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A Need For More Realistic Ice-Sheet Models

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A NEED FOR MORE REALISTIC ICE-SHEET MODELS

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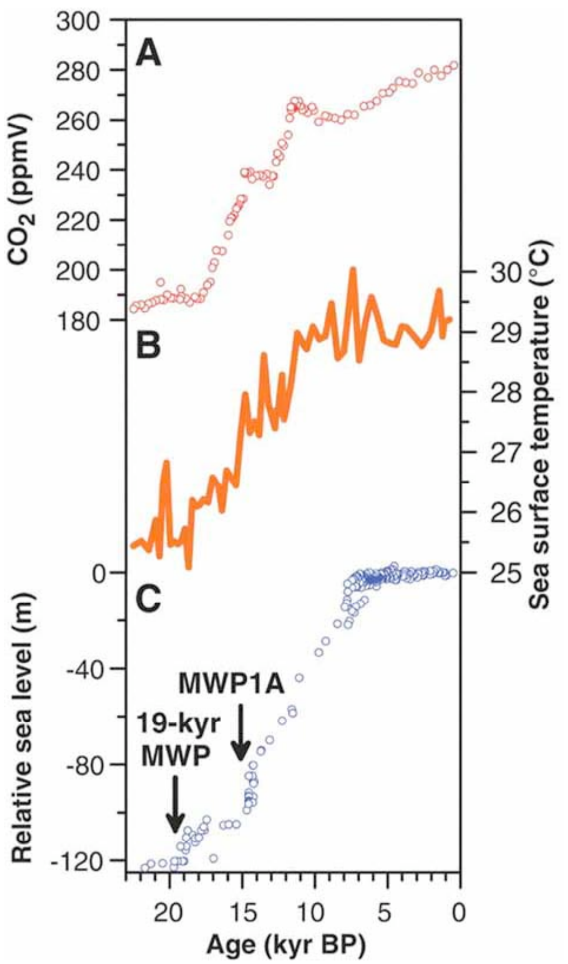
1. BACKGROUND

Over the past two decades or so, evidence for active ice sheets – both present-day and in the past – has mounted and the traditional view of ice masses responding sluggishly to external forcings has been replaced by the understanding that large ice sheets can undergo rapid changes. For example, analysis of deep-sea sediment cores has revealed major oscillations in the volume of the Laurentide Ice Sheet, believed to be associated with surges of the Hudson Bay lobe (e.g. Broecker, 1994). On a smaller scale, at the height of the Last Glacial Maximum (~17,000 years B.P.) the margin in the Great Lakes region of North America advanced and retreated over several hundreds of km within a short time span (Karrow and others, 2000). During meltwater pulse 1A, *ca*. 14,000 years B.P., sea level rose approximately 20 m over a period of some 400 years, at an average rate of 5 cm/yr (Kienast and others, 2003), compared to the rate of sea-level rise of 1-2 mm/yr during the 20th century (Church and Gregory, 2001). Such rapid changes cannot be explained by “conventional” models that have been used so far to simulate ice-sheet evolution. The majority of these models is based on the shallow-ice approximation in which the driving force is balanced by friction at the glacier bed. Allowance has been made for basal sliding where the bed is lubricated, or for deformation of a soft sediment layer. However, these models may

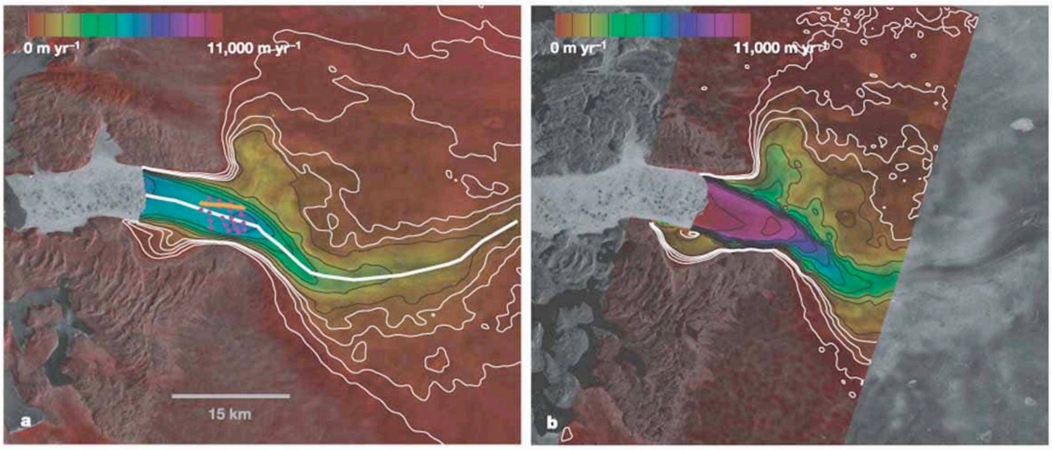
Box 1. Time series of key variables encompassing the last interval of significant global warming (last deglaciation). (**A**) Atmospheric CO_2 from Antarctic ice cores. (**B**) Sea surface temperature in the western equatorial Pacific based on Mg/Ca measured in planktonic foraminifera . (**C**) Relative sea level as derived from several sites far removed from the influence of former ice-sheet loading . MWP, meltwater pulse.

From: Alley and others (2005).



not fully capture the controlling dynamics of ice streams and outlet glaciers, and do not allow for rapid along-flow propagation of force perturbations including those that may result from removal of peripheral ice shelves.

Recent studies have focused attention on dynamic behavior of the Greenland Ice Sheet and on rapid changes occurring that cannot be explained by existing ice-sheet models. Peripheral outlet glaciers are thinning rapidly at rates that cannot be explained by increased ablation, pointing to ice-dynamical effects (Thomas and PARCA, 2001; Van der Veen, 2001; Thomas and others, 2004a; Rignot and others, 2004a; Rignot and Kanagaratnam, 2006). At the same time, the speed of many outlet glaciers has increased significantly, with Jakobshavn Isbræ speeding from 6.8 km/yr in 1995 to 12 km/yr in 2003 (Joughin and others, 2004a; Rignot and Kanagaratnam, 2006), likely as a result of enhanced basal lubrication (Zwally and others, 2002; Rignot and others, 2004a) or possibly a dynamical response to weakening and subsequent disintegration of buttressing tidal ice tongues (Thomas, 2004; Joughin and others, 2004a). For some of the outlet glaciers, the speed up was short lived and velocities have returned to earlier values (Howat and others, 2007). No consensus has emerged about processes controlling onset and stoppage of fast glacier flow and improving our understanding of these issues is imperative for developing better numerical ice-sheet models and constraining future evolution of the Greenland and Antarctic ice sheets.

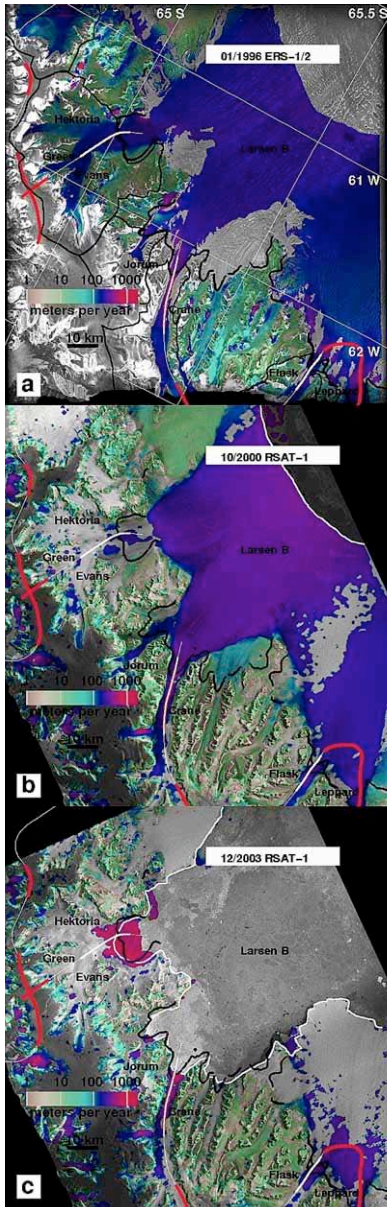


Box 2. Ice-flow speed as colour over SAR amplitude imagery of Jakobshavn Isbræ in February 1992 (left) and October 2000 (right).

From: Joughin and others (2004a).

In West Antarctica, ice streams draining into the Ross and Ronne-Filchner ice shelves have been identified and some of these have been the subject of extensive field campaigns aimed at better understanding the controls on ice streams (c.f. Alley and Bindschadler, 2001, for a collection of papers summarizing earlier findings). These ice streams appear to be capable of rapid changes including margin migration and complete shut-down. Based on mapping of grounding-line positions using satellite radar interferometry, Rignot (1998) inferred a retreat of 1.2 ± 0.3 km/yr between 1992 and 1996, corresponding to a thinning rate of 3.5 ± 0.9 m/yr. Satellite radar altimetry over the period 1992 to 1999 confirmed this inferred thinning rate and also showed thinning to extend far into the interior (Shepherd and others, 2002). Comparison of the satellite altimetry data with airborne laser altimeter surveys showed that thinning rates near the coast during 2002-2003 were significantly larger than those observed during the 1990s and the Amundsen Sea sector of the West

Antarctic Ice Sheet appears to be out of balance by as much as 60% (Thomas and others, 2004b). In the Antarctic Peninsula, several of the peripheral ice shelves have disintegrated or retreated, with a total area in excess of 14,000 square kilometers lost over the past two decades. Vaughan and others (2003) linked ice-shelf collapse to southward migration of the -9 ^{o}C isotherm, presumed to correspond to the thermal limit of ice-shelf viability. While Vaughan (1993) reported no significant acceleration of input glaciers following the breakup of the Wordie Ice Shelf, elsewhere in the Peninsula ice-shelf break-up has led to flow acceleration of grounded glaciers (e.g. Rott and others, 2002; De Angelis and Skvarca, 2003; Rignot and others, 2004b). These observations are suggestive that the West Antarctic Ice Sheet may be on the verge of contributing to future sea-level rise and have reinvigorated the long-standing debate about the stability of this marine-based ice sheet and to what extent buttressing ice shelves control drainage from the interior.



Box 3. Ice velocity (meters per year, logarithmic scale) of Hektoria, Green, Evans, Jorum, Crane, Flask, and Leppard Glaciers, Antarctic Peninsula, measured using (a) ERS-1/2 data from Jan. 1996, and Radarsat-1 data from (b) Oct. 2000; and (c) Dec. 2003. Grounding lines are thick black and ice front locations are thin white.

From: Rignot and others (2004b).

In a recent editorial, NASA's James Hansen stated that "ice sheet models cannot be used with confidence for assessing expected sea level change until they demonstrate an ability to reproduce ice sheet disintegration such as the Heinrich events, with realistic forcing yielding realistic rates of ice sheet demise" (Hansen, 2005, p. 273). Indeed, many recent as well as past events suggestive of rapid ice-sheet change cannot be reproduced by the current generation of whole ice-sheet models on which the predictions issued by the IPCC are primarily based. Developing the next generation of more realistic ice-sheet models requires a multi-tiered comprehensive approach, designed to include a hierarchy of models ranging from theory and perhaps intuitive conceptual models based extensively on observations, to time-evolving flowline models aimed at simulating individual glaciers or drainage basins, to fully three-dimension time-dependent thermo-mechanical models that simulate evolution of the entire ice sheet.

Challenges facing the glaciological modeling community are two-fold. First, many of the controlling processes are not well understood. For example, the transition from grounded ice to floating ice shelves and the role of longitudinal stress gradients in this transition remains disputed despite more than two decades of theoretical and modeling efforts. Fortuitously, ongoing ice shelf collapse in the Antarctic Peninsula as well as collapse of fjord glaciers in Greenland, offer the opportunity to study grounding-line instability at various stages of ice-shelf retreat and collapse. Coupled with advances in remote-sensing techniques that allow us to better constrain subglacial conditions (small-scale bed topography, wet versus frozen bed, etc.), major advances in understanding physical processes are possible through concerted multi-disciplinary efforts that involve targeted field campaigns, spaceborne remote sensing, and theoretical developments and data interpretation and assimilation.

A second challenge is posed by the need to implement processes acting on small spatial scales into large-scale numerical ice-sheet models. For example, the ice streams draining into the Ross Ice Shelf are laterally bounded by shear margins no more than a few km wide. Over this narrow region, a hundredfold speed contrast is developed between the near-stagnant interstream ridges and the fast-moving ice streams and the stress regime transitions from lateral shear dominating on the ice-stream side to vertical shear on the ridge side (e.g. Van der Veen and others, 2007a). Adequate modeling of these shear margins would require a horizontal grid spacing that is small compared to the width of the shear zone, that is, no more than a few hundred meters. For computational reasons, it is not feasible to apply such fine spatial resolution to the entire Antarctic Ice Sheet, thus necessitating either the inclusion of small-scale processes in a parameterized form, or the use of “nested” models in which high-resolution models of ice streams are embedded within models that calculate the evolution of the whole ice sheet on a coarser grid. This problem is not unique to glaciology and in particular the climate modeling community has made great progress towards incorporating local processes (e.g. cloud formation) in General Circulation Models. Some of these methods and techniques may be applied to glaciological whole ice sheet models.

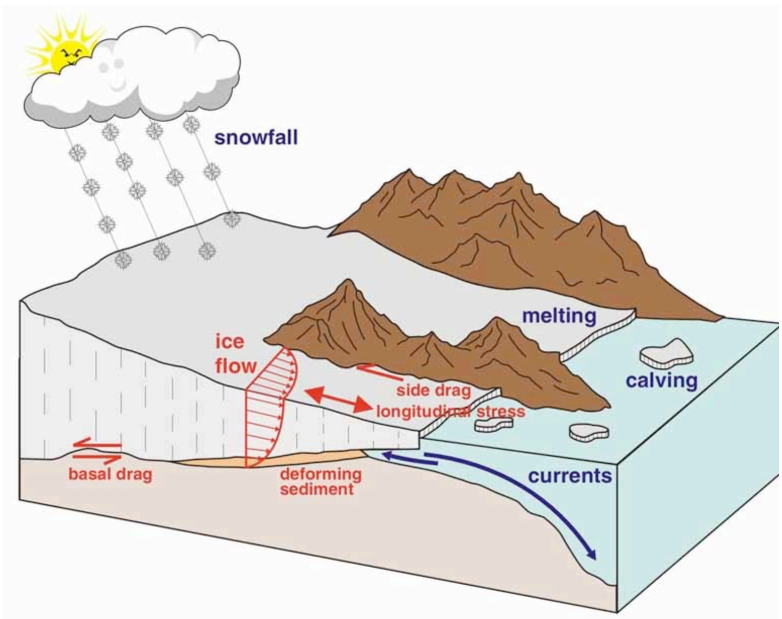
Recognizing the need for improving numerical ice-sheet models, the objective of this document is to identify major challenges and to suggest strategies for overcoming these. First, we discuss some of the over-arching uncertainties and processes that require better understanding before being incorporated into the next generation of ice sheet models. To address each of these, we outline possible modeling strategies and specific research questions. We realize, of course, that as our understanding improves, adjustments of the modeling strategies and research priorities may be appropriate. Therefore, this document should be viewed as modifiable as needed.

2. PROBLEM STATEMENT

Prognostic ice-sheet models are based on conservation of momentum, energy and mass. Stress-balance equations, coupled with a constitutive relation linking strain rates or velocity gradients to stresses, allow ice fluxes to be evaluated throughout an ice sheet. Evolution of the ice sheet is then evaluated by solving the time-dependent mass-continuity equation. Because flow of glaciers is strongly dependent on the englacial and basal temperature regime, the thermodynamical equation must be solved simultaneously for the ice temperature and freezing or melting at the glacier base (c.f. Van der Veen and Payne, 2004). The governing equations are solved on a three-dimensional numerical grid with a horizontal spacing of 20-50 km and typically some 20 vertical layers. For the whole Antarctic Ice Sheet, this amounts to *ca.* 10^6 gridpoints, while for the smaller Greenland Ice Sheet, the total number is *ca.* 10^5. To keep the integration scheme stable for finite-difference models, the time step must be of the order of 1 year.

In principle, the horizontal grid spacing employed in numerical models could be decreased to better capture small-scale features such as ice-stream margins and fast-moving outlet glaciers. However, this would require a smaller time step and hence make the model unpractical to conduct multiple simulations spanning longer time periods (e.g. one or more glacial cycles). Therefore, there is a need to develop schemes to incorporate small-scale processes into whole ice-sheet models. It should be noted that even if interest is in the evolution over the next few hundred years, longer time integrations are required to capture the slow component of ice-sheet adjustment to climate forcings on time scales corresponding to glacial cycles.

A second difficulty is associated with the non-linearity of the constitutive relation for glacier ice. As a result, the momentum equations cannot be solved directly and either the stress distribution has to be derived iteratively for each time step, or simplifying assumptions have to be made to estimate the stress distribution in the ice sheet. The most common simplification is the so-called Shallow Ice Approximation (SIA) in which the gravitational driving stress is taken to be balanced by drag at the glacier bed. Other sources of flow resistance (lateral drag and gradients in longitudinal stress) are not included. While the SIA provides a good description of the flow in interior parts of the ice sheets, this approximation breaks down on ice streams and outlet glaciers where friction at the lateral margins is often important (e.g. Hindmarsh, 2004). Further, on floating ice shelves, gradients in longitudinal stress are important and hence there must be a transition zone over which the portioning of flow resistance changes from the ice-shelf regime, to the ice-stream regime, to the interior flow regime. The nature of this transition region, and how it evolves over time, is not well understood.



Box 4. Schematic ice-sheet cross section illustrating the transition from shear-dominated laminar flow in the interior, to streaming flow characterized by low basal resistance due to the weak underlying bed, to ice shelf-spreading.

The above considerations apply to numerical issues associated with developing ice-sheet models, and to physical processes associated with the englacial environment, or the modeled ice sheet itself. In addition, to run prognostic models, conditions at the model boundaries have to be imposed. These include the upper ice surface, the ice-bed interface, and the lateral ice margin (the ice-ocean or ice-land boundary). For each of these boundaries, conditions are only partially known and improved understanding of current, past, and future boundary conditions is essential for developing more realistic scenarios for ice-sheet evolution in the (near) future.

Based on the natural division of the model domain into the glacier itself and the various physical boundaries, Section 3 discusses in more detail the overarching processes that need to be addressed. Section 4 outlines in more detail some specific research questions to be solved and data requirements, as well as issues pertaining to model numerics.

3. RESEARCH FOCUS AREAS

Based on recent results and observations, we have identified the following processes for which understanding is currently lacking but that are crucial to include in ice-sheet models:

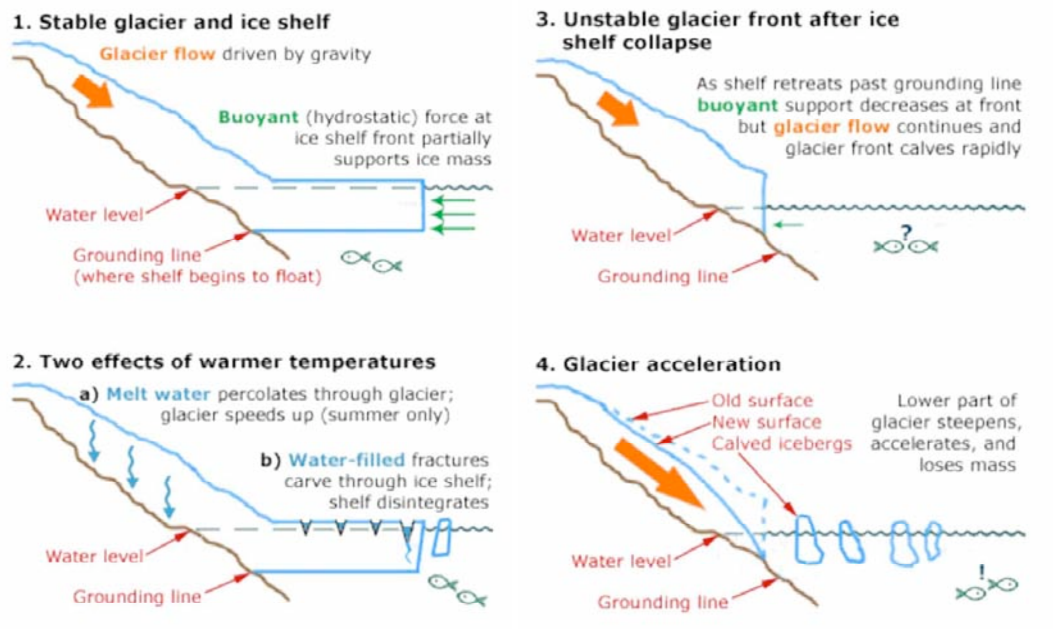
- Englacial: interaction between ice-marginal processes and discharge from the interior through upstream transmission of longitudinal stress perturbations.
- Lateral margins: environmental controls on ice-shelf collapse, and controls on iceberg calving.
- Basal boundary: geologic and topographic controls on fast glacier flow, and the role of subglacial hydrology on glacier discharge.
- Surface boundary: controls on surface mass balance.

Below, each focus area is described in more detail. From this discussion, it will become clear that these areas are interrelated and should be addressed in close collaboration rather than in isolation.

3.1. Interaction between ice-marginal processes and discharge from the interior

Ever since the pioneering work of Mercer (1968; 1978) and Weertman (1974), glaciologists have speculated that removal of peripheral ice shelves and floating ice tongues will result in increased discharge of interior ice. Over the last two decades, ice shelves in the Antarctic Peninsula have disintegrated, believed to be in response to a local warming trend that caused the thermal limit of ice-shelf viability to migrate progressively southward (Vaughan and Doake, 1996; Vaughan and others, 2003). Velocity measurements on grounded glaciers formerly draining into these ice shelves indicate a speed up followed collapse of the ice shelves (De Angelis and Skvarca, 2003; Scambos and others, 2004). In Greenland, the speed of Jakobshavn Isbræ nearly doubled from 6 km/yr in 1995 to 12 km/yr in 2005 (Rignot and Kanagaratnam, 2006) possibly in response to the weakening and subsequent break up of the floating terminus of this glacier (Joughin and others, 2004a; Thomas, 2004).

While many in the glaciological community have interpreted these observations as evidence for the instability models proposed by Mercer and Weertman, and many others since, a more careful analysis of the sequence of events is needed to establish unambiguously to what extent forcings at the calving front propagate upstream and influence discharge from the interior. For example, it could be that increased speeds on the grounded portions resulted from the same surface melt event(s) that led to the collapse of the floating part, rather than reflecting the glacier response to loss of ice-shelf buttressing. If, indeed, discharge from the interior is



Box 5. In Antarctica, glaciers flowing to the coast form ice shelves — thick platforms of ice that float on the ocean. Together, the glacier and ice shelf form a stable system, but this system can lose its stability in response to warmer temperatures.

Warmer summer temperatures sometimes result in glacier acceleration as melt water percolates through the glacier to its base. Here the water lowers the friction between the glacier and the underlying rock. This effect is seasonal, and with the ice shelf in place, the glacier returns to a lower flow speed once summer (and surface melting) ends.

Warmer summer temperatures can also lead to rapid ice shelf disintegration. As temperature rises, melt water accumulates on the shelf surface. Although only a tiny fraction of the ice shelf melts, the water infiltrates the shelf through small cracks in the ice. Over time, the weight of the melt water in the cracks shatters the shelf. This happened in the Antarctic Peninsula in 1995 and again in 2002.

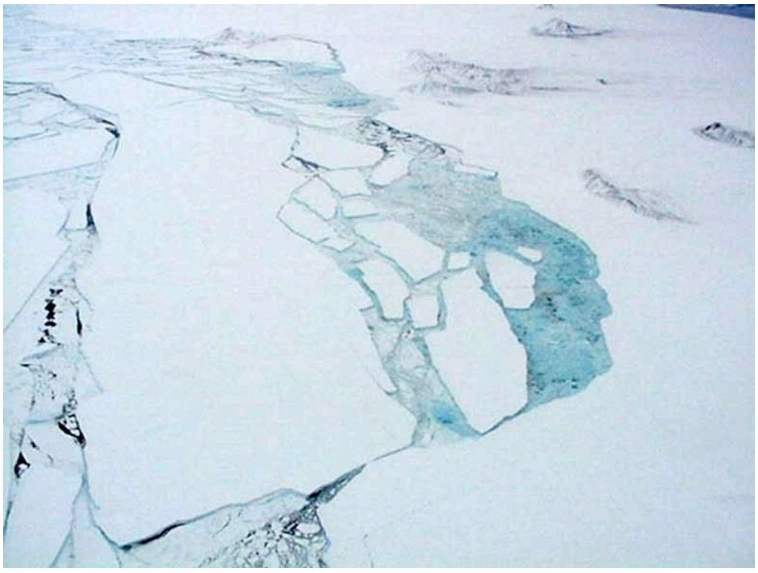
Removal of the ice shelf causes much more dramatic glacier acceleration by reducing two forces that counteract glacier flow. One counteracting force is "backstress" produced by islands or coastline underlying the original shelf. Another is the buoyant (hydrostatic) force of the seawater against the front of the shelf or glacier. A full explanation will require numerical modeling of glacier flow, but observations to date suggest that ice shelves act as "braking" systems on the glaciers behind them.

Diagram by Ted Scambos and Michon Scott, National Snow & Ice Data Centre

affected by ice-marginal processes, an important question is whether ice-shelf collapse necessarily leads to irreversible glacier retreat or whether interior flow will adjust to the perturbation towards a new equilibrium (e.g., Payne and others, 2004). Csatho and others (in press) document thinning of the lower reaches of Jakobshavn Isbræ over the period 1944 to 1953, following thinning of the terminal region and transition from a grounded terminus to a floating calving front. Thinning did not persist on this glacier, however, and for most of the second half of the 20th century, a period during which mean summer temperatures declined, Jakobshavn was close to steady state culminating in a brief period of thickening during the 1990s, before the onset of current rapid thinning as summer temperatures increased. Similarly, modeling experiments on the response of Pine Island Glacier to a fixed perturbation at the grounding line indicate that a new equilibrium is reached after ~150 years following an imposed instantaneous change on the ice plain (Payne and others, 2004).

3.2. Environmental controls on ice-shelf collapse

As noted, disintegration of ice shelves in the Antarctic Peninsula has followed the southward migration of the -9 °C mean annual isotherm, suggesting that atmospheric temperature is the main determinant on whether floating ice shelves can maintain their integrity. Scambos and others (2000) proposed that ice-shelf collapse is initiated by meltwater ponding in surface crevasses. If the influx of meltwater is sufficiently large to fill crevasses almost entirely, the tensile force introduced by the water (which has a greater density than ice) allows the crevasse to propagate downward and ultimately penetrate the full thickness of the shelf (Van der Veen, 1998, 2007). MacAyeal and others (2003) suggest that subsequent capsizing of narrow ice-shelf fragments forces rifts open, thus contributing to further ice-shelf fragmentation. While this mechanism appears to explain current break-up events, the Holocene record of Antarctic Peninsula ice shelves suggests other factors may play a role in ice-shelf stability.



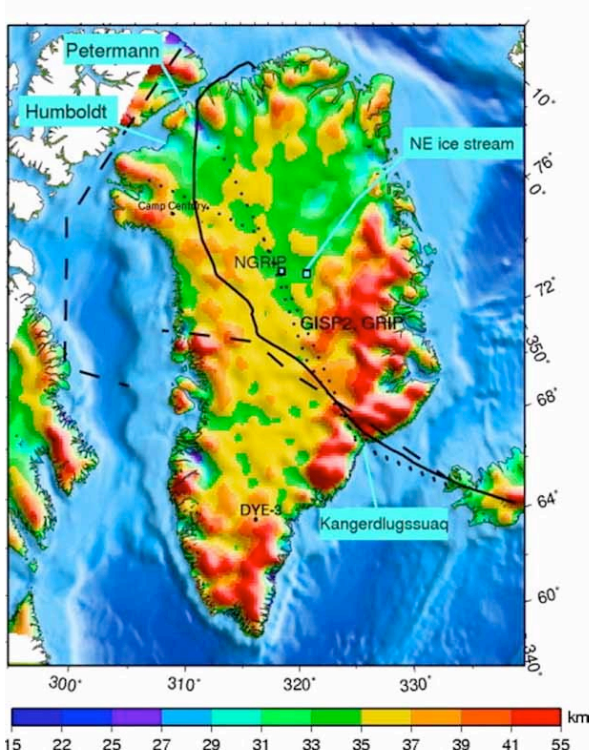
Box 6. Disintegration of Larsen B Ice Shelf south of the Seal Nunataks section (to the right).

Photo taken on 13 March, 2002, during the Global Positioning Satellite survey by Pedro Skvarca, Instituto Antartico Argentino.

Ice-shelf records reconstructed from sediment cores indicate that the Larsen B Ice Shelf, which has now collapsed, remained stable throughout the Holocene (Domack and others, 2005). Other ice shelves in the Antarctic Peninsula, however, experienced retreat during warm periods in the early and middle Holocene. The Prince Gustav Channel Ice Shelf, which collapsed in 1995, was also absent *ca* 5000-2000 years B.P. (Pudsey and Evans, 2001). The George VI Ice Shelf, which is present today, retreated in the early Holocene, *ca* 9600-7900 years B.P. (Bentley and others, 2005) when warmer surface waters were present over the continental shelf of the Peninsula (Leventer and others, 2002). Hodgson and others (2006) suggest that both atmospheric temperature and warm ocean currents contributed to previous retreats of ice shelves in the Antarctic Peninsula in a spatial pattern that is different from what is seen today.

3.3. Geologic and topographic controls on fast glacier flow and the role of subglacial hydrology

Several recent studies have pointed to the influence that subglacial geology may have on the onset of fast glacier flow. Blankenship and others (2001) reviewed available aerogeophysical observations to identify geological controls on the initiation of fast glacier flow in the southeastern Ross Embayment. These authors proposed three possible geological controls, namely (1) the availability of sediment deposits; (2) spatial gradients in geothermal heat flux associated with variations in crustal thickness (where crustal thickness is comparatively small, geothermal heat flux is likely to be large); and (3) focused geothermal flux associated with localized volcanism or, alternatively, with extensive interlayered (or capping) volcanics within fault-bounded sedimentary basins. Additionally, geothermal heat concentrations may occur where pronounced topographic relief is present (e.g. Lachenbruch, 1968; Van der Veen and others, 2007b).

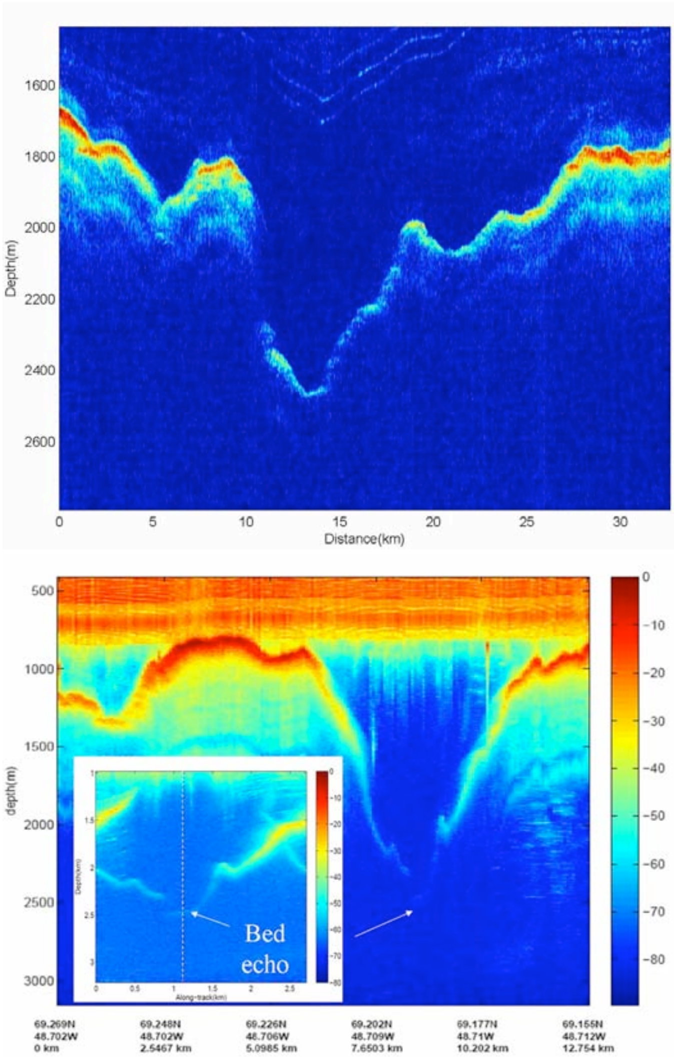


Box 7. Crustal thickness in Greenland obtained from generalized isostatic analysis. Crustal thickness minima correlate with identified sub-glacial melt and fast glaciers. In particular, the NorthEast Ice Stream coincides with the thinnest crust in central Greenland.

From: Braun and others (2007).

An important control on fast glacier flow is the basal temperature regime and the presence of lubricating meltwater or water-saturated sediments. Observations suggest localized melting in northern Greenland that contradict the distribution of present-day basal melt under the Greenland Ice Sheet derived using a high-resolution, three-dimensional thermo-mechanical ice-sheet model (Huybrechts, 1996). Assuming a spatially-uniform geothermal heat flux of 42 mW/m^2, Huybrechts (1996) found that most of the ice sheet is frozen to the bed with melting confined to lower elevations in the northeast and western regions. However, Fahnestock and others (2001) inferred basal melt rates up to 0.2 m/yr under the onset region of the Northeast Ice Stream and its southern tributaries that can only be explained by a geothermal heat flux much greater than the continental background value used by Huybrechts (1996). Similarly, at the NGRIP core site, basal melt rate reaches 7.5 mm ice per year, and the modeled geothermal heat flux is between 90 and 160 mW/m^2 along the flowline originating 50 km upstream of the drill site (Dahl-Jensen and others, 2003).

Whereas for the Siple Coast ice streams in West Antarctica there is no obvious topographic control on their position, the tributaries feeding these ice streams are well defined topographically (Joughin and others, 1999). Many other fast-moving glaciers are associated with subglacial valleys. Preliminary analysis of bed profiles derived from ice-penetrating radar indicates that many of the major drainage routes in the Greenland Ice Sheet are associated with bedrock troughs. Many of these troughs are narrow and deep and not well-captured on the continental-scale bed-elevation map, but can be readily seen on radar profiles or inferred from seismic traverses. Such valleys can have a twofold effect on the flow of outlet glaciers. First, geothermal heat flux is concentrated on the floor of deep trenches (e.g. Lees, 1910) thus providing more heat to the glacier base than in surrounding areas. For the subglacial trench under Jakobshavn Isbræ, Van der Veen and others (2007b) estimate a local doubling of heat flux. Second, the greater ice thickness over subglacial valleys increases the driving stress as well as basal temperatures. Thus, subglacial valleys tend to create conditions favorable for fast glacier flow. This flow enhancement is opposed by the greater contact area between basal ice and the bed compared to a flat bed topography.



Box 8.

Top panel: Ice-penetrating radar profile across a subglacial trench under Petermann Glacier, northwest Greenland. This trench, with a depth of 500 m, is overlain by 2 km thick ice.

Bottom panel: radar profile across the subglacial trench under Jakobshavn Isbræ, west Greenland. Inset shows the bed echo from the deepest part in greater detail .

From: Van der Veen and others (2007b).

It is well-established that the presence of a lubricating water layer at the base of a glacier allows the ice to slide over its bed, thereby permitting speeds to be reached well in excess of those resulting from internal deformation. Further, on smaller mountain and tidewater glaciers, short-term fluctuations in ice drainage are generally associated with variations in subglacial water pressure, ice front position, and/or water storage under the glacier (e.g. Meier and others, 1994).

Zwally and others (2002) observed increased speed on the Greenland Ice Sheet, about 45 km north of Jakobshavn Isbræ following surface melt events. Similar speed-up events have been observed on tidewater glaciers (e.g. Columbia Glacier, Alaska; Meier and others, 1994) as well as on mountain glaciers Findelengletscher, Switzerland, Iken and Bindschadler, 1986). The observations near Jakobshavn are somewhat surprising, however. At the location of the measurements reported by Zwally and others (2002) the ice thickness is ~1.5 km and it is not clear how water can migrate from the surface to the glacier bed over periods of days, considering that the upper portion of the glacier is sufficiently cold to refreeze any meltwater percolating downward. While measurements conducted on Jakobshavn Isbræ proper in 1984 indicated a lack of seasonal variation in glacier speed (Echelmeyer and Harrison, 1990), more recent observations indicate a seasonal variation in velocity near the grounding line in 1995, prior to retreat and speed up (Luckman and Murray, 2005). If the observations of Zwally and others (2002) are broadly indicative of ice-sheet behavior, future warming may lead to increased water availability at the glacier bed, thereby increasing ice discharge. Parizek and Alley (2004) showed that including the effect of enhanced basal lubrication by surface meltwater increased the rate of ice-sheet discharge and consequent sea-level rise compared to model simulations that did not include this feedback.

3.4. Controls on surface mass balance

The mass balance of the polar ice sheets is determined by the difference between mass input and mass loss from surface ablation and meltwater runoff, sublimation, bottom melting under peripheral floating parts, and from iceberg calving. Snowfall constitutes the main source of mass added to the ice sheets and, consequently, changes in snowfall may significantly alter their net balance. For predicting future evolution of the Greenland and Antarctic Ice Sheets it is imperative to include more realistic models or parameterizations of interior snow accumulation and sublimation. On short time scales, climate forcing of ice sheets is achieved through changes in the net surface mass balance, usually related to temperature changes. For example, Huybrechts and De Wolde (1999) make the assumption that changes in precipitation (snow and rain) may be related to the present distribution and perturbed for different temperatures according to sensitivities derived from ice cores. The assumption that temperature can be used as sole predictor for changes in accumulation rate may be questioned, however. Cuffey and Clow (1997) investigated the relation between temperature and accumulation as recorded in the GISP2 record and found that long-term (500 yr) averages are inversely correlated during most of the mid and late Holocene (from *ca.* 7 ka B.P. to the present). On time scales of a century, accumulation rate and temperature are essentially uncorrelated and Cuffey and Clow (1997) conclude that there is little historical evidence for expecting climate warming to result in increased precipitation over the Greenland Ice Sheet. Kapsner and others (1995) inspected the same GISP2 record and found that atmospheric circulation, rather than temperature, was the primary control on snow accumulation in central Greenland over the past 18,000 years. On a much shorter time scale, Monaghan and others (2006) derived a 50-year time series of snowfall accumulation over the Antarctic continent by combining model simulations with in situ observations primarily from ice cores. They conclude that there has been no statistically significant change in snowfall over the second half of the 20^{th} century despite winter warming of the overlying atmosphere. Moreover, a detailed analysis of temporal variability in accumulation over the last two centuries using firn cores and snow radar shows no significant increase in accumulation in most of East Antarctica (Frezzotti and others, 2007; Urbini and others, in press). These findings cast doubt on sea-level predictions suggesting a negative contribution from the Antarctic Ice Sheet resulting from increased precipitation.

Modeling the surface mass balance over the Greenland and Antarctic ice sheets requires a spatial resolution that exceeds that used by GCMs used to simulate climate change over longer time periods. This need is primarily dictated by the narrow and steep margins of the ice sheets where orographic uplift can exert an important influence on the precipitation. Failure to incorporate these steep slopes, means that most GCMs tend to overestimate precipitation in the interior (Ohmura and others, 1996; Glover, 1999; Murphy and others, 2002). At the same time, the smooth topography prescribed in GCMs tends to over-estimate the extent of the low-altitude marginal zone, resulting in over-estimation of surface melting (Glover, 1999; Wild and others, 2003). While increasing the grid resolution of atmospheric models improves the computed surface mass balance, considerable uncertainty remains as to the predicted increase in accumulation following climate warming (Van de Wal and others, 2001; Huybrechts and others, 2004). Box and others (2006) used the fifth-generation Pennsylvania State University – National Center for Atmospheric Research Mesoscale Model modified for use in polar regions (Polar MM5) to investigate variability in the surface mass balance of the Greenland Ice Sheet over the period 1988-2004. They conclude that, despite calibration with independent observations and apparently coherent and systematic increases in both accumulation and ablation as temperatures increased, an improvement by at least a factor of two in the accuracy of the model is required to arrive at more definitive assessments of the terms in the surface mass budget.

Regarding the surface mass balance in Antarctica, large uncertainties remain due lack of knowledge about processes driving precipitation and post-depositional redistribution (Eisen and others, in press). The area of katabatic winds covers about 90% of the Antarctic surface and transport processes represent an important component of snow accumulation at scales ranging from 10 meters to tens of kilometers (Frezzotti and others, 2005). Various difficulties complicate measurements of the surface mass balance, in particular unrepresentative spatial and temporal variability, measurement inaccuracy, and lack of quality control (Magand and others, 2007). It is known that sublimation is not negligible in Antarctica (e.g. Bintanja, 1998; Gallée, 1998; Gallée and others, 2001; Van den Broeke and others, 2004) but the snowdrift process (e.g. wind sublimation, transport, export into the ocean, etc.) is not explicitly included in numerical weather forecasting and general circulation models (Gallée and others, 2001; Genthon and Krinner, 2001; Krinner and others, 2007). One of the largest areas of uncertainty regarding surface mass balance is the role of surface and wind-driven sublimation (e.g. Genthon and Krinner, 2001). Frezzotti and others (2004) pointed out that snow precipitation is homogeneous on large scales (hundreds of square kilometers) but wind-driven sublimation phenomena determined by the surface slope in the prevailing wind direction have a considerable impact on the spatial distribution of snow over short (tens of meters) to medium (kilometers) spatial scales (Frezzotti and others, 2002) and can export large quantities of snow into the atmosphere and into the surrounding oceans (Frezzotti and others, 2007). These processes are very significant for evaluating past, present, and future surface mass balance and must be taken into account in atmospheric models and in comparisons of model results against observations.

Box 9. The wind blown ice surface at Law Dome, Antarctica.

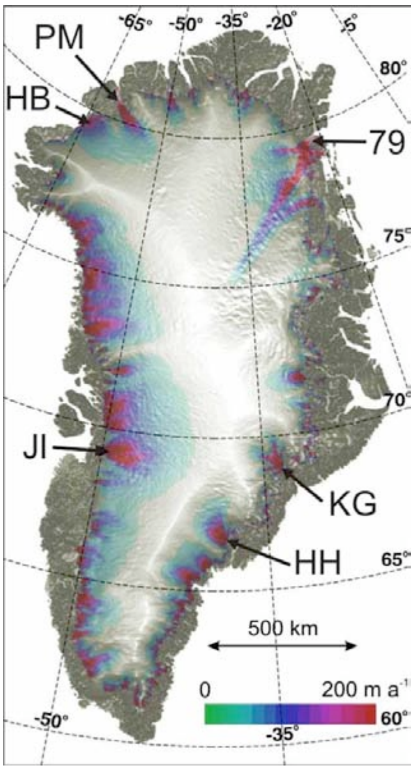
Photo by Tas van Ommen, Australian Antarctic Division and Antarctic Climate & Ecosystems, Cooperative Research Centre.

4. SPECIFIC RESEARCH EFFORTS

The ultimate objective is to develop prognostic numerical models that better incorporate non-linear ice-sheet response to environmental forcings such as change in surface mass balance, loss of buttressing from floating ice shelves and ice tongues, and rising sea level. This will require a targeted approach by the glaciological community. Process-oriented studies should be driven by observations and indicate where additional measurements are needed. The resulting improved physical understanding must then be incorporated into prognostic numerical ice-sheet models. In this section, we outline in broad terms some pressing research questions and strategies for addressing these.
Several recent developments allow a systematic investigation of how changes at the ice margin affect drainage from the interior, including more complete temporal data coverage and advances in numerical modeling. Quantitative records of elevation and velocity change are available for a number of outlet glaciers, and these observations can be used to test various models that have been proposed for ice-sheet response to ice-marginal forcing.

4.1. Potential study sites

The most complete velocity measurements pertain to Jakobshavn Isbræ, dating back to 1985, well before the recent velocity increase (Fastook and others, 1995; Joughin and others, 2004a). This quantitative record can be extended farther back to the mid 1940s using vertical and oblique aerial photographs collected at various times from 1944 to 1987 and archived by Kort- og Matrikelstyrelsen (KMS, National Survey and Cadastre of Denmark) and available as diapositives. Combining these data with high-resolution Digital Elevation Models (DEM) derived from laser altimetry and ICESat, and basal elevation maps obtained from radar sounding, allows quantitative analysis of the stress fields across the grounding line, and of temporal changes therein. This particular glacier is further noteworthy as the largest drainage conduit in the Greenland Ice Sheet, and has been identified as being prone to irreversible retreat. This combination of more than half a century of data and potential instability, make this glacier a natural candidate for coordinated investigations. Other Greenland outlet glaciers for which recent velocity changes have been documented include Helheim and Kangerdlugssuaq glaciers in the south-east (Howat and others, 2005, 2007).

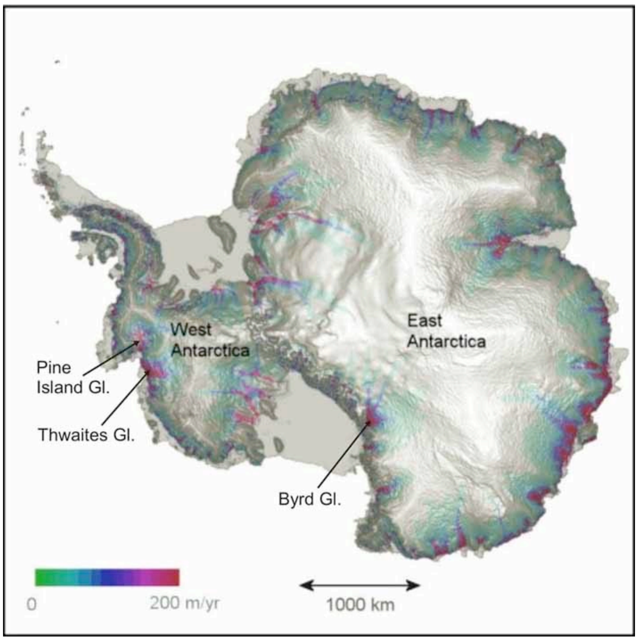


Box 10. Shaded topographic map of the Greenland Ice Sheet with ice flow velocities superimposed in colour to show the location of the key outlet glaciers. Maximum velocity shown is 200 m/yr, although ice flows faster than this toward the margins. JI: Jakobshavn Isbrae; HB: Humboldt Gl.; PM: Petermann Gl.; 79: 79-North Gl.; KG: Kangerdluggssuaq Gl.; HH: Helheim Gl.

From: Bamber and others (2007).

In addition to focusing on outlet glaciers currently undergoing change, observation programs should be initiated on those glaciers that have the potential for exhibiting rapid retreat and thinning as climate continues to warm. These include a number of glaciers that have an overdeepend basin below sea level in the north of Greenland (Humboldt, Petermann and 79-North glaciers) and that are expected to respond in a similar way as Jakobshavn Isbrae in a wamer climate. Further, there are a number of tidewater glaciers north of Ilulissat that are of similar size and these could advance in velocity and thinning if melt water lubrication or ice-front retreat is a driving force.

In Antarctica, several potential study areas can be identified. While most efforts have thus far concentrated on the large ice streams draining into the Ross and Ronne ice shelves, Pine Island and Thwaites glaciers have recently undergone dramatic changes whose documentation permits quantitative investigation of what has been termed the "weak underbelly of the West Antarctic Ice Sheet" (Hughes, 1981). Indeed, with ice-shelf collapse continuing its progression southward, these two glaciers as well as numerous others in the Antarctic Peninsula and Amundsen Sea sector, offer excellent opportunity to study their adjustments to grounding-line forcing. The ice-ocean interaction might be a driving force in some of the ice shelf thinning and marginal changes in that region. Oceanographic measurements are needed to monitor the vertical structure of the ocean the seasonal to interannual variability. Ice shelf models couples with oceanic models (currents, energy transport) are needed to study the potential feedback on these floating ice tongues.



Box 11. Shaded topographic map of the Antarctic Ice Sheet with ice flow velocities superimposed in colour to show the location of the key outlet glaciers. Maximum velocity shown is 200 m/yr, although ice flows faster than this toward the margins.

From: Bamber and others (2007).

Another major drainage route long believed to be relatively stable is Byrd Glacier, draining a large part of East Antarctica through the Transantarctic Mountains into the Ross Ice Shelf. Comprehensive velocity mapping was conducted in 1979 using aerial photogrammetry (Brecher, 1986) and these results have been used to investigate the flow dynamics of this glacier (Whillans and others, 1989). By registering the aerial imagery to ICESat data, Schenk and others (2005) found that the grounded part of this glacier was close to balance over the period 1979 to 2004, but large thinning was observed on the floating part, possibly associated with sub-shelf melting. Subsequent comparison of DEMs derived from ASTER images indicated this thinning occurred somewhere between January, 2001, and January, 2002 (L. Stearns, pers. comm., 2007). This significant change directly downstream of the grounding line did not have an immediate effect on the discharge from the interior (Stearns and Hamilton, 2005). More recent velocity determination suggest, however, that the speed along most of Byrd Glacier increased by *ca.* 50 m/yr (L.

Stearns, pers. comm., 2007). Quantitative assessment of stress-regime changes on Byrd Glacier is hindered by the lack of available bed topography.

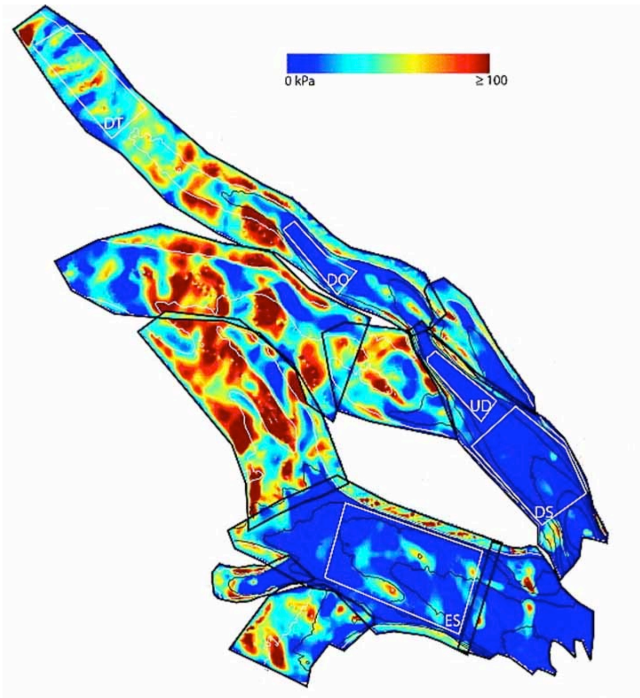
The importance of observing a range of grounding-line behaviors is evidenced by Vieli and Payne (2005) who compared various model formulations applied to the study of marine ice sheets. They found that predicted grounding-line migration is dominantly controlled by the way grounding line motion is treated in the numerical model (e.g. fixed grid versus a moving grid), and how the governing equations are discretisized. Physics incorporated into the numerical models, such as longitudinal momentum coupling between the ice shelf and the grounded ice sheet, appeared to be of secondary importance only. The implication of this model comparison is that there is an urgent need to develop better models whose predictions are not dictated by numerical specifics. As concluded by Vieli and Payne (2005), “further model development also requires a better observational history of grounding line migration (in terms of both the timing and spatial extent) and also indicates the need of a test data set for the modeling community.”

4.2. Hypothesis testing

Hughes (1986, 1998) proposed the Jakobshavn Effect, an instability whereby a set of positive feedbacks may result in substantial loss of ice from the interior. The trigger for this effect is increased calving and retreat of the calving terminus, allowing more rapid ice discharge upstream of the calving terminus, resulting in lowering of the ice surface and, ultimately, increased discharge of ice from the interior. Thomas (2004) presents a force-perturbation analysis to model upglacier propagation of ice-marginal perturbations associated with calving events and weakening of the floating ice tongue. According to that analysis, acceleration of Jakobshavn Isbræ may have resulted from a large calving event in 1997/98. Inspection of satellite images shows that several of the speed increases on this glacier coincide with calving events as the floating ice tongue disintegrated (Joughin and others, 2004a).

On the other hand, the velocity increase on Jakobshavn Isbræ may have resulted from enhanced meltwater penetration to the glacier bed, or from continued thinning and approach to flotation. This latter mechanism was proposed by Pfeffer (2007) to explain irreversible retreat of tidewater glaciers such as Columbia Glacier, Alaska. Further, the comparatively short duration of speed-up on Kangerdlugssuaq and Helheim glaciers in east Greenland indicates that perturbations may be modulated as the glacier adjusts and returns to near-balance conditions (Thomas, 2004; Howat and others, 2007). Howat and others (2005) investigate the effect of retreat of the calving front of Helheim Glacier by calculating the net gravitational driving force acting on its lower trunk. They propose that the observed speed-up resulted from two effects, namely increased effective stress over the glacier’s main trunk following retreat of the calving front, with subsequent thinning propagating upglacier resulting in steeper surface slopes and velocity increase. Retreat of the calving front terminated where the bed slope reversed, allowing the glacier to re-equilibrate (Howat and others, 2007).

It has long been recognized that the subglacial topography can be a deciding factor in the stability of grounding lines and tidewater calving fronts (Thomas, 1979) thus accurate mapping of the subglacial topography is necessary, especially in the vicinity of the grounding line. Earlier seismic studies revealed the existence of a deep trench under Jakobshavn Isbræ (Clarke and Echelmeyer, 1996) but only recently has improved radar technology allowed high-resolution mapping of the bed under this glacier (Gogineni and others, 2006; Lohoefener and others, 2006). Similar mapping campaigns are planned for other outlet glaciers, particularly Thwaites Glacier.



Box 12. Basal shear stress from the viscous bed inversion for Ice Streams D and E. Thick black lines outline the regions over which individual inversions were performed. Flow speed is contoured at 50 m/yr (thin white line) and at 100 m/yr intervals (thin black lines).

From: Joughin and others (2004b).

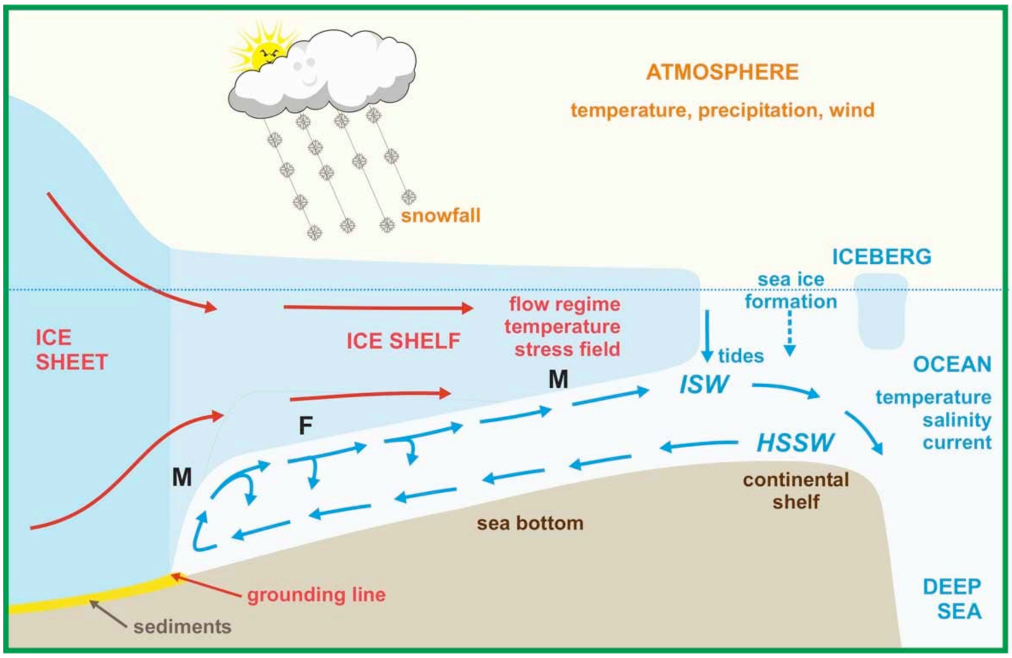
The transmission of stress perturbations from the terminus to the interior can be investigated by analyzing velocity fields. The procedures for this are well established and have been applied successfully to many fast-moving glaciers. Van der Veen and Whillans (1989) developed the so-called force-budget technique in which measured surface velocities are inverted to resistive stresses using the flow law for glacier ice to identify sites where resistance to glacier flow is located. A different approach, based on control methods, was developed by MacAyeal (1992). Both methods are compatible and can be applied to available data (e.g. Van der Veen, 1999; Joughin and others, 2004b).

4.3. Ice-shelf collapse

While collapse of floating ice shelves is commonly linked with the average summer temperature and the so-called "thermal limit of ice-shelf viability" the details of this process are not unambiguously established. In Greenland the onset of surface melting correlates with increased iceberg production, suggesting meltwater ponding affects the integrity of floating ice tongues. However, for most of the second half of the 20th century the position of the calving front of Jakobshavn Isbræ remained stable with annual advance and retreat cycle of *ca*. 2.5 km amplitude (Sohn and others, 1998). In 1999 progressive retreat of the calving front started immediately following onset of rapid thinning (Csatho and others, in press), and continues to this day. It is not clear what caused the terminus to retreat after the prolonged period of stability. Joughin and others (2004a) suggest that abnormally high melting during the 1998-2001 period may have resulted in increased calving and retreat of the ice tongue beyond its stable position attributed to resistance from fjord walls and pinning points.

As noted above, the Holocene record of ice-shelf change in the Antarctic Peninsula suggests oceanic circulation may be a contributing factor in break up. Thinning rates on the floating part of Jakobshavn Isbrae were on the order of several hundred meters per year from 1999 onward, following a period of thickening starting *ca*. 1985

(Csatho and others, in press). Such great mass loss cannot be explained by surface ablation and strongly suggests strongly enhanced bottom melting, likely as a result of altered circulation and temperatures in the fjord. Without observations and modeling of water circulation in the fjord and in the sub-ice cavity, controls on basal melt rates remain speculative. Standard oceanographic CTD profiles need to be measured in the fjord to constrain models for water circulation and sub-ice melting.



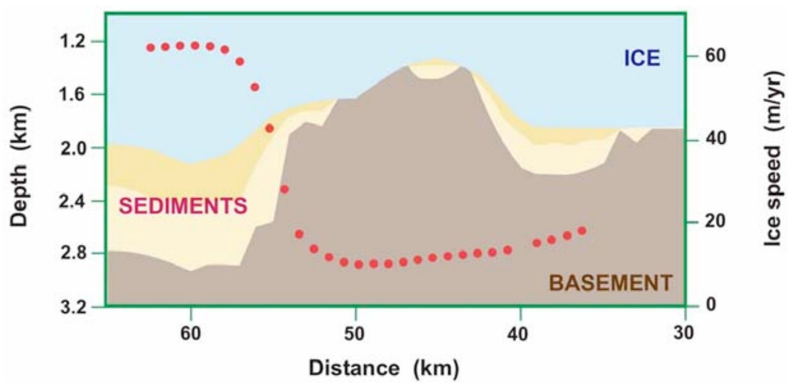
Box 13. Schematized cross section of a coupled ice shelf–ocean system. Shown are the main components, some basic influence quantities, and selected dynamical processes, governing the system evolution under the climatic boundary conditions. M: basal melting; F: basal freezing; ISW: ice-shelf water; HSSW: high-salinity shelf water.

After: Grosfeld and Sandhäger (2004).

4.4. Basal boundary conditions

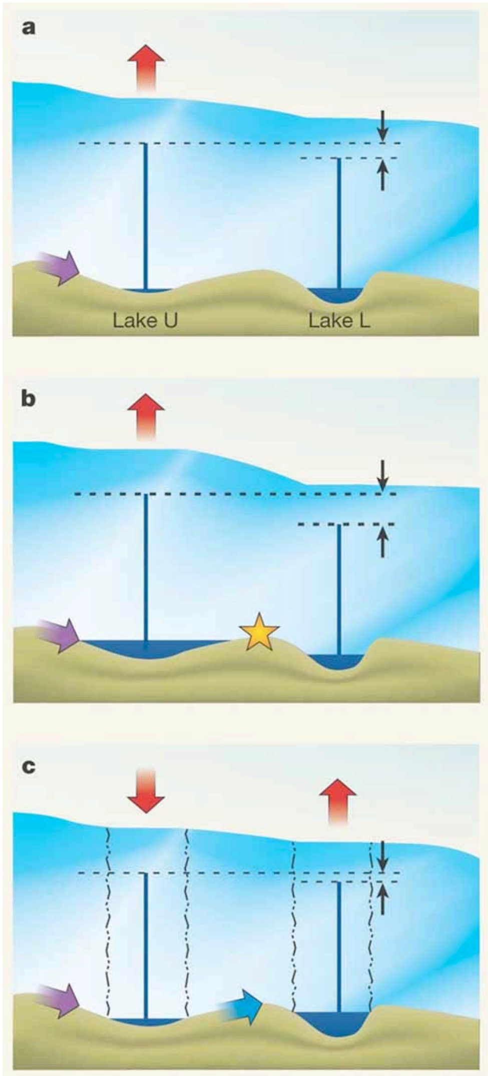
An important control on fast glacier flow is the basal temperature regime and the presence of lubricating meltwater or water-saturated sediments. Observations suggest localized melting in northern Greenland that can only be explained if large spatial variations in geothermal heat flux are invoked. Such variations may be associated with pronounced subglacial topography, or they may result from lateral variations in crustal thickness (Braun and others, 2007). Clearly, with the exception of incidental boreholes, direct access to the bed under the ice sheets is limited and inferences about subglacial geology will have to be made based on surface-based seismic studies, or airborne or spaceborne geophysical observations, for example using the methodologies recommended by Bell and others (1998) and Studinger and others (2001).

Surface-based seismic observations first led to the important inference that one of the active West Antarctic ice streams was underlain by a saturated porous layer believed to be actively deforming (Blankenship and others, 1986; Alley and others, 1986). A seismic profile across the lateral shear margin of a tributary in the onset region of Kamb Ice Stream provides evidence that this margin directly overlies the boundary of a sedimentary basin. The ice stream is within this basin, while outside the basin the ice moves at slow speeds (Anandakrishnan and others, 1998), suggesting that sedimentary fill offers much reduced frictional resistance compared to the subglacial material outside he basin, thereby allowing the glacier to reach greater speeds. An inversion of a depth-averaged ice-stream model reveals that this distribution of till is widespread (Joughin and others, 2004b). Furthermore, thermal modeling constrained by velocity data indicate strong



Box 14. Sediment layer thickness under a tributary in the onset region of Kamb Ice Stream. Layer thickness is determined by seismic-refraction processing and depends on assumed shear-wave velocity. Two solution are shown. Ice-flow velocities along the line are shown with red dots, with the corresponding scale on the right.

From: Anandakrishnan and others (1998).



Box 15.

Upper panel: Lakes U (upper) and L (lower) are isolated from each other and have differing hydraulic potentials (indicated by the water levels in the hypothetical manometers). Meltwater flows into lake U (purple arrow), raising its level and that of the ice floating on the lake (red arrow).

Middle panel: The level of lake U and the overlying ice column continues to rise, increasing the hydraulic potential difference between lakes U and L (black arrows) until a drainage pathway that connects the lakes develops (star).

Lower panel: Water transfer from lake U to L (blue arrow) causes the level of lake U to fall and that of L to rise. The motion of the ice columns overlying the two lakes can be monitored from space7, and can be likened to that of a pair of ice pistons moving through an ice cylinder. Ice deformation at the contact between the piston and cylinder resists this motion, possibly attenuating the flood.

From: Clarke (2006).

freeze-on beneath Whillans Ice Stream (Joughin and others, 2004c), which may explain deceleration over the last three decades (Joughin and others, 2005). Recent satellite observations also suggest water moves beneath the ice streams through a series of discharge events through a “cascading” network of subglacial lakes (Gray and others, 2005; Fricker and others, 2007), which suggests the need for new models for subglacial water flow.

Further characterization of the nature of the ice-bed interface can be inferred from airborne radar sounding. Oswald and Gogineni (subm.) developed a method to map the extent of subglacial water by recovering basal reflection characteristics that yield robust discrimination of subglacial water. Potentially, this method could be applied to the Greenland and Antarctic ice sheets to yield maps of basal water that can be used as important validations for numerical models. Additionally, high-resolution radar mapping across ice stream shear margins could provide independent corroboration of the suggestion that meltwater production is greatest just outboard of the shear margins (Van der Veen and others, 2007a).

In addition to the physical characterization of the bed under fast-moving glaciers (sedimentary, wet, etc.), constraining better the geothermal heat flux is an important requirement for developing more realistic ice-sheet models. Geothermal heat supplied to the basal ice is an important control on the basal thermal regime, and spatial variations in heat flux may be a determinant in concentrating drainage into fast-moving ice streams and outlet glaciers. Relative to the mantle, the overlying crust has enhanced heat production and reduced thermal conductivity properties. Thus, crustal thickness variations are important factors in controlling the transfer of mantle heat through the crust to the base of the ice sheet. Dahl-Jensen and others (2003b) analyzed time series of 16 broadband seismographs installed in the coastal regions of Greenland as well as a few locations on the ice sheet. By applying receiver-function analysis, the depth to Moho was mapped, revealing the presence of two distinct Proterozoic blocks in central Greenland. In northern Greenland, the crustal thickness is significantly less than in the south, with consequently higher heat flux in the north. Modern terrain models suggest that crustal thickness variations may be considerably more detailed than the generalized seismic characterization of Greenland as two crustal blocks (Braun and others, 2007). Airborne and satellite gravity (GRACE) and magnetic (CHAMP, Ørsted) data offer opportunities to extrapolate geologic features from Greenland’s marginal regions into the ice-covered interior, to obtain more detailed spatial models of inferred geothermal heat flux.

4.5. Numerical challenges

Rapid ice-sheet changes originate in, and spread from restricted regions of fast flow such as ice streams and outlet glaciers. Existing models are based on the shallow-ice approximation (SIA) and do not include longitudinal stresses and the buttressing effects of ice shelves that may restrain ice-stream flow in these key regions. The SIA is appropriate only for slow-moving inland ice where resistance to glacier motion is entirely concentrated at the glacier bed. On the other end of the modeling spectrum is the Morland-MacAyeal (MMF) formulation for ice-shelf spreading in which basal drag is set to zero and ice flow assumed depth-independent. It may be expected that flow of fast-moving ice streams and outlet glaciers falls somewhere in between these two model regimes. Thus, the best way to model the dynamically important ice streams is to solve the full stress equations without *a priori* simplifying assumptions. Doing so on a sufficiently fine grid to resolve the ice streams is too computationally intensive for the current generation of computers and one approach is to follow the climate modelers’ lead and develop variable-resolution models, through use of nested meso-scale models embedded in coarse-grid models or variable-element size models with adaptive regridding if needed.

Traditional models of ice sheets employ a fixed horizontal resolution over the whole domain and so either fail to resolve these features adequately (if they employ a relatively coarse ~20 km resolution) or would require unfeasible computing resources (if they employ a more reasonable ~1 km resolution). A solution to this dilemma is the application of a nested grid in which the whole domain is modeled at a coarse resolution, while areas of supposed importance such as ice streams are modeled at a series of finer resolutions with the remainder of the slow-flowing interior omitted. A wide range of variable resolution techniques are available ranging from simple ones in which the areas to be modeled at finer resolution are predetermined and held fixed, to ones in which the solution algorithm itself determines which areas are to modeled at the finer resolution. Similarly, the way information passes between the various grids can vary in sophistication from the coarse grid providing boundary conditions to the finer grid, to full multi-grid techniques in which information flows both ways in an iterative fashion. This type of approach is well developed in ocean and atmospheric modeling, and a number of software libraries exist to facilitate the use of nested grids.

To place any confidence in model predictions, it is first necessary to demonstrate that the models can successfully reproduce past glacier variations. Consequently, it is imperative that data sets be developed against which the skill of the numerical models can be tested. In this respect, it is important to separate data used for model calibration (i.e. parameter adjustment) from those used to evaluate the model performance.

5. CONCLUDING REMARKS

The recently published IPCC Fourth Assessment Report has highlighted the need for more realistic prognostic ice-sheet models that better capture non-linear behavior of the Greenland and Antarctic ice sheets. The current generation of models is based on the shallow-ice-approximation in which the gravitational driving stress is balanced primarily by drag at the glacier base. Such models produce notoriously sluggish ice sheets that respond slowly to imposed environmental forcings. For time scales of interest to IPCC projections, these models are inadequate because of their inability to reproduce observed rapid changes.

Developing more realistic ice-sheet models will require a concerted effort from the community, involving data collection and interpretation, improved process understanding and incorporation into numerical models. This paper provides some suggestions for such an effort.

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