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Rapid response of modern day ice sheets to external forcing

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Abstract

The great ice sheets covering Antarctica and Greenland were, traditionally, believed to take thousands of years to respond to external forcing. Recent observations suggest, however, that major changes in the dynamics of parts of the ice sheets are taking place over timescales of years. These changes were not predicted by numerical models, and the underlying cause(s) remains uncertain. It has been suggested that regional oceanic and/or atmospheric warming are responsible but separating the influence and importance of these two forcings has not been possible. In most cases, the role of atmospheric versus oceanic control remains uncertain. Here, we review the observations of rapid change and discuss the possible mechanisms, in the light of advances in numerical modelling and our understanding of the processes that may be responsible.
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1. Introduction/background

The Greenland and Antarctic Ice Sheets (shown in Fig. 1, with key regions noted) contain about 80% of the Earth's freshwater, and cover 10% of the planet's land surface. If they were to melt completely they would raise global sea level by some 70 m, so even a small imbalance may be important. In simulations of future climate, such as those summarized for the third IPCC assessment exercise [1], the primary impact of increasing temperatures on the ice sheets is growth over the next century as increases in melt along coastal Greenland are exceeded by combined increases in snowfall in Antarctica and Greenland. Changes in the dynamics of the ice sheets have been, in general, considered to take place over much longer timescales than the length of typical transient climate simulations (100–200 yr; [1]). Recent evidence from observational and modelling studies of both Antarctica and Greenland suggests that this assumption may be invalid. The evidence suggests that profound changes in flow and, hence, mass balance are possible over timescales of a few years to decades.

1.1. Timescales and historical context

The greatest changes in the size, and even existence, of continental-scale ice sheets take place over glacial–interglacial timescales of tens of thousands of years. During the last glacial (between about 115 and 12 kyr BP) ice sheets covered North America, Eurasia and Scandinavia, and the Antarctic and Greenland ice sheets

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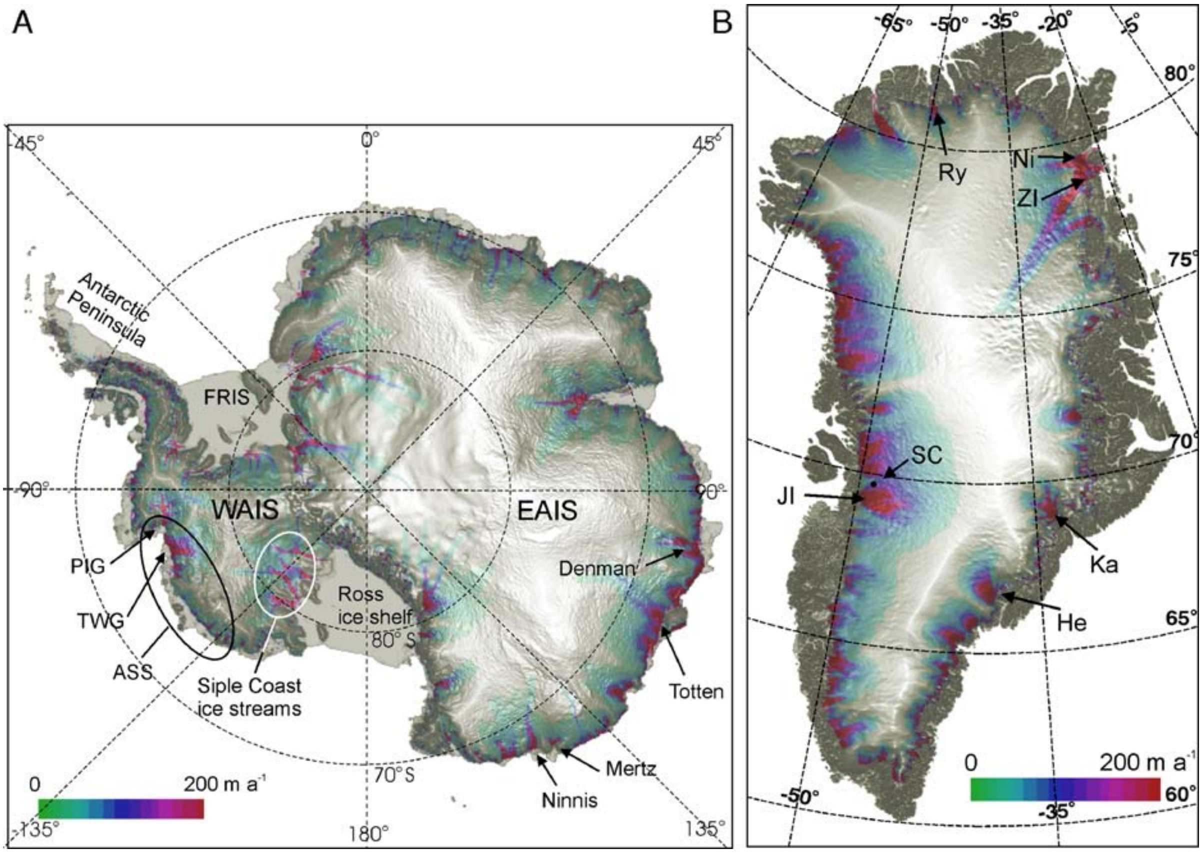

Fig. 1. a. Surface topography and steady-state flow rates for the Antarctic ice sheet. Key locations and features, discussed in the text are labelled. PIG =Pine Island Glacier; TWG = Thwaites Glacier; ASS = Amundsen Sea Sector; FRIS = Filchner Ronne Ice Shelf. b. As for a but for Greenland. JI = Jakobshavn Isbrae; SC = Swiss Camp; Ry = Ryder Gletscher; Ni = Nioghalvfjerdsbræ; ZI = Zachariae Isstrøm; Ka = Kangerdlugssuaq Gletscher; He = Helheim Gletscher.

were notably larger, often extending to the edge of the continental shelf. These changes were the result of major shifts in the climate system affecting the whole planet, and took place over thousands of years. The ice-age ice sheet in Hudson Bay produced a series of apparently internally generated rapid-discharge events [2], and similar ice-age events have been suggested from the West Antarctic Ice Sheet (WAIS) [3]. Here, we focus instead on externally-forced non-cyclic instabilities.

It is important to consider the timescales over which such instabilities might operate and what we mean by “rapid”. To do this, we first consider the steady response time, T_r, of an ice mass to a perturbation in one or more forcing fields. T_r can be defined as how long it takes to reach a new equilibrium state after a small, instantaneous perturbation to one or more boundary conditions. T_r is a function of the ice thickness and rate of mass turnover. For an ice-deformation response away from enhanced flow features (defined below), inland Antarctica has a value of order of 10,000 yr or more. For example, Alley and Whillans [3] found ~8000 yr for response to sea-level rise, ~10,000 yr for response to accumulation-rate change, and much longer times for ice-flow response to temperature rise. The Antarctic Ice Sheet is, therefore, amongst the slowest responding components of the climate system. Part of present-day ice-sheet behaviour, as a consequence, may be due to climate changes at the end of the last glacial, around 12 kyr BP. The Greenland Ice Sheet’s (GrIS)’s mass turnover is faster than for interior Antarctica, yielding a mean response time on the order of a few thousand years. These timescales relate, however, to the parts of the ice sheet system dominated by slow sheet flow. Locally and regionally, parts of the system can respond more rapidly. Perturbation theory suggests that the local response time is inversely proportional to the local velocity [4]. With enhanced flow features extending far inland [5], response times may be much shorter. It is important, therefore, to differentiate in the discussion that follows between local-regional effects and how these may propagate inland to have an impact on the large reservoir of slow-moving inland ice (see Jargon Box).

The primary role of this paper is to present and synthesise the various observational studies on rapid (years-decades) changes in ice-sheet dynamics and to attempt to explain these changes with the aid of results from recent modelling work and theory. To achieve this, it necessary to introduce the reader to some concepts and definitions associated with ice sheet dynamics, (see Jargon Box).

Jargon box

Ice motion is the result of three mechanical processes: internal deformation of the ice (also known as *creep*), sliding of the ice over its substrate (rock or unconsolidated sediment), and shear within any underlying deformable sediment. One or more of these processes may occur at any particular place across an ice mass. Rapid motion by either of the latter two processes, however, requires the presence of water at the bed and, in the case of the third process, a substantial layer of water-saturated sediment. When present, basal motion can be responsible for 90% or more of the total ice velocity. As a consequence, understanding the factors that influence basal motion is crucial to our understanding of ice sheet dynamics and our ability to adequately model ice flow. Small changes in basal boundary conditions can provide a switch between a sliding/no sliding condition and, as a consequence, significantly impact ice motion [6].

Fast-flow features (often taken to mean velocities >100 m a^{-1}) drain much of the ice sheet. They are usually broadly divided into two categories: *ice streams* and *outlet glaciers*. Ice streams are considered to be fast flow features bounded by ice while outlet glaciers are bounded by rock (e.g., fjords). Although different distinctions can be made, more recent work tends to separate ice streams from outlet glaciers more by their dynamics rather than by their morphological setting. The gravitational driving stress, τ_d, for ice streams is typically lower than for the surrounding ice. Streaming flow is generally controlled more by the presence of weak deformable till rather than bedrock topography. In contrast, most outlet glaciers flow through deeply eroded troughs (e.g. Fig. 6). The thicker ice in these troughs generally produces a larger layer of warmer basal ice (and therefore greater deformation and basal melt) and a higher driving stress than for the surrounding regions. Outlet glaciers are also more likely to flow over hard crystalline bedrock, and to have creep supplying an important fraction of the total velocity. Early work suggested an abrupt transition between "inland flow" and "streaming flow", but more recent studies have revealed a complex network of *enhanced flow tributaries*, which penetrate several hundred kilometers into the ice sheet interior in both West and East Antarctica [5,7]. Typical tributary velocities are 25–100 m a^{-1}, and are likely due to a combination of internal deformation and basal motion [8]. The gravitational driving stresses are typically higher than for regions of streaming flow, where basal motion dominates, and the velocities are intermediate between inland and streaming flow values [9]. For many outlet glaciers in Greenland, there is little tributary flow and rapid flow begins abruptly in highly convergent regions near the coast, likely at the heads of subglacial extensions of the fjords through which they drain [10] (Figs. 1b and 6).

The next section discusses the likely mechanisms that could, theoretically, be responsible for a rapid change in flow regime. Following this, we present a suite of recent results, primarily from satellite remote sensing observations, but also from a small number of numerical modelling studies of rapid and significant changes in flow regime for both Antarctica and Greenland. Finally, we discuss the implications of these studies for our understanding of the response of the ice sheets to external forcing, our ability to predict their future behaviour, and the limitations of our present knowledge.

2. Mechanisms

The possibility that the West Antarctic Ice Sheet (WAIS) could react rapidly to external forcing was first proposed in the late 1960s [11]. This work suggested that ice shelves (Fig. 1a) act as buttresses to ice streams, regulating their discharge as a consequence. Ice shelf removal, therefore, might yield ice-stream acceleration ultimately leading to disintegration of the WAIS [12,13].

This hypothesis has undergone revision and re-evaluation since its first introduction and is still an open debate. We discuss this and other mechanisms that can induce a rapid response next.

2.1. Melt water drainage

The effect of meltwater drainage to the bed of temperate alpine glaciers on ice motion is a well established phenomenon [14,15], and has even been associated with velocity fluctuations of a predominantly cold-based glacier in the Arctic [16]. Drainage through a substantial thickness (~1 km) of cold ice within an ice sheet has received less attention until recently (e.g., [17,18]).

Debris-free meltwater from the Greenland Ice Sheet's surface drains into holes, called moulins. In many places, debris-rich streams emerge from beneath the ice onto land or into the ocean, and simple energy-budget calculations indicate that the surface melt must dominate the water fluxes. As the only debris source is at or near the bed, it seems likely that surface meltwater reaches the bed.

This inference is supported by observations (currently available from only one site: Swiss Camp, Fig. 1b) that the rate of ice motion increases when surface melting begins in springtime and decreases with the autumnal end of surface melting [19]. Years with more melting show greater seasonal speedup. The changes are not large (between 10 and 25%) and are relatively short-lived, but significantly exceed measurement errors. In common with behaviour of many mountain glaciers, the ice exhibits a late-winter nearly steady velocity. A reasonable interpretation is that springtime acceleration occurs as meltwater is supplied to the bed more rapidly than removal in well-developed subglacial streams, allowing the water to spread across and lubricate the glacier bed. Enhanced water-flow paths then develop during the summer, increasing water drainage and reducing lubrication especially when the meltwater input slows in the autumn. Fig. 2 shows the calving front of a south Greenland glacier. The blue bands and darker patches indicate the location of channels and cavities in the ice caused by englacial meltwater flow (now refrozen). The base of the ice is below the sea level but it is clear that at least in the upper ice column water routing can be widespread.

The mean summertime velocity at Swiss Camp, Greenland (Fig. 1b) slightly exceeds (by ~2%) the mean wintertime velocity, so the meltwater provides at least some enhancement to the total ice motion over the year [19]. In assessing the importance of this mechanism to the ice sheet's future mass balance in a warming world, one of the main uncertainties is the potential for surface meltwater to reach the bed farther inland [20]. Should abundant surface meltwater reach a frozen bed, rapid warming to the melting point and thawing are expected, accelerating flow. Drainage of surface meltwater to the bed likely occurs through propagation of water-filled fractures (crevasses) followed by collapse of water flow to one or a few moulins [14,18,21]. The propagation of cracks through the full ice sheet thickness probably requires a large volume of water such as that provided by the many surficial lakes in the ice sheet's ablation zone that form during the summer ablation season [21].

Should warming allow the inland migration of the zone in which meltwater lakes form on the surface of the ice sheet, and should ice-flow stresses be large enough to open crevasses in the vicinity of those new lakes, then thawing and enhanced lubrication of the bed in those regions will be likely. Even in the present climate, large "slush swamps" form in closed basins in the upper percolation zone, which would easily transition to lakes with increased melt. The total speed-up of flow will depend on the conditions produced by basal thawing—if thick, soft, smooth subglacial tills are present there, order-of-magnitude changes could be possible, but in the more-likely event of bumpy bedrock, factor-of-two or smaller changes seem more likely [20].

2.1.1. Glacier surges

Many glaciers around the world, including several in Greenland [22,23], are known to undergo a quasi-cyclic, non steady-state behaviour known as surging. Typically, ice in the downstream (near-coastal) part of the glacier advances suddenly several kilometres in a few months to a year and then slows or stagnates, allowing the

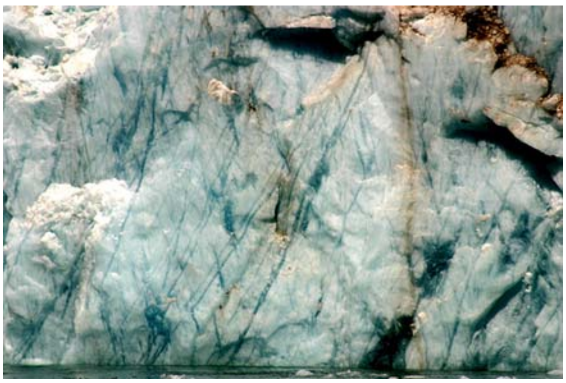

Fig. 2. The calving front of a south Greenland outlet glacier, Kangersuneq qingordleq, showing an extensive network of vertically orientated relict meltwater pathways and cavities.

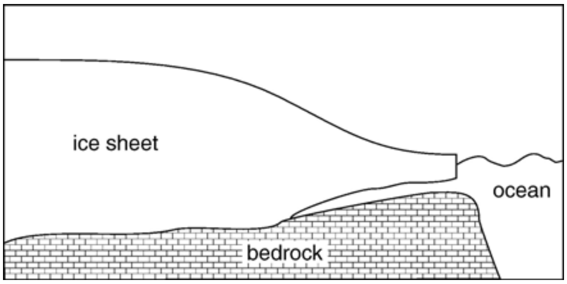

Fig. 3. Schematic cross-section of ice sheet grounding zone with a backward sloping bed profile.

margin to retreat over a period of tens to a few hundred years during what is often termed the quiescent phase of a surge cycle [24,25]. During this phase the upstream part of the glacier gains mass. The cycle of build-up and rapid advance varies in duration from tens of years to a few centuries. It is possible that individual ice streams can surge [26], but if different parts of an ice sheet were capable of surging, they would do so with different cycle durations because of differences in physical conditions. The mechanisms for a rapid response discussed here, however, are not directly associated with the classic surge-cycle behaviour described above. Both may be due to changes in basal hydrology but surges are generally considered to be internally-driven, cyclic instabilities in ice dynamics whereas the focus here is on externally-driven changes.

2.2. Ice sheet–ocean interactions

An ice sheet with its bed below the surface of an adjacent sea or lake—a marine ice sheet—has more mechanisms of rapid change available than does an ice sheet ending on land. If the ice floats at the edge but does not immediately break off to form icebergs, then a floating ice shelf forms attached to the ice sheet (Fig. 3). Water circulation generally delivers orders of magnitude more heat to the undersides of ice shelves than the Earth delivers to grounded (non-floating) ice. In addition, icebergs that break off from ice shelves, or from grounded tidewater ice fronts if no ice shelf forms, can float away and melt elsewhere, so an ice sheet calving icebergs can shrink without requiring heat delivery from the environment.

Modern and past ice sheets included marine margins. The WAIS is largely marine, resting on a bed that is almost everywhere well below sea level and that deepens toward the ice sheet's center in some places (Fig. 4).

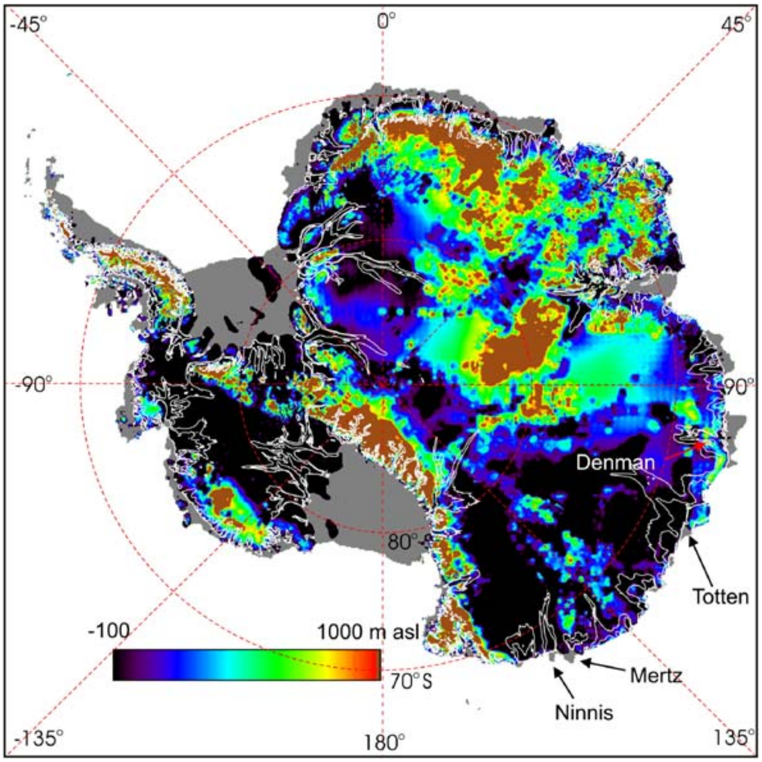

Fig. 4. Bedrock topography for Antarctica highlighting areas below sea level (in black), fringing ice shelves (in dark grey) and indicating areas of enhanced flow with contours of estimated steady-state velocities, known as balance velocities, in white.

Many glaciologists have argued that this may allow especially rapid response to changes in climate, or even internally generated instabilities [13,27,28].

Weertman [28] argued that, in the absence of ice-shelf buttressing as discussed below, a marine ice-sheet margin on a bed that deepens towards the center of the ice sheet is inherently unstable. Consider the situation in Fig. 3. An ice shelf's extensional thinning rate increases with thickness. Thus, if the grounding line (which separates floating from grounded ice) advances seaward for some reason, then its thickness decreases, reducing the ice shelf's creep thinning, thereby promoting further grounding line advance. Alternatively, a process that initiates grounding line retreat toward the ice sheet's center yields thicker ice at the grounding line and additional creep thinning, causing the grounding line additional retreat. Thus, a positive feedback exists that amplifies initially small grounding line perturbations. In the case where the bed is higher beneath the ice than beneath adjacent water, the feedback is negative and the position of the grounding line remains more stable. Subsequent work by van der Veen [29,30] identified feedbacks omitted from the earlier work that would tend to promote stability but did not directly address the Weertman conjecture [28], and more recent work indicates that the Weertman instability does exist in the idealized geometry of Fig. 3 [31].

As van der Veen noted [29], WAIS is there, and has persisted for at least tens of thousands of years and probably longer, so a simple and rapid instability model cannot be correct. Many processes are missing from that simple model, including geometric differences (much of the ice-sheet bed does not fit the simple cartoon in Fig. 3, although some does). As a stabilizing mechanism, ice-shelf buttressing has received the most discussion, so we consider this next.

2.3. Ice shelf backstress/buttressing effect

The ice sheets spread and thin under their own weight, but the rate of spreading is reduced if a resistive push, or backstress, is applied from the edges (or from anywhere else) to oppose the spreading. Dramatic evidence of this effect is given by the response of fast-moving weak-bedded ice streams to tidal fluctuations. The extra backstress from ~1 m rise in tide is enough to slow one ice stream almost twofold, and to completely stop another, with motion occurring only during short-lived events on the falling tide [32,33].

The backstress provided by ice shelf buttressing has long been deemed important in overcoming the marine ice sheet instability (e.g., [11,34]). Ice shelves rarely spread freely, as assumed in the Weertman model [28]. Instead, they form in embayments or run aground on islands, and the friction with embayment sides and with islands resists flow. Building on earlier work (e.g., [3]) and using faster computers, a new generation of ice-flow models is being used to solve for higher-order stress terms (including, for example, longitudinal stresses) with more-realistic geometry [35].[1] No fully three-dimensional, thermomechanical, full-stress tensor whole-ice-sheet model has yet been used for long simulations, so much work remains to be done, but progress is rapid, and insights from simplified-geometry, basin-scale or short-interval modelling are proving quite instructive [36].

Where an ice shelf is fed by a fast-moving ice stream with a well-lubricated bed, a reduction (or increase) in ice-shelf buttressing can speed (or slow) flow inland as far as ~100 km almost instantaneously [32]. Faster flow in the ice stream advects thicker ice from upglacier, and thinning in downglacier regions steepens the ice surface, increasing the gravitational driving stress and speeding flow to produce a "wave" of thinning that diffuses upglacier. This advective-diffusive response is typically dominated by the diffusive term [37]. Perturbations can travel hundreds of kilometres in a fast-moving ice stream in decades, but may take millennia to notably affect central, slow-moving regions of an ice sheet [37–40]. The modelled magnitude of the velocity change in response to a change in ice-shelf buttressing depends on a host of variables, and may range from trivial to large. Plausible forcings can produce factor-of-two velocity responses, and order-of-magnitude velocity responses are possible.

One example is given in Fig. 5, from Dupont and Alley [39]. A steady, idealized version of Pine Island Glacier (Figs. 1a and 4) was generated in which an ice shelf offset 50% of the ice spreading tendency at the grounding line. This ice-shelf buttressing was then permanently removed while the thickness was held constant at the upglacier end of the ice stream where it meets the inland ice sheet. The modelled ice velocity initially increased about 50% at the grounding line, although some of this perturbation was lost as ice thinning reduced the driving stress. Enough ice was transferred to the ocean from the ice stream to raise sea level ~1 mm in ~40 yr. Allowing the perturbation to propagate into the inland ice would have increased the potential contribution to sea-level rise; allowing the ice-shelf to re-form in its prior position would have tended

[1] For a comprehensive review of advances and concepts in numerical modelling of ice sheets see 36 S.J. Marshall, Recent advances in understanding ice sheet dynamics, Earth and Planetary Science Letters 240 (2) (2005) 191–204.

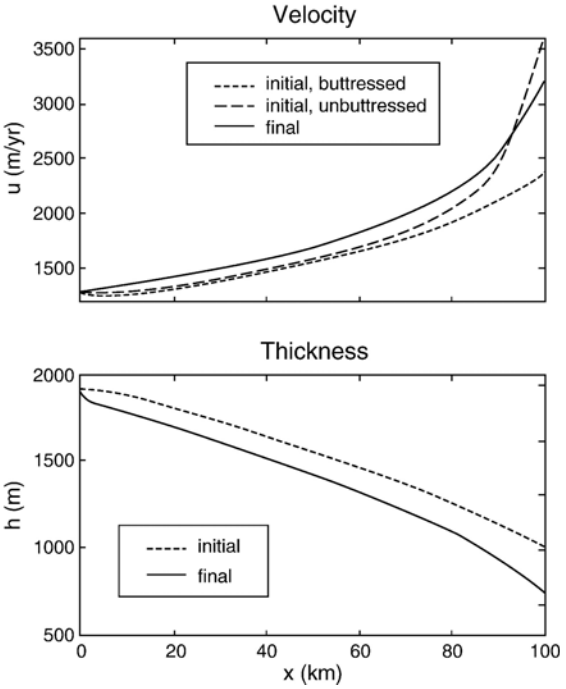

Fig. 5. Effect of ice shelf buttressing on velocity of grounded ice. Removal of ice-shelf buttressing causes ice-flow speed-up and thinning, as described in the text [32].

to reduce the changes and the contribution to sea-level rise, whereas allowing additional forcing to remove buttressing that developed from new ice shelf that developed upglacier of the original grounding line would have increased changes. Among the many uncertainties in this modelling (including lack of complete knowledge of basal topography and the distribution of lubricating water and soft sediment under the ice), poor knowledge of the ice-ocean interactions and the ability of an ice shelf to re-form or a recently floated region to re-ground may pose the biggest uncertainties.

3. Observations of rapid changes

3.1. Greenland

The Greenland Ice Sheet loses mass by three mechanisms: surface melting or ablation, bottom melting under floating tongues, and iceberg calving. Ablation accounts for about half the direct mass loss [41,42]. Ablation also may affect ice flow and thus discharge across the grounding line. Recent observations show significant changes in outlet glacier discharge and seasonal variability on inland ice.

3.1.1. Inland ice

As noted above, observations of surface motion (from *in-situ* GPS measurements) taken at Swiss Camp in southwest Greenland (Fig. 1b), situated close to the equilibrium line suggest that the motion is influenced by the amount of melting taking place. Not only is there a springtime speed-up of up to 25% (partially offset by the autumnal slowdown) but there is also a vertical uplift of around 50 cm suggesting a build up of subglacial water during the summer melt period [19]. As noted earlier, a key requirement appears to be the presence of lakes at the surface, which have been observed to drain rapidly (e.g. [43]) throughout the region between the margin and the equilibrium line or above. While it is difficult to attribute drainage of the lakes to many of the recent outlet glacier speedups as described below, a short-lived speed-up of a major outlet glacier in northern Greenland (Ryder Gletscher; [44]) was observed from repeat satellite measurements of surface velocity. Here, a roughly three-fold increase in velocity occurred over a 7-week period during a period when lakes on its surface appear to have drained.

3.1.2. Outlet glaciers

It is estimated that the increase in speed of glaciers south of 70°N has more than doubled the net mass imbalance from the ice sheet between 1996 and 2005 [45]. Enhanced surface melting may be partly responsible for the remarkable speed-up (by up to a factor of two) [45–49]. Most of the speedups on outlet glaciers are sustained through the winter, with Ryder Gletscher being the one exception as mentioned above. The glaciers that have sped up typically move at speeds of 1 to 10 km/yr, suggesting a well-lubricated bed even in the absence of active surface-meltwater input. Furthermore, while observations are sparse, there is little evidence of seasonal variability in speed on these glaciers to suggest a sensitivity to surface melt water that could cause accelerations of 50% and greater. Thus, it is likely that causes other than an increased supply of meltwater to the bed are more important.

The speed increase on Greenland's largest outlet glacier, Jakobshavn Isbrae, has been attributed to the loss of buttressing as its floating ice tongue thinned and then disintegrated [47,50]. Many of the other outlet glaciers that accelerated over the last decade are tidewater glaciers with grounded termini with little or no floating ice tongues [45]. The loss of significant portions of grounded ice, however, can provide a greater force imbalance since, unlike the case with floating ice, there is a loss of basal resistance. For example, Helheim's grounded trunk retreated by ~7 km from 2003 to 2005, during which time the remaining ice sped up by ~40% [48]. Thus, loss of buttressing as grounded and floating ice fronts have retreated may be the cause for much of the acceleration. If ice-front retreat caused the acceleration, the reasons for this retreat are not yet clear. Calving rates on Jakobshavn Isbrae are up to six times higher in summer, suggesting a sensitivity to temperature and/or sea ice cover [51]. Warmer temperatures might yield more ponded meltwater in crevasses, causing hydro-fracturing similar to that which may have caused the breakup of the Larsen ice shelves along the Antarctic Peninsula [52]. Floating ice, and non-floating ice near the grounding line, are often in tension and near fracture; hence, only small amounts of water in crevasses may hasten calving events. Alternatively, warmer water in fjords during the summer may contribute to greater calving [53]. Changes in sea ice extent may also have an impact on both sub-shelf circulation [54] and wintertime calving rates. It has been suggested that this may partly be responsible for the seasonal differences in calving rate for Jakobshaven Isbrae, and is another potential feedback between ice dynamics and external forcing [51].

It is also likely that the process of retreat may involve important positive feedbacks. For example, acceleration may lead to more extensional flow and greater crevassing, which may lead to greater calving and further retreat and acceleration. Regardless of the exact mechanism, the retreat and acceleration occurred over a period of warmer summer temperatures [55]. During this period, both the incidence and seasonality of glacial earthquakes detected from teleseismic observations increased [56,57]. These short-lived, 30–60 s, seismic events are associated with sporadic glacier activity and may be related to calving events or grounded ice dynamics [55]. For Helheim and Kangerdlugssuaq Glaciers, both of which have accelerated markedly recently, it has been suggested that thinning may have led to flotation of lightly grounded and highly crevassed ice, which almost immediately disintegrated to form icebergs [49].

3.2. Antarctica

In Antarctica, surface melting or ablation is negligible (except in the Antarctic Peninsula, see below) and mass loss is dominated by iceberg calving and sub-shelf melting. It seems unlikely, therefore, that surface melt could have an influence on ice dynamics to the degree that it does for Greenland. No seasonal changes in ice velocity have been observed in Antarctica, unlike Greenland. Significant, roughly coincident, increases in ice motion have been noted in several different sectors of Antarctica, however, suggesting a common external forcing.

3.2.1. Peninsula

The Antarctic Peninsula (AP) is the warmest and most northerly region in Antarctica. Summer surface melt has been observed over some of the AP ice shelves, and the strong warming in the AP over the last 50 yr probably was responsible for the disintegration of the northern Larsen Ice Shelf [58–60]. Oceanic warming may also have contributed to the ice shelf thinning and collapse [61]. Of relevance here are observations of rapid increases in speed (by as much as a factor eight) of glaciers after the buttressing ice shelf was removed [62,63]. The AP glaciers are relatively small by Antarctic standards (a few tens of kilometres in length at most) and drain a high, narrow mountain range, so direct analogy to the rest of Antarctica may not be appropriate.

3.2.2. Amundsen Sea sector

Interestingly, three adjacent glaciers in dynamically independent basins in the Amundsen Sea sector (Figs. 1a and 4) of the WAIS have sped up [64,65], suggesting a response to a common forcing. Numerical modelling and satellite-observations point towards ocean warming as this forcing [37,66,67]. The three glaciers (Pine Island, Thwaites and Smith) all have floating ice shelves that have progressively thinned over at least a ten-year period [66].

Some of the most dramatic Amundsen Coast changes have taken place on Pine Island Glacier. This glacier has accelerated by more than 25% over the period from 1974–2003 [68,69], and satellite altimetry shows strong thinning on the floating ice ($\sim$4 m/yr) and extending well inland at lower rates ($\sim$10 cm/yr) [70]. Steady-state basal melt rates of Pine Island Glacier's floating ice tongue average about 11 m/yr, so thinning of $\sim$4 m/yr represents an increase of just over one third of the steady state rate, explainable by ocean warming of a few tenths of a degree C [66].

There are two ways in which the strong thinning near the grounding line may have caused the acceleration. The first may be a reduction in back-stress from a thinner ice shelf. Alternatively, loss of basal resistance as ice ungrounds in response to strong thinning could yield a speedup. There was at least one instance where the grounding line retreated by several kilometres just prior to a period of acceleration [71]. Either, or a combination of these causes, would produce strong longitudinal stress gradients near the grounding line. Higher-order numerical modelling (including longitudinal stresses) shows that the observed response is consistent with the inferred forcing [35,37].

The rapidity of the response has two implications. First, the sensitivity to temperature is large enough that a sustained warming of the ocean likely would lead to much larger imbalances along the Amundsen Coast with a consequent impact on sea-level rise [72]. Second, the ongoing changes may be a response to annual-to-multi-decadal scale variability in temperature beneath Pine Island Glacier's floating ice shelf rather than to climate warming or some other slower forcing.

3.2.3. WAIS

Changes in the small Amundsen Sea (AS) ice shelves of WAIS are linked to large ice-flow changes, whereas smaller ice-flow changes including thickening are associated with the much larger Ross and Filchner–Ronne Ice Shelves of WAIS (which together represent ~60% of the total area of Antarctic ice shelves) [73,74]. Whereas circumpolar deep water can circulate vigorously beneath the Amundsen Sea Coast ice shelves, causing basal melt rates of up to 40 m/yr and potentially much larger if warming affects the water temperature, such vigorous ventilation is not observed beneath the deeply embayed larger ice shelves [75]. Instead, slower ventilation involves dense, cold water formed during sea ice production that sinks beneath the large ice shelves to the grounding lines, where higher pressure gives a slightly lower melting point. This allows the descending water, which is at the surface melting point, to produce some sub ice-shelf melt (a few m/yr at grounding lines and average rates of 10–20 cm/yr) [72,75,76]. Although basal melt may have driven the retreat that formed these ice shelves, as they became more deeply embayed and less well ventilated, a decline in melt may have halted their retreat. Melt rates beneath the deeply embayed Ross and Filchner–Ronne ice shelves might even decrease in a warming climate due to a reduction in the sea ice production that drives much of the sub-shelf circulation [54].

Whole ice-sheet numerical models have attempted to incorporate the effects of ice-shelf basal melting, but in a limited way [77,78]. They do not include the effect of propagation of longitudinal stresses inland, and severely underestimate the sensitivity of basal melt to ocean temperature [72]. As a consequence, none of these large-scale modelling studies predicted either the magnitude or rate of response recently observed in the Amundsen Sea sector. One of the most extensive modelling studies of future ice sheet behaviour suggests that the whole ice sheet will have a positive mass balance during the third millennium for temperature increases up to 5.5 °C [79], without the incorporation of enhanced basal melt. With the addition of 5 ma^{-1} basal melt, distributed uniformly beneath the ice shelves, the mass balance was negative, with a contribution of 168 cm to sea level rise by the year 3000. These models distribute melt evenly over the base of the ice shelf. Observations indicate that melt rates are the greatest at the grounding lines of fast moving outlet glaciers and ice streams [72], so that sensitivity to basal melt may be higher than models indicate.

3.2.4. EAIS

A number of East Antarctic glaciers in the sector from ~100–160°E appear to be thinning near their grounding lines. The magnitude is less than for the AS sector, but it is interesting to note that these glaciers (Ninnis, Mertz, Totten and Denman) all have floating tongues or ice shelves susceptible to increasing ocean temperature (Fig. 4). A multidecadal warming trend in this part of the Southern Ocean has been detected from ship-borne measurements [80]. Another common feature of these glaciers is that they lie on bedrock significantly below sea level at the coast (Fig. 4). Thinning or loss of even quite small ice shelves can have important effects on tributary ice flow [81], and East Antarctica has numerous fringing ice shelves.

4. Discussion

4.1. Greenland

No comprehensive whole-ice-sheet models reproduce the observed behaviour of Greenland's outlet glaciers during the moderate recent oceanic and atmospheric warming. The only significant positive feedback in these models is an increase in the ablation area as temperatures rise, which in turn increases ablation due to a lower albedo. Any dynamic feedback is slow and moderate in magnitude, resulting from changes in ice sheet geometry [79]. It has also been suggested that if the external forcing in Greenland is oceanic then this will have a limited and short-lived impact as most of the ice sheet margin is grounded on bedrock above sea level [82] (Fig. 6). Thus, once the glaciers have retreated inland they are no longer in contact with the ocean. This is true for southern and eastern Greenland but not for several major outlet glaciers in the north including Humboldt, Petermann, Ryder, Nioghalvfjerdsbræ and Zachariae Isstrøm (Fig. 6). If the external forcing is partly atmospheric, as seems likely [19,45,83], then loss of calving fronts may only partially impede the accelerated retreat of the ice sheet. The importance of the surface melt/ice dynamics feedback is an area of current research.

4.2. Antarctica

For the immediate future, any rapid changes in the Antarctic Ice Sheet are most likely to be triggered by inland propagation of the effects of ice-shelf shrinkage or loss, especially in response to increased sub-shelf basal melting, but comprehensive whole-ice-sheet simulations are lacking. Improved simulations will have to tackle a number of challenging issues including ice shelf-sheet coupling and grounding line migration. Of particular note, oceanic heat flux beneath ice shelves remains poorly understood (even such basic information as water depth is missing in many cases), and is not well-modelled in any global-scale simulations. Hence, estimates of forcing cannot be considered reliable, but are critical in assessing future changes.

Rapid inland propagation of a grounding line perturbation is strongly dependent on the strength of basal traction, τ_b. If this is high, and basal sliding is not present, dissipation of the perturbation is high and transmission inland limited. If, however, τ_b is low, in which case longitudinal stresses are a dominant component of the force budget, then inland transmission is efficient, as is the case for Pine Island Glacier [84]. “Flow laws” for the dependence of basal velocity on the applied stress range from linear to highly nonlinear so that doubling the applied stress may double the basal speed (linear), or increase it far more (non linear), based on these “laws”. Geophysical studies of the glacier bed (e.g., [85]), laboratory work, and process modelling are helping constrain models but much remains uncertain in constraining the characteristics of basal motion.

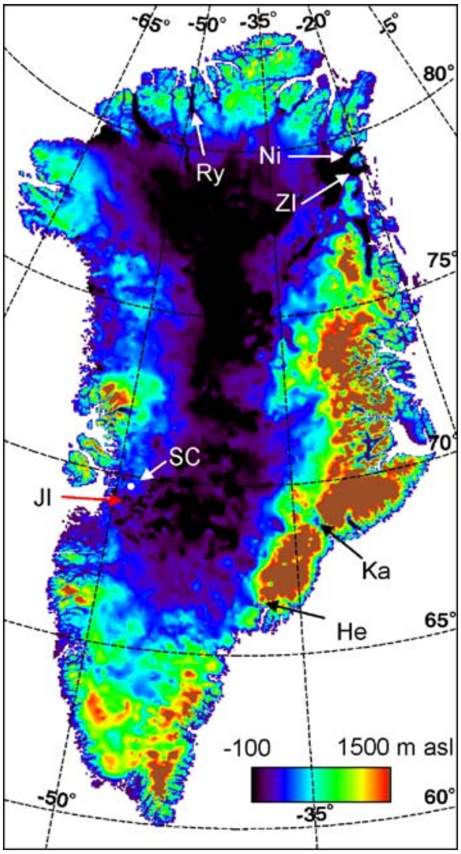

Fig. 6. Bedrock topography for Greenland highlighting areas below sea level (in black) and the deep incised valleys that many of the outlet glaciers referred to in the text flow through. JI = Jakobshavn Isbrae, SC = Swiss Camp, Ni = Nioghalvfjerdsbræ, Ry = Ryder Gletscher, Ka = Kangerdlugssuaq Gletscher, He = Helheim Gletscher, ZI = Zachariae Isstrøm.

Accurate, large-area constraints on basal drag and its relation to flow velocity are obtained from inverse modelling driven by ice-sheet thickness, surface elevation, and velocity as determined by remote sensing or surface measurements. These techniques have been employed to derive basal traction values for the ice streams feeding the Ross and Filchner Ronne ice shelves [86]. The results indicate that the Siple Coast ice streams have uniformly weak beds in their lower reaches, but the tributaries feeding them have extensive patches of strong bed resistance, with the potential to limit the propagation of grounding line perturbations inland. For the ice streams feeding the FRIS, the picture is more complex. Recovery glacier is the only flow feature with an extended area of uniformly weak bed. Elsewhere there are alternating regions of weak and strong bed [9]. The impact of changes at the grounding line are, therefore, not obvious and no higher-order models have been used to simulate the behaviour of these flow features.

5. Conclusions and outlook

Important marginal regions of both the Greenland [45,47,48] and Antarctic [64,66,71] ice sheets have exhibited ice-flow speed-up contributing to sea-level rise in recent years, with warming (either atmospheric, oceanic or both) the likely cause. In the case of Greenland, a recent estimate suggests that this has resulted in an almost threefold increase in its contribution to sea level rise in less than a decade [45]. Results from three independent satellite-based approaches indicate that the WAIS is losing significant mass [74,87,88]. There is uncertainty in the absolute magnitude of losses, but there is growing evidence that mass loss has increased in the last decade and it seems likely that this trend will be maintained if the amplified[2] warming in the polar regions continues [89,90].

Projecting the future of the ice sheets, and whether these speed-ups are soon-to-stabilize perturbations or harbingers of larger future changes, remains very difficult. Although the surface mass balance of ice sheets can be estimated with reasonable accuracy from global

[2] GCMs and observations indicate that the Arctic is warming at a rate up to three times the global mean.

weather forecast re-analyses using downscaling or regional climate models [42,91], similar skill has not been demonstrated for oceanic heat fluxes. Potentially important physical processes remain poorly understood, such as the potential for additional surface meltwater to speed ice flow in regions currently affected, and for the affected regions to spread. Thus, the rapid changes in ice dynamics reported here, for both Greenland and Antarctica, were not predicted by any large-scale numerical models of ice sheet flow, and our ability to simply reproduce the present-day observed velocity fields of the ice sheets remains limited. No ice-flow model incorporating longitudinal stress terms has been used to simulate whole-ice-sheet behaviour over appropriate time intervals, yet it is evident that these stresses are contributing to the changes we have described here. It is clear, therefore, that the current generation of numerical models can provide only limited insights into the future behaviour of the ice sheets. Given this great uncertainty, we are able to confidently state that modern-day ice sheets *can* respond rapidly to external forcing, but if asked "*Will* ice sheets in the near future respond rapidly to external forcing?" we must give a qualified "maybe".

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