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Ice streams—fast, and faster?

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Abstract

Rapid flow of ice streams is caused either by great thickness, or by effective basal lubrication especially from deforming tills. Competing thermal processes act to stabilize and to destabilize the well-lubricated ice streams, and may contribute to their observed short-term variability yet long-term persistence. Increasing evidence indicates that ice streams are subject to speed-up in response to warming, through thinning or loss of ice shelves, and possibly in response to meltwater penetration to ice-stream beds. ***To cite this article: R.B. Alley et al., C. R. Physique 5 (2004).***

Résumé

Fleuves de glace : toujours plus vite ? L'écoulement rapide des fleuves de glace est causé soit par leur grande épaisseur, soit par une lubrification efficace à la base, en particulier due aux sédiments sous-glaciaires déformés. Les différents processus thermiques en compétition peuvent stabiliser ou déstabiliser ces fleuves de glace lubrifiés, ce qui pourrait contribuer à leur variabilité à court terme comme à leur persistance à long terme. Les évidences s'accumulent en faveur d'une accélération des fleuves de glace en réponse au réchauffement climatique global, par un amincissement ou un retrait des plate-formes glaciaires, et peut-être par la pénétration d'eau de fonte vers le lit glaciaire. ***Pour citer cet article : R.B. Alley et al., C. R. Physique 5 (2004).***

Keywords: Ice-stream; Ice-sheet; Antarctica; Greenland; Till; Sea-level; Ice-shelf

Mots-clés : Fleuves de glace ; Calottes polaires ; Antarctique ; Groenland ; Sédiments sous glaciaires ; Niveau des mers ; Plate-forme glaciaire

1. Introduction

An ice stream is a region of rapidly moving land ice with slower-moving ice on either side, e.g., [1]. Ice streams discharge much of the flux from big ice sheets. The possibility of large and rapid changes in this discharge flux, e.g., [2–4] focuses attention on ice streams.

Features identified as 'ice streams' range greatly, from those sitting in deep bedrock troughs and transitional toward rock-walled outlet glaciers to those with scarcely any topographic control, and from those with steep surface slopes only slightly different from surrounding non-streaming ice to those with hardly any surface slope and more similar to ice shelves in many

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ways. In this short review, we use a series of questions and answers to address the behavior of ice streams. The various types of ice streams, and differences between, say, Jakobshavn in Greenland and ice stream B in West Antarctica, are emphasized through these questions and answers. Recent developments are highlighted here; for earlier work, see [1].

2. Why do ice streams flow rapidly? Great depth, or lubrication especially by water-saturated till

Enhanced flow velocity in an ice stream might be generated in any of several ways. These include development of an anomalously slippery bed (thawed *versus* frozen, smooth *versus* bumpy, or with soft and continuous till *versus* absent or discontinuous or stiff till), anomalously soft ice (from higher temperature, or effects related to impurities or c-axis-fabrics), or anomalously high shear stress (from greater thickness or surface slope).

Of these, enhanced surface slope is difficult for an ice sheet to maintain because of relaxation by transverse ice flow or other mechanisms. We know of no documented cases of strong lateral inhomogeneity in ice properties causing an ice stream, and we suspect that this would be difficult to accomplish. Most or all ice streams appear to arise either from increased thickness or enhanced basal lubrication.

2.1. Thick ice

The effect of increased ice thickness is easy to understand. Bed-parallel shear stress increases linearly with thickness of overlying ice, shear strain rate in ice increases with the cube of the shear stress, and integrating this shear-strain rate through thickness yields a fourth-power dependence of surface velocity on thickness, see e.g., [5]. Basal sliding and till deformation, if active, also increase with ice thickness through its effect on shear stress. Sliding over a hard bed may increase roughly with the square of the basal shear stress [6]. (Bed deformation is more difficult to summarize easily, as discussed below.)

As an example of an ice stream in which thickness plays an important role, Jakobshavn Isbrae in Greenland has a maximum center-line thickness in the main channel approximately two and one-half times that of adjacent slower-moving ice [7,8]. By itself, this would yield a 40-fold increase in surface velocity arising from ice deformation compared to the adjacent, slower-moving ice if the channel were wide enough, although side drag offsets this somewhat. The greater thickness in the channel and greater basal shear stress there also cause basal velocity from sliding and any bed deformation to be higher than for equivalent conditions under adjacent ice. Jakobshavn flows rapidly primarily because it has a deep bed.

Clearly, this raises interesting questions about the origin of deep valleys (selective linear erosion [9]), but these questions extend beyond the scope of this review. Briefly, many glacial valleys follow geologic features (faults, etc.) that served to localize erosion, but glaciation can greatly enhance topographic contrast, as shown by erosion of fjords far below sea level. The bedrock under Jakobshavn is incised 1.5 km below sea level, with adjacent rock near sea level, as an example [7]. High basal shear stress and rapid basal ice motion likely increase erosion rates by abrasion and plucking [10–12], so a strong feedback should exist in which the low-elevation bed of a thick ice stream enhances ice velocity and shear stress, increasing erosion rate and thus the tendency to ice streaming.

2.2. Slippery bed

Perhaps of greater interest, as discussed below, are those ice streams that arise from enhanced basal lubrication. Many ice streams seem to have thawed beds but to flow between frozen-bed regions, as in the Northeast Greenland ice stream [13,14] and broad regions of the Siple Coast of West Antarctica [15]. Subfreezing sliding is quite slow [16]; thawing increases basal velocity, and the increase may be quite large. Whether ice streaming is solely a response to a freeze-thaw boundary, or whether additional lubrication is required beyond simply thawing, is not well known in at least some cases.

Most attention has focused on those ice streams in which high velocity is achieved despite very low basal shear stress, as on the Siple Coast (Figs. 1–3) and in paleo-ice streams of the southern margin of the Laurentide ice sheet [17] and elsewhere [18,19]. In all such cases of which we are aware, extensive and fine-grained tills contribute(d) to the extreme basal lubrication, see e.g., [20,21]. Bedrock typically is sufficiently heterogeneous that differential erosion produces a rough bed that interferes with rapid sliding, whereas till can smooth irregularities or allow motion of irregularities (ploughing; [22]), thus yielding faster motion. High basal water pressure is required to allow rapid till deformation, or to promote ice-till separation and rapid sliding or ploughing, e.g., [17,23–25]. In the ten cases of which we are aware where observations have been made to learn whether subglacial till was or was not deforming, all exhibited deformation (reviewed by [15,26]). However, in deforming beds beneath glaciers, considerable variability has been involved in the total depth of deformation, and whether deformation was broadly distributed, focused near the base of the ice, or focused at some greater depth in the till.

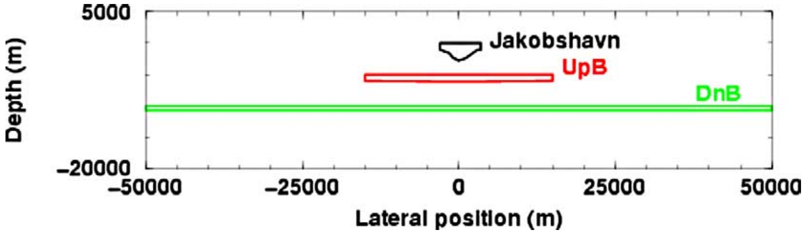

Fig. 1. Ice-stream cross-sections, with no vertical exaggeration, for Jakobshavn [8], Greenland, and ice stream B [1] along one of its main tributaries (UpB) and in downstream reaches (DnB), West Antarctica. Ice fluxes at UpB and Jakobshavn are not too dissimilar, with DnB carrying the flux from both major tributaries and so having roughly twice the flux of UpB.

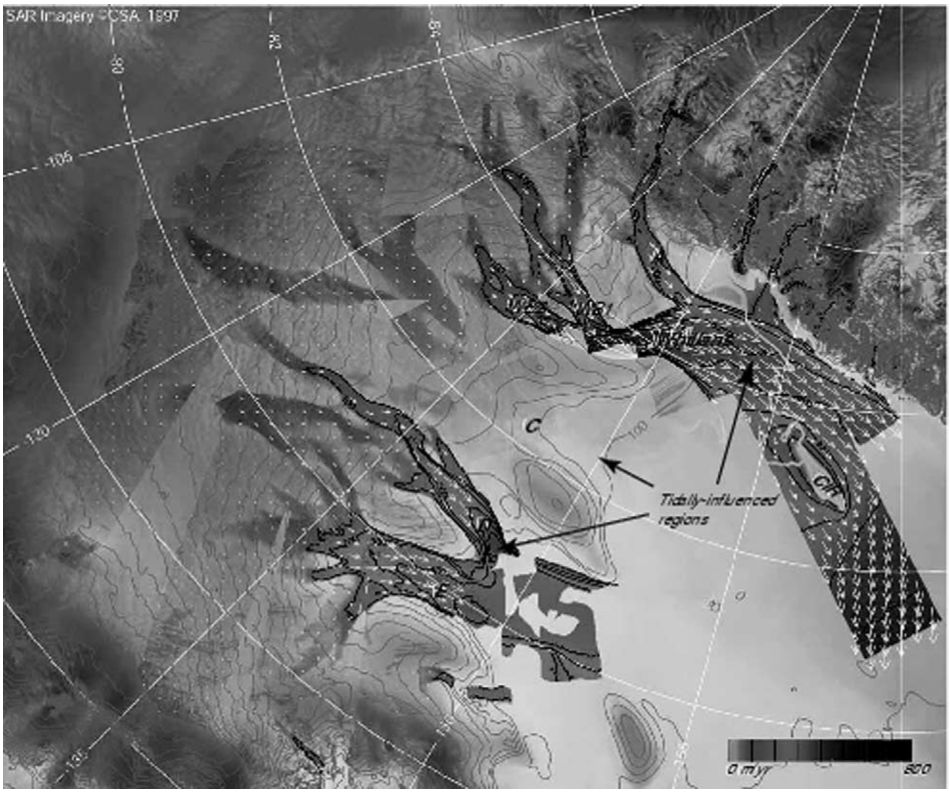

Fig. 2. Ice-stream velocities of Siple Coast of West Antarctica [68], together with locations discussed. Regions associated with documented tidal influences are indicated.

3. Why don't ice streams flow more rapidly? Basal drag, side drag and ice-shelf buttressing

Thick ice or efficient basal lubrication especially by water-saturated, fine-grained till explains the fast motion of ice streams. Restraint can arise from many causes, including basal drag (here applied to the contact with rock or till, whether that contact is nearly horizontal or not), side drag (here applied to drag with adjacent, slower moving ice), back-stress from downglacier and tension from upglacier. Of these, basal drag along the channel perimeter is likely dominant in the deep-bedrock-channel ice streams, side drag is quite important in well-lubricated ice streams, and ice-shelf buttressing may offer the greatest opportunity for ice-stream perturbations.

In the case of Jakobshavn and other thick ice streams, the viscosity of the ice is important, as is friction with the bed. Basal friction arises within any subglacial tills, from viscosity of ice flowing around bedrock bumps, and from the resistance of ice to regelation around bumps [6]. Friction between substrate and rocks within ice may also be quite important [24,27,28]. Driving stress is supported on the bed beneath and beside the ice in deep channels, and by lateral transmission through the ice-stream

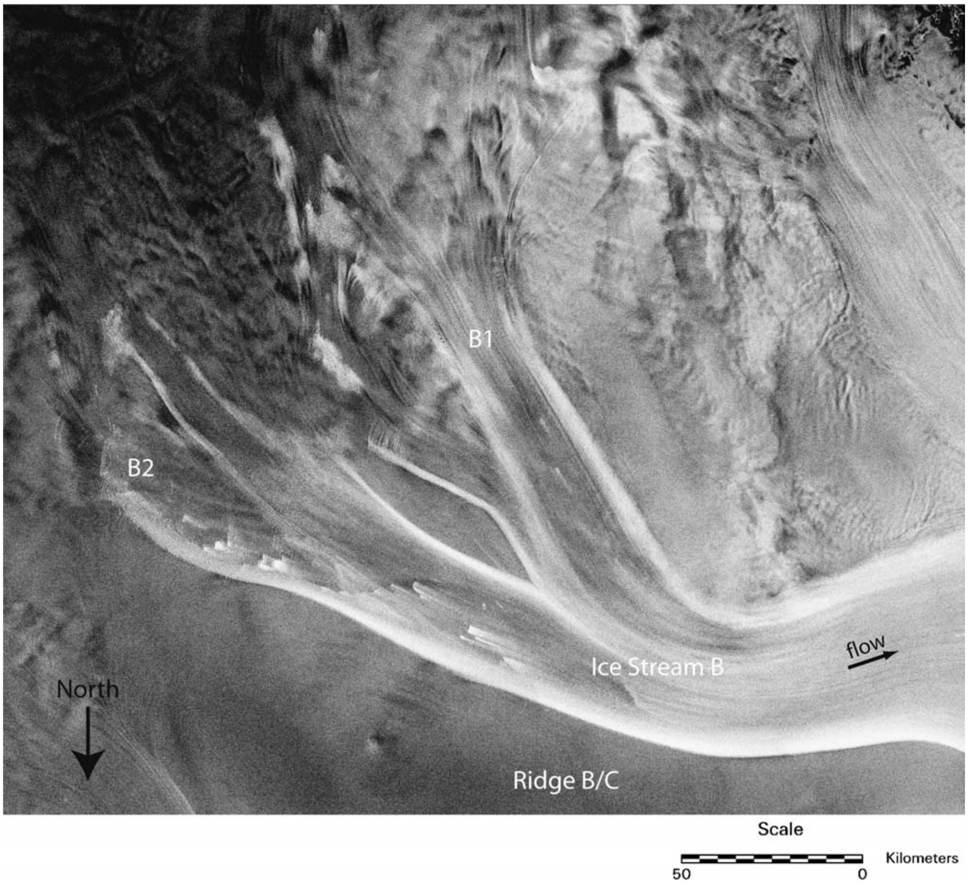

Fig. 3. SAR-derived image of Ice Stream B, modified from the RAMP image gallery [100]. North is down in the image. The ice-stream margins show up as white lines, with the flow being roughly from left to right and the arrow near the left margin serving purely to indicate the qualitative direction of flow.

sides above the bedrock channel and then to the bed. Gradients in longitudinal stresses may be nonzero in force balance of such deep-channel ice streams, but typically are not especially important compared to the basal and side drags [29].

For ice streams with especially efficient basal lubrication, such as those of the Siple Coast, side drag is important in addition to basal drag, but restraint from ice shelves downglacier also matters [30–32]. Several efforts at assessing controls on ice flow (e.g., [29,33,34]) have shown that side drag is important in most of the Siple Coast ice streams, with the stress transmitted to the bed under interstream ridges. This is a rather striking result, as these ice streams can be 50 to 100 times wider than they are thick (Figs. 1 and 2). Where side drag is dominant, the center-line velocity of an ice stream is expected to increase with the cube of the ice–air surface slope averaged across the ice stream, and with the fourth power of the ice-stream width [35, Eq. (15)], such that steeper or wider ice streams can move much faster.

As summarized in [35], the ‘softness’ of the ice in the shear margins of ice streams, and especially those on the Siple Coast, has proven difficult to estimate. Preferred c-axis fabrics or other features can develop in ice and change its viscosity, softening or hardening the ice for a particular pattern of deformation (see e.g., [36]). However, the high stresses in ice-stream shear margins likely produce recrystallization (nucleation and growth of new strain-free grains in ‘soft’ orientations), giving multiple-maximum fabrics only slightly softer than in random ice [36,37]. There is no evidence of overall hardening in ice-stream shear margins, but estimates of the softening of shear-margin ice in a particular region along a tributary of ice stream B range from 1- to 10-fold, complicating efforts to partition resistance to flow. The lower end of this range (3-fold or less softening, and probably close to 1) appears most likely [37,38].

Temperature also may be perturbed in shear margins, affecting the ‘softness’ of the ice. Shear heating from the high stress and rapid deformation in shear margins will warm and soften the ice there. However, surface crevassing trapping wintertime cold air will cool and harden the ice [39]. Furthermore, in those common (but not universal) regions where the ice stream surface is lower in elevation than surrounding slow-moving ridge ice, flow from the ridges to the stream will carry ice through the shear zone where temperature anomalies are generated. If this flow is fast enough, large temperature anomalies will not be produced [35,39]. Overall, the effects of temperature perturbations do not appear to be exceptionally large. Hence, whereas beds of ice streams can differ greatly from beds of surrounding regions, as described next, differences between the “softness” of ice-stream shear-margin ice and other ice probably are not large.

3.1. Till

Basal drag can be considered in various ways. One common terminology is to divide basal drag into those conditions applying over the majority of the bed (for low-shear-stress but fast-moving ice streams, this is the basal drag of subglacial till) and a higher level of drag on ‘sticky spots’, e.g., [38,40–44], which may be regions where till discontinuity reveals bedrock, local regions of high elevation in the bed, or other features that restrain flow.

Two quite different ‘end member’ behaviors have been suggested for the subglacial till, with implications for basal drag on till-lubricated ice streams. Steady behavior of till is expected to be closely approximated by a Coulomb-plastic rule, e.g., [15, 23]. In this case, deformation does not occur for stress below some threshold, but once deformation is initiated, arbitrarily rapid deformation can occur without increasing the stress. If this is valid, then where the bed is already soft and deforming, velocity of ice flow can change greatly without change in basal drag from till, and basal drag from till beneath already fast-moving ice streams will not increase if ice velocity increases. (A more-accurate treatment allows very large changes in velocity for nonzero but very small changes in stress [15].)

However, nonsteady mechanisms exist that can give radically different behavior, e.g., [26,45–47]. In particular, onset of till deformation is associated with dilatancy, as clasts move from a more closely packed configuration to a more open one in which motion over neighbors is easier. Dilatancy increases volume, and in the subglacial environment, that requires inflow of water to occupy the additional space created. Water inflow in turn requires that the hydrological potential, hence pressure, in the dilating region be lowered below the value in the surroundings. However, this strengthens till because till strength increases as local water pressure is reduced below overburden pressure. If onset of deformation is faster, dilation is larger, water-pressure reduction is larger, till strengthening is larger, and a larger stress is required. Hence, faster deformation requires larger stress, thus giving an apparent till viscosity even if the steady-state behavior is plastic. In this case, faster ice velocity would require a higher stress on the till bed.

The response of ice streams to tidal forcing [43,47,48] may show which case applies. Surveys on ice streams B, C and D on the Siple Coast have shown that they exhibit strong responses to the small ($\sim$1 m amplitude) ocean tide, with the response extending $\sim$100 km upglacier from the grounding line. The 0.1-bar fluctuation in sea-water pressure caused by the rising and falling of the ocean tide acting on the $<$1 km-thick ends of the ice streams is equivalent to a change in net driving stress for the ice streams of only $\sim$0.001 bar averaged over the region affected by the tidal signal. This is a rather small amount in comparison to the typical driving stress of $\sim$0.1 bar for the tidally influenced regions of ice streams C and D, yet large responses are produced. The tidal stress fluctuation is larger in comparison to the very low driving stress on the ice plain in the mouth of ice stream B, where especially large tidal effects on ice flow are observed.

On ice stream C, which slowed to only $\sim$10 m/year or less from a likely value of hundreds of meters per year a century or two ago; see [49], the magnitude of the tidally induced velocity fluctuations is not well known. Instead, basal seismicity caused by forward motion of the ice stream is strongly modulated by the tides [43,47–49]. On ice stream D, the velocity varies about twofold with the tides [48]. On ice stream B, the ice lurches forward (accelerating in as little as 30 seconds, then decelerating over 2 to 5 minutes at a site) just after high tide, sits without moving for about six hours as the tide falls, lurches forward again, then sits for about 18 hours until after the next high tide [47]. In all cases, the falling tide favors motion as the stress on the ice streams from sea water is reduced.

The motion of ice stream B in response to falling tide begins at some location(s), and then propagates across the ice stream at a speed similar to that of shear waves in subglacial till, roughly 100 m/s [47]. On ice streams C and D, upglacier propagation is observed but at much slower speed, of about 2 m/s on C and 5 m/s on D [43,48,49].

The behavior on ice stream B has been modeled successfully as a frictional slider [47]. This in turn can be interpreted as the behavior of a plastic till. Starting with the downglacier portion of the ice stream at rest, flow of ice from upglacier ‘loads’ the stationary ice until the basal shear stress exceeds the failure strength of the subglacial till, triggering motion that propagates elastically. Stress then falls to some residual strength threshold, when motion stops. No ‘damping’ term from viscosity of till is required.

On ice streams C and D, however, some damping term is required to explain the slow propagation of the tidal signal along the ice streams. One can conceive of several possibilities, including relaxation of side shear stress within the ice-stream ice where it abuts slow-moving interstream ridges, phenomena within a basal water system, or viscous relaxation within till.

On ice stream C in regions upglacier of the tidal influence, microearthquakes on sticky spots were observed to trigger cascades of additional quakes on other sticky spots up to 1.5 km away, with a propagation velocity of about 2 m/s [43]. The local nature of this behavior largely excludes any influence of the far-away sides of the ice stream. Similar propagation velocity of the tidal signals in downglacier regions, and the very small side restraint owing to the low ice-flow velocity, then indicate that the damping term for ice stream C arises at the ice-stream bed rather than along the sides. Similarity of the propagation velocities of tidal signals on ice streams C and D then suggests that basal delay is also dominant on ice stream D.

From these observations, it appears unlikely that shear within ice explains the delay, although fully three-dimensional modeling would be useful for testing these ideas. The water systems on C and D are known to exhibit large differences yet the subglacial tills are somewhat similar [15], and no successful model of how the water system could generate the observed delays has yet been proposed. In contrast, a (pseudo)viscous till model simply and naturally explains the behavior and yields appropriate material properties [48]. Thus, it seems likely that the tills under ice streams C and D are exhibiting a (pseudo)viscous behavior that explains the time-delays of the tidal signals. This in turn suggests that faster motion would require higher stresses to deform the tills, and that ice streams do not move faster in part because the subglacial tills resists faster motion.

The apparently plastic behavior of subglacial till of ice stream B, *versus* the apparently viscous behavior of ice streams C and D, may arise from the different stress levels on the beds. Ice stream C exhibits numerous sticky spots [42], leaving stress on till between sticky spots reduced from the driving stress, and ice stream D likely has much stress supported by side shear. Ice stream B, in its downglacier reaches, is especially wide with minimal side shear, and has thick till and thus apparently few sticky spots [50]. Sufficiently high stresses on basal till may cause failure producing plastic behavior, with lower stresses allowing pseudo-viscous behavior [47]. The relatively larger ratio of the tidal stress to the mean driving stress for the tidally influenced region of ice stream B in comparison to the tidally influenced regions of ice streams C and D is consistent with this view. In turn, this at least suggests the possibility that a sufficiently large increase in basal shear stress on an ice stream with (pseudo)viscous subglacial till, perhaps from loss of ice-shelf buttressing, could cause a switch to plastic behavior, reducing one of the stabilizing influences on the ice streams.

3.2. Longitudinal stresses

Some attention has been given to the possibility that tension from upglacier is important in restraining ice streams. However, as reviewed by [33], although this term is nonzero, it probably is not very important.

Much of the study of ice streams, especially in Antarctica, has focused on the possibility that ice shelves restrain the ice motion. This is an old idea (see e.g., [51–53]) that figures prominently in West Antarctic collapse scenarios. Surveys in the Ross Ice Shelf project, e.g., [30,31], showed that stretching rates at the upglacier end of the Ross Ice Shelf were smaller than expected for a freely spreading ice shelf, and thus that there must be a 'back stress' arising from interaction of the ice shelf with its sides or with pinning points. The magnitude of this stress, about 1 bar, is roughly one order of magnitude larger than the tidal stress changes responsible for the large modulation in ice-stream velocity. It remains possible that the response to a perturbation over times longer than one day would differ from the largely diurnal tidal changes, but the potential effects of ice-shelf changes must be considered and may be quite large.

In summary, ice streams are restrained by friction with their beds and their sides, and by resistance provided by ice shelves at the downglacier ends. For Jakobshavn-type thick ice streams, friction is very large at the bed and with the sides, but end effects may matter ([54]; see below). In Siple-Coast-type ice streams, resistance arising from the bed and the sides is also important with the sides relatively more important, and the end effects likely matter as well and probably are relatively more important. Resistance from the bed includes 'sticky spots' (probably regions of till discontinuity, or bedrock-cored features that stick up into the ice) and the strength of till; both Coulomb-plastic and (pseudo)viscous behavior seem to occur.

The end-member cases of the Siple Coast and Jakobshavn are useful. Perhaps equally instructive and more typical are the other modern ice streams, and particularly those entering the Filchner-Ronne Ice Shelf and Pine Island Bay in West Antarctica. Pine Island Glacier, Thwaites Glacier, the Rutford Ice Stream, and others seem to occur between the end-member cases [55–57]. Rutford, for example, exhibits a moderately deep bedrock channel, elements of soft-bedded and hard-bedded behavior, and much bedrock on one side reaching to the surface of the ice but thick ice on the other side [55]. Restraint by basal drag, side drag, and ice-shelf resistance are probably most important.

4. Can the ice streams change to flow more rapidly? Yes; warming is a concern, but predictions are still difficult

4.1. Historical insights

Abundant evidence testifies to the possibility of large changes in ice streams. Notable changes in the past include those of ice streams of the Laurentide ice sheet, and recent or ongoing changes of West Antarctic ice streams.

Fluctuations in the Lake Michigan lobe of the Laurentide ice sheet during the most recent (Wisconsinan) ice age included ice-marginal changes of approximately 500 m/year [58]. The Lake Michigan lobe was, at least in downglacier areas, a till-lubricated feature; topographic control was very weak in downglacier regions, more prominent in the Lake Michigan basin, but still not nearly so strong as for Jakobshavn Isbrae, for example. Other ice streams of the Laurentide ice sheet likely experienced large, rapid changes as well (e.g., [19]), and changes also are indicated for ice on Svalbard [59] and the Fennoscandian ice sheet and elsewhere [60].

Of special importance are the very prominent Heinrich events of the Hudson Strait ice stream [61], which had a till-lubricated aspect but with somewhat more topographic control than on the Siple Coast or south of Lake Michigan. The discharge of iceberg-rafted debris changed greatly and rapidly in association with the Heinrich events, with rates of deposition peaking at more than an order of magnitude above background values in the iceberg drift band of the open North Atlantic [62]. Numerous hypotheses have been proposed to explain the features, including mechanisms involving ice shelves and outburst floods. Both may have been involved (see [63,64]). However, neither floods nor ice-shelf mechanisms retrodict the very large changes in iceberg rafting–floods do not require icebergs at all, and large ice shelves where observed today allow net basal melting and loss of englacial debris before calving, serving as 'filters' to reduce or eliminate debris in icebergs, so formation and break-off of an ice shelf should release less debris than if the ice shelf never formed and bergs were calved directly. The leading hypothesis, that of sudden and rapid increase of flow velocity of the Hudson Strait ice stream, continues to appear to be accurate [2–4].

Considering recent Antarctic behavior, abundant evidence (e.g., [65,66]) attests to the highly nonsteady condition of the Siple Coast ice streams. The most dramatic change is probably that of ice stream C, which appears to have slowed from 'normal' ice-stream behavior to very slow velocity, $O(1\text{ m/yr})$ in many places, just over a century ago [49]. Extensive evidence of additional change can be found on the Siple Coast, with the likelihood that other regions around Siple Dome and elsewhere once exhibited streaming flow but do not now, and with slowdown observed on the ice plain of ice stream B near the grounding line [67,68]. Importantly, evidence of past slowdown is easier to observe than of speed-up, because active ice flow sweeps evidence out of the ice sheet whereas slowdown preserves the evidence inland. But, [35, Sections 4.1 and 4.2] reviews evidence of ice-stream widening as well as narrowing, with different rates of motion in both directions, and careful study of the Ross Ice Shelf provides probable evidence of speed-up as well as slowdown [69].

Several lines of evidence show that the glaciers flowing into Pine Island Bay have been changing rapidly over recent decades. Ice-stream widening (Pine Island Glacier; [70]) and strong near-coastal thinning of more than 1 m/year (affecting Pine Island, Thwaites and Smith Glaciers, with >4 m/yr on Smith between 1991 and 2001 [71,72], with slower thinning extending well over 100 km inland [71]) have occurred with ice-shelf thinning and grounding-line retreat [73–75] and very rapid basal melting of ice shelves [76].

4.2. Mechanisms: ice-shelf changes and surface melting

A short review such as this cannot go into details of the possible mechanisms of instability. Instead, we highlight a few areas in which new results are appearing or which are likely to be especially relevant.

Perhaps the most prominent hypothesis for perturbation of ice-stream flow is that loss of ice shelves will allow speed-up (see review by [77]). This may apply to changes occurring in parts of West Antarctica now.

Although cause(s) of the rapid, widespread, coherent changes in the Pine Island Bay region of West Antarctica is (are) not established yet, it is increasingly evident that all can be explained as response to ice-shelf loss or ice-shelf thinning caused by increased basal melting of ice shelves [63,78,79]. There can be little doubt from these and other papers, and evidence cited above, that ice shelves provide back-stress if confined or pinned, and that ice streams are sensitive to even quite small changes in back-stress. The models just emerging simulate these features, and show large inland response to ice-shelf change. Ice-shelf shrinkage may have occurred in response to warmer waters circulating in Pine Island Bay, similar to those observed in the Ross Sea region ([80], cf. [81]).

Support for the idea that loss of ice-shelf buttressing can cause enhanced flow of inland ice comes from recent observations of the ice-flow response to loss of portions of the Larsen ice shelf in the Antarctic Peninsula (reviewed by [77], cf. [82]). Collapse of part of the Larsen ice shelf in response to warming [83] was followed by a large speed-up in flow (as much as a tripling of velocity) of five of six tributary outlet glaciers [84]. Speed-up may have been caused by additional meltwater reaching the bed and enhancing basal lubrication [85] or by force-balance change in response to ice-shelf loss. However, because melting increased well before ice-shelf break-up, whereas ice-flow speed-up was prominent just after ice-shelf break-up, the force-balance effects of ice shelves were probably most important [86].

Ice-shelf buttressing even appears to be important for Jakobshavn [54]. There, thinning of several m/year within 20 km of the ice front started in 1997, spreading inland and slowing. The behavior is not explainable based on surface-mass-balance data. A short ice shelf of about 15 km length occupies the zone of fastest thinning. Waters in the fjord have warmed, and the mass-balance behavior of the ice is best explained by reduced back-stress from the ice shelf in response to melting induced by the warmer waters [54].

The results of [85] constitute a second mechanism by which speed-up could be caused. Meltwater penetrates through more than 1 km of cold ice in western Greenland to reach the bed. Summers with more melt exhibit faster ice velocity, and annually averaged velocity is inferred to be higher now than if no melt reached the bed. Hence, if surface melting increases in the future, additional water may reach glacier beds and increase ice-flow velocities. High mean-annual temperatures are probably not as important as high summertime temperatures. Because summertime temperatures on large ice shelves and adjacent low-elevation regions of ice streams are already within a few degrees of the melting point (e.g., [87]), extreme warming may not be required to affect ice flow. A targeted research effort to understand processes by which basal meltwater reaches glacier beds, and how rapidly abundant meltwater could reach glacier beds following initiation of widespread melting, seems appropriate. Additional issues are related to the ability of surface meltwater to lubricate flow; too much meltwater in steady state is known to lower water pressures and slow ice flow, but fluctuating and rising meltwater supplies favor high basal water pressures and rapid ice flow (reviewed by [88]).

4.3. Mechanisms: thermal effects

Many other hypotheses might be proposed by which ice-stream flow could change in the future. For example, loss of lubricating till through transport in excess of erosion could slow ice streams. However, among the many hypotheses, changes in thermal conditions at ice-stream shear margins and beds seem especially interesting.

The behavior of shear margins at the sides of ice streams was summarized by [35]. Because shear margins support so much of the stress of well-lubricated ice streams, and because centerline velocity increases with the fourth power of the width in such cases, widening could cause great speed-up, and narrowing great slow-down. As reviewed by [35] and discussed briefly above, both widening and narrowing have been observed or are recorded in paleoglaciological features.

Unstable behavior of shear margins is possible [89]. As one moves from the center to the side of an ice stream, velocity drops to low values more typical of non-streaming flow. If those ice-stream regions just inside of the shear margins are well-lubricated, the low velocity means that there is very low frictional heating (which is the product of velocity and basal shear stress). Freeze-on then is favored, which would narrow the ice stream, reduce the velocity further, and in turn tend to favor freeze-on inside this new marginal position. In contrast, the stress transmitted from a fast-moving ice stream through its shear margin is supported on the bed just outside of the shear margin. The stress there on the bed is very high, causing strong basal heating and favoring thawing, widening of the ice stream, hence higher side-shear stress and additional widening. Unless stabilized by topographic boundaries, by lateral limitations on till lubrication, by transport of lubricant from upglacier, or by other processes, unstable behavior may be possible.

Basal thermal behavior across the width of an ice stream may play an important role here. Where basal shear stress is large, as in deeply channelized ice streams, frictional heating maintains basal melting, allowing much basal sliding to contribute to ice motion. However, where basal shear stress is small owing to effective basal lubrication, frictional heating is small. Yet the large acceleration of ice as it flows into ice streams causes thinning that brings cold surface ice close to the bed, steepening thermal gradients and cooling the bed by heat conduction into the overlying ice [2–4,15,25,68]. This could provide a strong stabilizing feedback, in which faster flow of a well-lubricated ice stream leads to freeze-on and slow-down [23].

Initiation of basal freeze-on need not stop ice-stream flow, however, because of the effects of the subglacial hydrological system. Regions of thick ice upglacier of ice streams usually will have warmer beds than the ice streams, in part because the cold upper surface is farther away, favoring basal melting. The meltwater produced beneath inland ice will tend to flow to and along the ice streams. Initial freeze-on to ice streams can be supplied from this through-going water flow without loss of lubrication, and major slowdown or stoppage will occur only after this through-going water is (largely) exhausted [90,91]. Water loss associated with sub-ice-stream freeze-on has been implicated in the slowdown of ice stream C [25,68], although a special hydrological change upglacier may have been involved in routing inland water away from the lower part of ice stream C to ice stream B in a piracy event [49].

Observations under ice stream C [92] in one borehole show that at present it overlies a water-filled cavity. About 10 cm of debris-free, bubble-free ice has frozen onto the bottom, beneath about 1 m of debris-bearing ice with a high concentration of debris (perhaps $>50\%$). Above that is about 10 m of debris-bearing ice with a much lower debris content (also see [15,25,93]), which in turn underlies normal ice-sheet ice. A reasonable explanation is that the ice penetrated by the borehole experienced inland melting, and that freeze-on from through-going water producing dilute debris-bearing ice began following flow into the ice stream with associated vertical thinning [34,38], cf. [25]. After ice-stream slowdown and loss of some or all of the through-going water system, freeze-on of high-debris-concentration till began [94]. Finally, as the remaining slow flow moved the ice over a water-filled cavity (or as the water-filled cavity expanded under the ice?), freeze-on of clean, bubble-free-ice occurred. Certainly, calculations from the relatively recent slowdown and the measured basal thermal gradient [15] indicate that the dilute debris-bearing ice is too thick to have formed post slow-down. The low observed bubble concentration argues against entrainment without freeze-on [93]. Freeze-on far upglacier is unlikely because melting is expected there [34,38]. Hence, freeze-on from a through-going water system under active ice seems highly likely [34,90,91,93].

Water likely remains widespread under ice streams even in a freeze-on situation because of an inverse relation between basal shear stress and the difference between ice-overburden pressure and water pressure [40,95]. Where water intervenes between ice and bed, even if freeze-on is occurring from that water, the basal shear stress is supported on the upglacier sides of obstacles in the bed. This locally raises the contact pressure above the average ice-overburden pressure, producing corresponding pressure drops in regions downglacier of obstacles. Water is squeezed from high-pressure to low-pressure regions, and so accumulates in cavities in the lee of obstacles. Steady state with a through-going, distributed water system requires that the water pressure in these low-pressure cavities be such as to make them sufficiently widespread to maintain interconnection. The water-pressure drop from occupation of lee-side cavities would be zero in the absence of a basal shear stress, and increases with basal shear stress. The functional dependence on basal shear stress is not known, but a linear assumption is simplest. The similarity of the magnitudes of the basal shear stress and the water-pressure drop on Siple Coast ice streams [96] then indicates that water-pressure drop equal to basal shear stress is a reasonable first approximation. If lack of lubricating water or incipient freeze-on were to increase basal resistance to ice-stream flow on an incipient sticky spot, the increase in basal shear stress would lower the basal water pressure locally, causing water flow from surroundings to increase lubrication and supply of water, thus opposing sticky-spot formation. A sticky spot could locally generate differences in shear stress of $\sim$1 bar over $\sim$10–100 m [38,42], giving larger water-potential gradients than the Siple Coast regional mean from ice–air and bed slopes of $\sim$0.1 bar/km. Hence, continued lubrication is likely until the ice stream runs out of water [90,91].

These considerations on thermal balance, together with the calculations of [90,91], suggest that large and surprising flow changes can occur in well-lubricated ice streams; however, averaged over the appropriate time and space scales, these changes have little effect on the ice sheet as a whole. Local, perhaps unstable migration of shear margins can strand formerly fast-moving ice or accelerate formerly slow-moving ice while the ice stream continues to flow rapidly. However, too much basal cooling can lead to depletion of basal meltwater and freeze-on.

Retreat of the grounding line in the Ross Embayment from its advanced glacial-maximum position near the continental shelf edge probably was delayed until near the end of sea-level rise from deglacial warming of northern-hemisphere ice sheets, indicating that sea-level control on ice sheets is not especially tight [97]. The retreat likely began when post-glacial warming had penetrated to the ice-sheet bed and increased basal melting [90,91]. Continued gradual retreat but with surprising local variations has occurred, and the thinning associated with the retreat may be cooling the ice-sheet bed and favoring freeze-on now. Flow-line simulations indicate that sufficient basal meltwater remains for continued retreat [90,91]; however, more-comprehensive modeling driven by better data on geothermal fluxes would improve confidence in this result.

From this discussion of mechanisms, it is clear that warming affecting ice shelves can lead to ice-stream speed-up and transfer of mass to the ocean, and this process seems to be active for Jakobshavn and in the Pine Island Bay region, and has occurred following breakup of the Larsen Ice Shelf. The possibility remains that future surface meltwater in a warming world could penetrate to the beds of ice streams and increase flow velocities. Highly sensitive response of the large ice sheets to small sea-level rise does not seem likely. Stabilization from basal cooling perhaps in response to future speed-up is possible but does not appear likely at this time.

Over longer times, perhaps the most interesting possibility is that a potential ice stream (based on till-covered, smooth bedrock) exists somewhere (perhaps in the East Antarctic ice sheet) but is not now active because it is frozen to the bed. Basal warming in such a place would have large potential to cause a speed-up that would transfer ice to the ocean. Just as new ice streams seem to have developed in the Laurentide ice sheet at times of basal warming [2–4,60], such an event could perhaps occur in the future. Geophysical exploration to find potential sites, and glaciological assessment of the possibility of changes, would be interesting.

5. What does the future hold? Much is known, but much more remains to be learned

Given the large uncertainties introduced by lack of understanding of controls on shear-margin positions, and related to the competition between basal freeze-on from enhanced thinning and basal melting from enhanced friction in response to an ice-stream speed-up, large difficulties remain in making accurate predictions of ice-stream behavior.

The state of the cryosphere has been reviewed by [98]. Results since then have shown that fast-moving ice can accelerate in response to warming-induced loss of ice shelves [84,86], and increased the likelihood that this applies to large ice streams as well as to smaller ones [78,79]. Contribution of increased surface melting of inland ice to increased ice velocity has also been demonstrated [85], and suggests greater shrinkage of ice in a warming world than previously calculated [99], although the applicability of the melt-induced speedup to ice streams remains conjectural. Much work remains to be done.

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