Ice shelf buttressing

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The Antarctic and Greenland ice sheets are the most massive bodies of ice in the world, containing about 30 million and 3 million km^3 of ice, respectively. Input of mass to the ice sheets is exclusively through snowfall on the surface densifying into glacial ice. Mass is lost through melting, either at the surface or at the base of the sheets, and through icebergs calving off the margins of the ice sheets into the ocean. The two ice sheets differ greatly in the importance of these outputs, however: in Greenland, surface melting is quite important and a large part of the ice sheet margin is land-terminating. In Antarctica, surface melting is negligible and ice flows into the ocean around its entire perimeter. These inputs and outputs control the volume of ice in the sheets, a quantity of great societal importance because it affects global sea levels.

In Antarctica, snow accumulation has increased in recent decades, possibly due to warming air masses. Loss of ice by flow across the ice sheet margin, however, is poorly characterized and its future behavior is almost completely unknown. Around most of Antarctica, the ice sheet goes afloat in the form of large ice shelves, which vary in area from the size of Greater London to roughly that of France. Since the ice shelves are in hydrostatic balance with the ocean, their direct sea level contribution would be negligible were they to melt entirely. On the other hand, the *indirect* effect of such an occurrence would have far-reaching effects, and in some cases could threaten the stability of large parts of the Antarctic ice sheet. This is due to "ice shelf buttressing," the term used to describe the resistive forces imparted to the ice sheet by its ice shelves.

Physical setting and theory

The transport of ice toward the ocean is not uniform around the continental margin; rather, it is concentrated in relatively narrow, fast-flowing *ice streams*. These ice streams empty into ice shelves, and the point at which the ice goes afloat is called the *grounding line*. The position of the grounding line is determined by a floatation condition: the weight of the ice column must equal the weight of the water column it displaces. In other words,

$$\rho g H = \rho_w g D \qquad (1)$$

where ρ and ρ_w are average ice and ocean densities, respectively, H is the ice thickness (distance from top to bottom), and D is the bedrock depth. Thus the grounding line is dynamic; if ice thickness changes, the grounding line will adjust to a position where equation 1 is satisfied. As ice shelves do not contribute to sea level, any changes in grounding line position, or in the flux of ice across the grounding line, translate to sea level change.

To appreciate the role that ice shelves play in grounding line position and ice flux, one should first examine the dynamics of a grounding line with no ice shelf attached, that is, where there is simply an ice cliff at the grounding line, and all ice that crosses it calves off instantly as icebergs (Figure 1). In this case there is an imbalance between the pressure within the ice and the
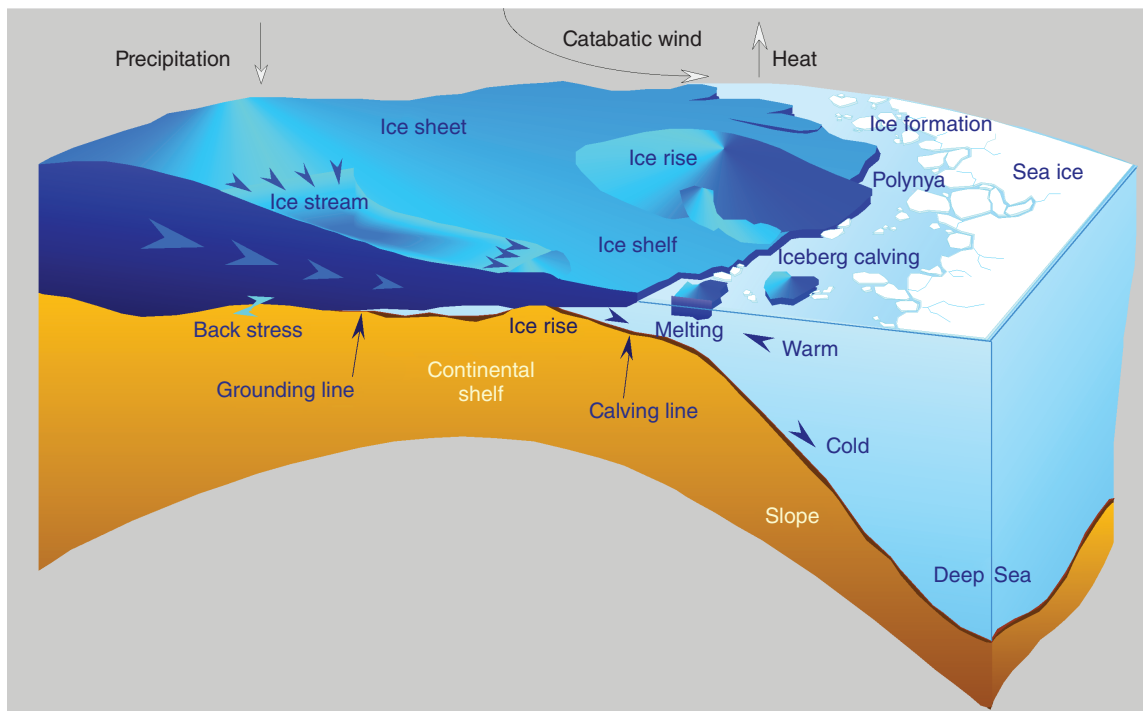
Figure 1 A visualization of many of the features and concepts discussed in this article. Hannes Grobe, Alfred Wegener Institute for Polar and Marine Research, Germany. Reproduced from Wikipedia (http://en.wikipedia.org/wiki/File:Antarctic_shelf_ice_hg.png).

ocean pressure along the ice face, and this results in a tensile stress acting at the grounding line in the seaward direction, pulling the ice forward. Averaged over the depth of the ice sheet this equals

$$\sigma = 1/4\Delta\rho gD \qquad (2)$$

where $\Delta\rho = (\rho - \rho_w)$. (Note: some sources quote a $1/2$ factor on the right–hand side; this is because they are discussing *membrane stress*, which in this context is twice the tensile stress.) This tensile stress induces a flow of ice across the grounding line; however, the relationship between equation 2 and ice velocity depends on a number of factors. Weertman (1974) derived an approximate relationship between the two, and deduced the result that ice flux across the

grounding line increases quite strongly with depth. More recently, Schoof (2007) derived the analytical result that ice volumetric flux per unit width is proportional to $\sigma^\alpha D^\beta$, where α varies from approximately 1.5 to 2.25 and β from approximately 2 to 2.5, depending on the nature of the ice-bed interface.

That ice flux increases with grounding line depth has far-reaching implications for ice sheet behavior and stability. A large sector of Antarctica, the West Antarctic Ice Sheet, is a *marine ice sheet*: it rests on a bed that is far below sea level, in part due to the weight of the ice and in part due to millennia of erosion. Moreover, it is deeper, on average, toward the center than at the margins. For such an ice sheet to be in steady state, input from snow accumulation, which is roughly

proportional to its area, needs to be balanced by the flux of its ice streams across its grounding line. If stress and flux are related as described above, such an ice sheet cannot be stable.

To see this, consider a small inland retreat of part of its grounding line: this decreases the input to the sheet by decreasing its surface area. At the same time, the grounding line has retreated onto a deeper bed, increasing the flux. Thus, in this new configuration, less ice is delivered to the grounding line than moves across it, and ice thins at, and upstream of, the grounding line. Since total accumulation input has decreased, this imbalance will not correct itself; rather, ice at the grounding line will go afloat and both the thinning pattern and the grounding line will move further inland, exacerbating the imbalance. With nothing to arrest the process, the entire ice sheet might eventually collapse, meaning a substantial portion would go afloat and subsequently calve into icebergs or melt. Similarly, a small advance of the grounding line would result in unchecked advance to the edge of the continental shelf. This instability is often referred to as marine ice sheet instability (MISI) (Mercer 1978). The instability has significant implications: if the entire West Antarctic Ice Sheet were to collapse, sea level would rise by 3–5 m globally.

An important point to make is that the above arguments hold not only for a grounding line with no ice shelf, but also for a grounding line with an ice shelf that is completely unconfined, that is, it does not exist in an embayment and there are no points of contact between the ice and the ocean bottom. However, such is rarely the case for Antarctic ice shelves. Most ice shelves are embayed, flanked by slow-moving grounded ice that limits the oceanward flow of the ice shelves; and the larger ice shelves – the Ross Ice Shelf and the Filchner–Ronne Ice Shelf – are pinned at large ice rises in their interior. These embayment walls and ice rises exert a resistive force on the ice shelf, which is carried through the ice shelf and exerted on the grounding line, lowering the tensile stress relative to equation 1, and in turn decreasing ice velocity relative to the unconfined case.

This ability of ice shelves to exert resistive force and decrease velocity of outlet streams is commonly known as buttressing: the ice shelf buttresses against the tendency of ice to flow under its own weight. Strictly speaking, the force is not originating from the ice shelf itself; the ice shelf is simply allowing forces along different parts of the grounding line to be felt nonlocally. If a part of the grounding line, for example a ridge along the side of the shelf, experiences strong friction at its base, the ice shelf allows this frictional force to slow other parts of the grounding line, such as the outlet of a fast-moving ice stream. In the following, different types of buttressing will be examined.

Buttressing by rigid sidewalls

Consider an ice shelf in a rectangular channel of width W and length L; at one end ($x = 0$), an ice stream flows into the shelf; and the other end ($x = L$) is open to the ocean. On either side, the ice is slow-moving, and at every point along these rigid sidewalls a shear stress τ resists flow in the ice shelf. It can be shown (Thomas 1973) that the tensile stress felt at the grounding line at $x = 0$ is lessened by F, where F is the total amount of force arising from this stress. If τ is assumed uniform, then tensile stress at $x = 0$ is given by

$$\sigma = 1/4 \, \Delta \rho g D - \tau (H_s/H) \, (L/W) \quad (3)$$

where H_s is a representative ice shelf thickness (in general smaller than H; see Figure 2). Let us consider how this might change grounding line velocity. If we consider a value of $750\,\text{m}$ for D, and 10^5 Pa for τ, then σ will decrease by a factor of approximately
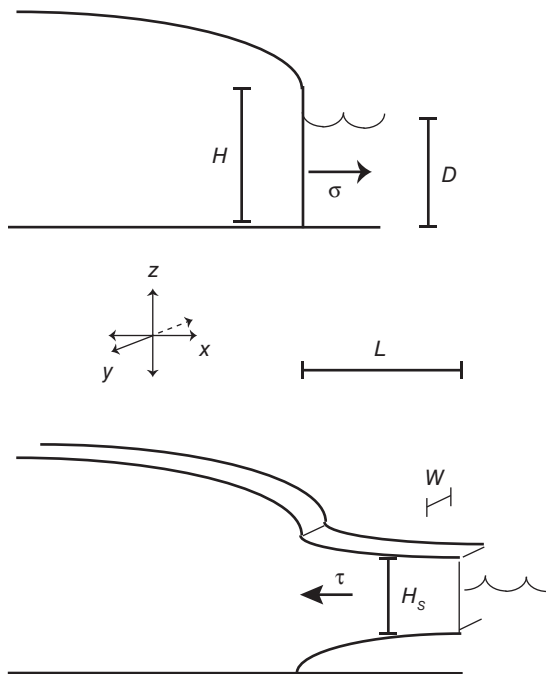
Figure 2 Two simplified cases of an unconfined and buttressed grounding line: (top) an unconfined shelf with depth D at the grounding line; (bottom) a confined ice shelf. There is stress τ along the sides of the ice shelf which effectively slows ice velocity.

$$0.5 \times (L/W) \times (H_s/H) \qquad (4)$$

relative to its unconfined state. From the arguments of Schoof (2007), this can be translated to velocity change. For instance, if the unconfined grounding line velocity (the velocity if τ is zero) is 1000 m per year, an ice shelf of aspect 1 ($L/W \approx 1$) and thickness $H_s/H \approx 0.5$ will reduce this speed to approximately 600–650 meters per year. A narrower, longer, or thicker ice shelf will reduce this speed even more.

It should be pointed out, though, that the above analysis is oversimplified: equilibrium ice shelf thickness would change with width and length. Additionally, τ would not be constant, but would decrease as the shelf widens, and σ is unlikely to drop to zero. To truly determine the effect of buttressing, the differential equations governing stresses within the ice must be solved. In general, though, the narrower, longer, and thicker an ice shelf, the more effective it will be at buttressing the flow of the ice stream that feeds it.

This has implications concerning MISI: if a grounding line retreats, the length of its attached ice shelf will increase, unless its calving front retreats at the same rate (which is unlikely because of the different physics determining the positions of the two). Thus, grounding line tensile stress could actually decrease as grounding line moves inland, even though the bed deepens, and the grounding line may stabilize. It should be pointed out that whether or not grounding line retreat is unstable depends on a number of factors, including the specific geometry and mechanical properties of the bed, and any factors influencing the ice shelf, such as crevassing and melting by ocean currents.

Buttressing by ice rises and ice rumples

Theoretical studies suggest that a relatively narrow ice shelf is capable of reversing MISI, but a wide ice shelf is not. The Ross and Filchner-Ronne ice shelves are 700–800 km wide at their calving fronts, and it is unlikely that these ice shelves could effectively restrain ice stream flow through sidewall buttressing alone. However, these ice shelves come in contact with the ocean bed in several places in the form of massive ice rises – regions of grounded ice that are not connected to the main ice sheet. Ice flows away from the rise in all directions, as if it were an isolated ice cap or ice sheet, implying a strong (likely frozen) bed. Ice rises influence ice shelves in much the same way as slow-moving ice at ice shelf boundaries, and in

doing so effectively narrow a wide ice shelf. In Ross Ice Shelf, Roosevelt Island and Steershead and Crary ice rises are spaced more closely than the width of the overall ice shelf, and so the sides of the rises exert a stronger buttressing force on the Siple Coast ice streams than the side margins of the ice shelf could. Roosevelt Island may have had a strong influence on the location of the grounding line of MacAyeal Ice Stream since the Last Glacial Maximum. Similarly, it has been suggested that ungrounding of these ice rises may have had a dramatic impact on the flow of Whillans and Kamb ice streams; See Figure 3.

Ice rumples are isolated regions of grounding as well, but ice slides over the contact; the ice shelf flow slows but does not stagnate. In general, the surface elevation contrast between rumples and the surrounding shelf is less than that of ice rises. Ice rumples provide buttressing as well, by transmitting frictional force from the bed back to the grounding line. While the bed is weaker than under ice rises, ice rumples can be as important as ice rises in providing buttressing. There are several locations in which ice rumples are thought to provide ice stream stability, or have in the recent past. As recently as a few decades ago, Pine Island Ice Shelf, West Antarctica, was in contact with an undersea ridge; ungrounding of the shelf from this ridge is thought to have led to acceleration of Pine Island Glacier (Jenkins *et al.* 2010). On the other hand, the adjacent Thwaites Ice Shelf is in contact with a seamount at its base, and while this local grounding has a noticeable effect on the flow of the ice shelf, recent studies indicate it does not play a strong role in the stability of Thwaites Glacier.

Buttressing by unconfined ice shelves

A frequently made statement is that an unconfined ice shelf does not exert any buttressing force on ice flow at the grounding line. This is not completely true, however. Ice shelf buttressing refers to any transfer of force across distances due to the presence of an ice shelf. If a section of the grounding line overlies a stronger bed than the rest, this stronger bed effectively provides a restraining force to weaker-bedded parts of the grounding line; the strong-bedded section is essentially pulled forward by the weaker-bedded sections. If the ice shelf were to be removed, the latter would accelerate (and the former would decelerate somewhat). In general, buttressing by unconfined ice shelves is likely less important than that of sidewalls and ice rises, but it may be of importance to short, narrow ice tongues, such as those in Greenland fjords whose margins are too crevassed and broken up to provide any support.

Climate impacts on ice shelf buttressing

As in the simple example of sidewall buttressing, thinning of an ice shelf will limit its buttressing ability. However, the cause of thinning is important. If the shelf thins because it is "starved" by the ice sheet, it may be due to changes in the ice sheet interior, such as long-term changes in precipitation. But if the thinning is due to melting at its lower surface by heat from the ocean, the impacts on the ice sheet can be very important. As the ice shelf thins due to melting, buttressing is lost and velocities at the grounding line increase. The thinning of grounded ice propagates inland because the lowering of the ice surface increases the basal slope, which increases velocities. If the ice sheet bed deepens inland, extensive grounding line retreat occurs as well. The velocity increase results in more mass being fed to the shelf; but if the ocean contains sufficient heat to remove this ice, the ice shelf does not thicken and the retreat does not reverse itself.
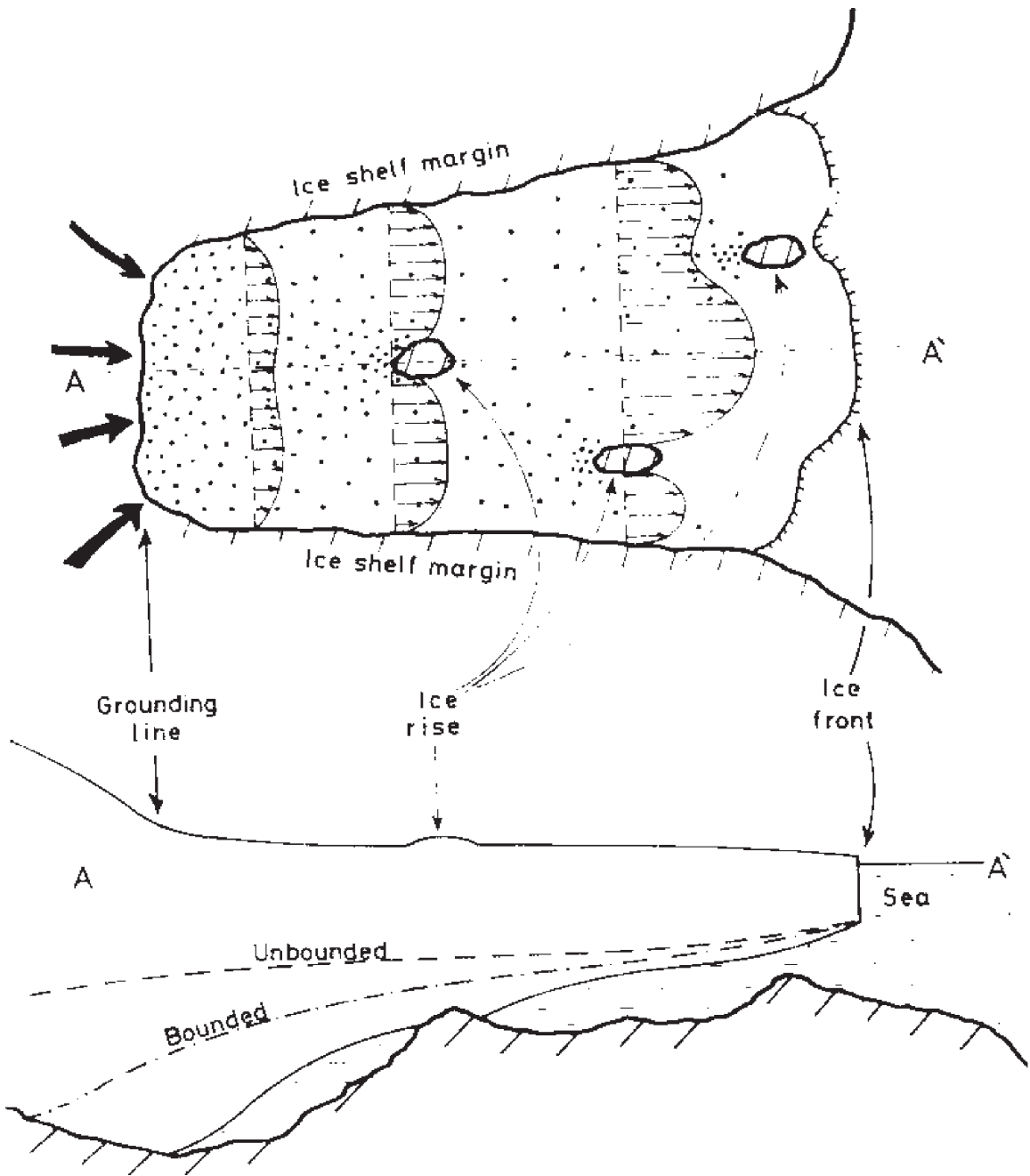
Figure 3 Plan view (top) and section (bottom) of an ice shelf buttressed by rigid sidewalls and ice rises. Arrows represent ice velocity magnitude and speckling density represents the level of resistive force carried by the ice shelf. Thomas 1979. Reproduced by permission of International Glaciological Society (IGS).

This process is currently occurring on several ice shelves in Antarctica, most notably in those of the Amundsen Sea Embayment in West Antarctica. Almost all of the ice shelves around Antarctica are exposed to some degree of melting in their deepest parts, since ocean waters can be no colder than the surface melting point (about −2°C) due to sea ice formation, and the melting point decreases with depth. However, due to deep underwater troughs in the Amundsen Sea, along with the fact that waters tend to be warmer at depth at high latitudes, these ice shelves are exposed to waters of 1.2°C or greater (Jenkins *et al.* 2010). Exposure to these warm waters results in large thinning rates; ice shelf thinning of 6–8 m per year has been inferred, which has caused extensive speedup and thinning of the Pine Island, Thwaites, and Smith glaciers (Shepherd, Wingham, and Rignot 2004). On Pine Island, thinning and speedup are seen hundreds of kilometers upstream from the ice shelf (Wingham, Wallis, and Shepherd 2009; See Figure 4). This thinning could have far-reaching implications: these ice streams (Pine Island, Thwaites, and Smith) drain a portion of the ice sheet that could raise global sea level by 1.3 m, and they rest on inland-deepening beds, making them unstable.

Ice loss due to melting ice shelves is not limited to the Southern Hemisphere. The floating tongue of Jakobshavn Isbrae, a fast-flowing outlet glacier in southwest Greenland, rapidly thinned and broke up when a shift in large-scale ocean circulation brought relatively warm waters from the Atlantic into its fjord. The loss of this small ice shelf led to a doubling of the glacier speed (Holland *et al.* 2008). Elsewhere in Greenland, marine-terminating glaciers without floating extensions can be exposed to high melt rates along their calving cliffs. It is important to realize, however, that this is different from ice shelf melting: in the latter case, grounded ice is
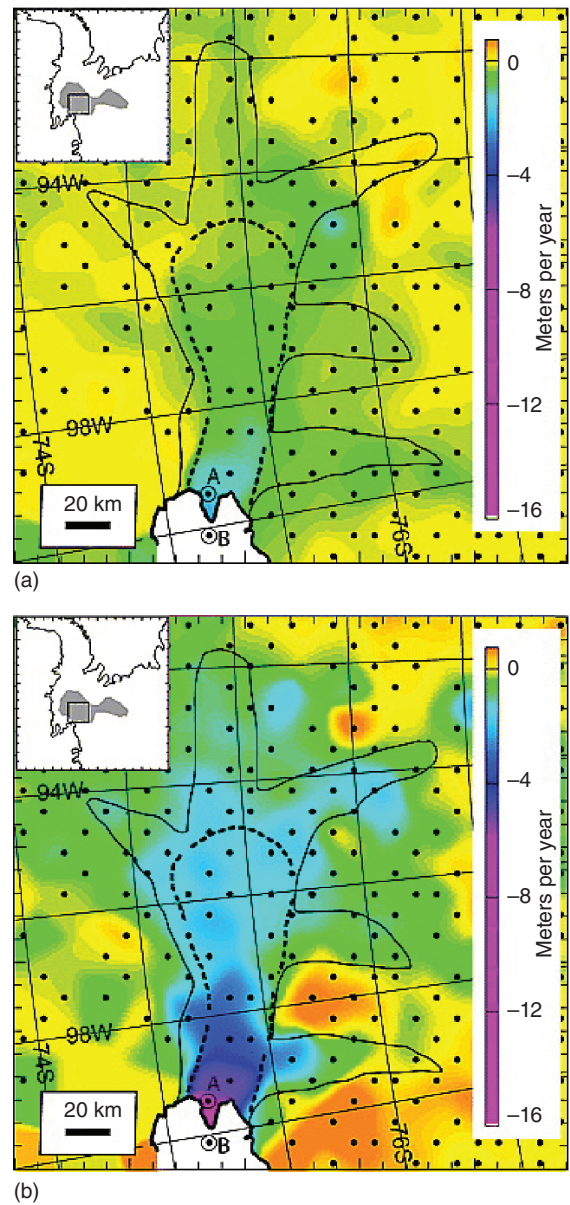

Figure 4 Thinning rates on Pine Island Glacier in 1995 (top) and 2006 (bottom). Ice flows from top to bottom; the ice shelf is uncolored. The pattern of increased thinning rates extends several hundred kilometers from the grounding line. Wingham, Wallis, and Shepherd 2009. Reproduced by permission of John Wiley & Sons, Ltd.
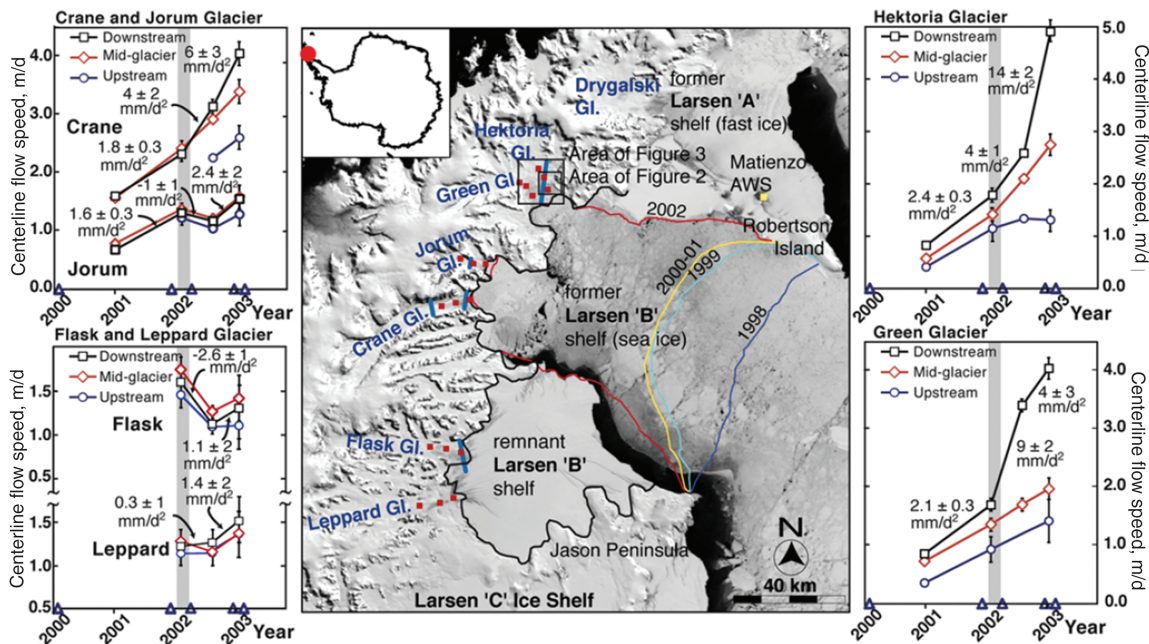
Figure 5 Center: Satellite photo of Larsen B embayment with historic ice shelf extents. Margins: velocities on input glaciers at different times. Crane, Hektoria, and Green glaciers have sped up at their terminus since the collapse. Flask and Leppard remain protected by the remnant Larsen B. Scambos *et al.* 2004. Reproduced by permission of John Wiley & Sons, Ltd.

affected through the loss of buttressing; in the former, melting affects grounded ice directly.

Loss of buttressing due to melting has been observed to be a slow process, operating over years to decades. However, far more rapid collapse of ice shelves has been observed. In the past two decades, several ice shelves along the Antarctic Peninsula have "disintegrated": seemingly intact, they are transformed into a mass of small icebergs in a matter of days and then drift away. The process is thought to be due to melting at the upper surface by solar radiation and heat from the atmosphere, as the Peninsula lies farther north than the rest of Antarctica. Extensive melt ponds form, seeping into surface crevasses; the crevasses then quickly deepen, wedged open by the weight of the water. The breakups may have been preconditioned through the removal (by

iceberg calving) of parts of the ice shelves vital to their structural stability. At any rate, after the disintegration of one of the largest of these ice shelves, Larsen B, large speedups were observed on the ice streams that fed it, indicating that the Larsen B provided buttressing for these streams (Scambos *et al.* 2004; see Figure 5).

Finally, it should be stated that the mechanisms through which the ocean and atmosphere affect ice shelves are still poorly characterized. An improved understanding of the processes involved is necessary before the effects of climate change on the Antarctic and Greenland ice sheets can be assessed.

SEE ALSO: Antarctica; Climate change and land ice; Glacial erosional processes and

landforms; Ice caps; Ice sheets; Ice shelves; Oceans and climate; Sea level rise

References

Holland, D.M., R.H. Thomas, B. De Young *et al.* 2008. "Acceleration of Jakobshavn Isbrae Triggered by Warm Subsurface Ocean Waters." *Nature Geoscience*, 1: 659–664. DOI:10.1038/ngeo316.

Jenkins, A., P. Dutrieux, S. Jacobs *et al.* 2010 "Observations beneath Pine Island Glacier in West Antarctica and Implications for Its Retreat." *Nature Geoscience*, 3: 468–472. DOI:10.1038/ngeo890.

Mercer, J.H. 1978. "West Antarctic Ice Sheet and CO_2 Greenhouse Effect: A Threat of Disaster." *Nature*, 271: 321–325. DOI:10.1038/271321a0.

Scambos, T.A., J.A. Bohlander, C.A. Shuman, and P. Skvarca. 2004. "Glacier Acceleration and Thinning after Ice Shelf Collapse in the Larsen B Embayment, Antarctica." *Geophysical Research Letters*, 31(18): L18402. DOI:10.1029/2004GL020670.

Schoof, C. 2007. "Ice Sheet Grounding Line Dynamics: Steady States, Stability, and Hysteresis." *Journal of Geophysical Research*, 112: F03S28. DOI:10.1029/2006JF000664.

Shepherd, A., D. Wingham, and E. Rignot. 2004. "Warm Ocean is Eroding West Antarctic Ice Sheet." *Geophysical Research Letters*, 31: L23402. DOI:10.1029/2004GL021106.

Thomas, R.H. 1973. "The Creep of Ice Shelves: Theory." *Journal of Glaciology*, 12(64): 45–53.

Weertman, J. 1974. "Stability of the Junction of an Ice Sheet and an Ice Shelf." *Journal of Glaciology*, 13(67): 3–11.

Wingham, D.J., D.W. Wallis, and A. Shepherd. 2009. "Spatial and Temporal Evolution of Pine Island Glacier Thinning, 1995–2006." *Geophysical Research Letters*, 36: L17501. DOI:10.1029/2009GL039126.