and Greenland ice sheets, Science 315, 1529-1532.

Siegert, M., S. Carter, I. Tobacco, S. Popov, D. Blankenship (2005): A revised inventory of Antarctic subglacial lakes, Ant. Sci. 17 (3) 453–460.

Siegert, M;, A. Le Brock and A. Payne (2007): Hydrological connections between Antarctic subglacial lakes: The flow of Water beneath the East Antarctic ice sheet and implications of sedimentary processes, Glacial Sedimentary Processes and Products. In press.

Tison J.-L., Khazendar A., Roulin E., 2001. A two-phase approach to the simulation of the combined isotope/salinity signal of marine ice, Journal of Geophysical Research, 106 (C12), 31387-31401.

Van der Veen, C., J. (1999): Fundamentals of Glacier Dynamics, A. A. Balkema, Rotterdam/ Brookfield, 459 p.

Vaughan, D. and R. Arthern (2007): Why is it hard to predict the future of ice sheets?, Science 315, 1503-1504.

Vieli, A. and A.J. Payne (2003): Application of control methods for modelling the flow of Pine Island Glacier, West Antarctica. Annals of Glaciology, 36, 197-204.

Vieli, A. and A.J. Payne (2005): Assessing the ability of numerical ice sheet models to simulate grounding line migration, J. Geophys. Res., 110 (F01003), doi:10.1029/2004JF000202.

Weertman (1974): Stability of a junction of an ice sheet and ice shelf, J. Glaciol., 13, 3-11.

Wingham, D., M. Siegert, A. Shepherd, A. Muir (2006): Rapid discharge connects Antarctic subglacial lakes, Nature 440, 1033-1036.

Wolff E.H. et al. (2006) Southern Ocean sea-ice extent, productivity and iron flux over the past eight glacial cycles. Nature 440, 491-496 doi:10.1038/nature04614