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Science **338**, 1172 (2012);
DOI: 10.1126/science.1226481

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The ice sheets of Greenland and Antarctica are losing ice at accelerating rates, much of which is a response to oceanic forcing, especially of the floating ice shelves. Recent observations establish a clear correspondence between the increased delivery of oceanic heat to the ice-sheet margin and increased ice loss. In Antarctica, most of these processes are reasonably well understood but have not been rigorously quantified. In Greenland, an understanding of the processes by which warmer ocean temperatures drive the observed retreat remains elusive. Experiments designed to identify the relevant processes are confounded by the logistical difficulties of instrumenting ice-choked fjords with actively calving glaciers. For both ice sheets, multiple challenges remain before the fully coupled ice-ocean-atmosphere models needed for rigorous sea-level projection are available.

There is long-standing concern about Greenland and Antarctica's contributions to rising sea level. Earlier projections indicated ice-sheet losses in Greenland would be offset largely by gains in Antarctica, where a warmer atmosphere is expected to increase snowfall more than surface melt (*1*). Such projections are at odds with recent observations, which indicate that both ice sheets are losing mass at accelerating rates. Increased ice discharge to the ocean (i.e., greater iceberg calving) has produced much of this increased loss as numerous glaciers and ice streams have sped up rapidly (months to years) over the past two decades (*2*, *3*). Largely unanticipated as recently as a decade ago, many of these changes were observed just as the Intergovernmental Panel on Climate Change (IPCC) was preparing its Fourth Assessment, prompting the panel to conclude that they could place no upper bound on the ice dynamics–driven contribution to sea level (*4*).

The recent changes in ice flow have highlighted the important role that ice-ocean interaction plays in ice-sheet stability, which has sometimes been overlooked. The ocean's tremendous heat capacity means that shifting currents or warming water can substantially alter the rate of melting at the ice-ocean interface (*5*). This interaction is particularly important in Antarctica, where little melting occurs at the ice-air interface. By contrast, considerable (tens of cm to tens of meters per year) melting occurs beneath the continent's fringing ice shelves, which are floating extensions of the grounded ice sheet (*6*). Removing ice from a land-terminating margin requires that heat be supplied to melt or sublimate ice in situ, limiting the rate of loss. Ocean currents, however, can rapidly carry away excess ice to melt elsewhere after it calves from an ice-ocean terminus in response to forcing at the boundary (e.g., oceanic or atmospheric heating), enabling rapid retreat. Lastly, the ice-sheet grounding line (Fig. 1), where the grounded ice sheet transitions to a floating ice shelf, often lies at a point of tenuous stability, such that small initial perturbations can trigger large-scale retreat (e.g., marine ice-sheet instability) (*7*).

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Oceanic Setting

Along an ice-sheet periphery, the ocean surface waters tend to be relatively fresh and cold (Fig. 2, C and D), typically at or near the surface freezing point. The properties of such waters typically are of polar origin and have only modest impact on melting beneath ice shelves. Below these surface waters, at depths typically ranging from 100 to 1000 m, there often resides a relatively warm and salty layer of water originating from the subtropical or subpolar regions (Fig. 2, C and D). These water masses are referred to as Irminger Water (IW) and Atlantic Water (AW) south and north of Greenland, respectively, and as Circumpolar Deep Water (CDW) surrounding Antarctica (Fig. 1). Although warmer than the polar surface waters, these deeper water masses actually are denser because of their greater salinity, and consequently they sink below the surface as they approach the polar regions. The fact that such warm water masses have no surface expression in the vicinity of the ice sheets makes them impossible (thus far) to detect via remote sensing, greatly complicating the task of monitoring water-mass changes. These warm waters have a large impact where they contact glacial ice, causing melting rates of orders of tens or more meters per year (*5*, *6*).

Accurate projections of the role of ocean-induced ice-sheet melting require improved models of the variability in circulation of CDW and IW/AW where they come in contact with glacial ice (Fig. 2D). From scant observations taken over the past few decades, these water masses are known to vary, often in response to fluctuations in atmospheric circulation. In particular, westerly winds blowing across the Southern Ocean and North Atlantic influence the movement and distribution of CDW and IW/AW water, respectively (*8*, *9*). As a result, what controls shifts in the winds is an area of intense current research, and part of the answer appears to be natural variability, internal to the climate system, making projections of future change all the more challenging (*8*–*10*).

Ice Shelves

Much of Antarctica and areas of northern Greenland discharge ice through large floating ice shelves (Fig. 3). Once afloat, ice largely completes its direct contribution to sea level, but ice shelves exposed to oceanic melting (Fig. 1) and other climatic processes still play an important role in regulating ice-sheet discharge. With no ice shelf, the column-integrated pressure difference across the ice-ocean boundary produces a seaward-directed force at the grounding line, which is resisted upstream by friction at the ice-bed interface (*11*). The ocean provides negligible traction, so an unconfined (i.e., ocean-only contact) ice shelf does not alter this stress distribution. Where an ice shelf is confined by an embayment or pinning points, however, the resulting drag produces ice-shelf buttressing that offsets some of the traction that the grounded ice upstream would otherwise provide (*11*, *12*). Thus, any loss of this buttressing from ice-shelf shrinkage or breakup must in turn be compensated for by increased drag from upstream. For the nonlinear viscous flow of ice, this restoration is achieved through increased strain rates (i.e., faster flow) (*11*, *13*–*16*).

The bedrock geometry on which an ice sheet rests provides an important control on its stability. With no ice-shelf buttressing, ice discharge scales nonlinearly ($n > 3$) with grounding-line thickness, making it difficult to stabilize the grounding line on slopes where the bed deepens with distance inland (*7*). Thus, for extended regions where ice rests on a bed that is both well below sea level and deepens toward the interior, this nonlinearity leads to the so-called marine ice-sheet instability (*17*). Stabilizing factors such as ice-shelf buttressing and locally interior-shallowing slopes have allowed the predominantly marine West Antarctic Ice Sheet to maintain a relatively stable geometry even while sustaining moderate regional retreat over the past several millennia (*18*, *19*). Ice-ocean interaction now appears to be accelerating retreat and ice-sheet loss with consequent impact on sea level.

Ice-shelf cavities generally can be classified as "warm," where water well above the local freezing point (e.g., CDW) flows beneath to cause strong (tens of m per year) melting, and as "cold," where such water is largely absent. With little heat introduced into cold cavities, relatively slow (centimeters to a few m per year) melting is primarily controlled by thermohaline processes related to sea-ice formation and by tidal mixing near the ice-shelf front, as illustrated in Fig. 1 (*6*). Little thinning is observed today on most ice shelves with cold cavities (Fig. 3), and some are thickening substantially (*20*). In contrast, most ice shelves with warm cavities are thinning (Fig. 3), and nearly all of these correspond to areas where the interior ice sheet also is thinning (*21*, *22*). Such thinning is strongest along Antarctica's Amundsen coast, where models and data indicate that recent shifts in the Amundsen Sea Low and an atmospheric Rossby wave response to tropical warming have caused enhanced offshore Ekman transport, increasing inflow of CDW to ice-shelf cavities (*9*, *10*) (Fig. 2C) and producing ~50% greater melt relative to the mid-1990s (*23*). Although a relatively strong correspondence between increased CDW flow and thinning of floating and grounded ice has been established observationally (*23*, *24*), less is known about the precise sensitivity of ice flow to the warmer waters and the full suite of processes and feedbacks contributing to thinning.

In the Antarctica Peninsula, glaciers have responded dramatically to the loss of buttressing as ice shelves have collapsed, which has largely been attributed to warmer atmospheric temperatures and surface melting (*14*, *15*). Along the Amundsen coast where the largest Antarctic contributions to sea level originate, the ice shelves have remained largely intact, while grounded ice flow has accelerated. In such regions, strong thinning (up to 10 m/year; Fig. 3) of thick (hundreds of meters) ice shelves likely has reduced buttressing by at most a few percent annually (*22*), which is too little to directly explain much of the observed speed-up. Instead, such gradual loss of buttressing may initiate a series of feedbacks, leading to far greater losses. For example, gradual loss of buttressing may induce moderate thinning and grounding-line retreat, removing basal traction to cause further speed-up, thinning, and grounding-line retreat, especially where the bed is prone to marine ice-sheet instability (*7*). In response, faster speeds may cause rifting (fracturing of the full ice thickness) or crevassing along ice-shelf margins, further reducing buttressing (*11*, *16*, *25*). Lastly, as the grounding line retreats, more ice is subjected to warm ocean temperatures (*24*).

Although we now have a reasonable understanding of the qualitative aspects of the ocean's interaction with large ice shelves and its key role in governing ice-sheet stability, the more quantitative understanding necessary to project sea level remains elusive. On the ocean side, high-resolution models are just beginning to resolve the details of circulation in ice-shelf cavities (*26*), which have largely remained unresolved at the scale of global climate models (GCMs). Recent work that does couple GCMs to regional-scale ice-ocean simulations suggests that the state-of-Texas–sized Filchner-Ronne Ice Shelf could transition from cold to warm by the end of the century, increasing basal melting by more than an order of magnitude (*26*). Such a change likely would lead to the ice shelf's rapid demise, which would remove buttressing and greatly increase outflow from several ice streams, potentially removing large marine portions of the West Antarctic Ice Sheet (*27*).

Once models providing accurate simulation and projection of ocean forcing are developed, they need to be coupled to ice-flow models to determine the ice-sheet response. Here, several obstacles remain. For example, grounding-line migration remains a challenge for numerical models (*28*); models tend to produce similar retreat but at vastly different rates (*7*). Much research has been discipline-focused, with rigorous coupling of ice-sheet, ocean, and climate models for long-term projection only recently becoming an area of active investigation. Until the completion of such fully coupled models that include realistic climate forcing, accurate projection of Antarctica's contribution to sea level will be highly uncertain.

Tidewater Glaciers

Although there are a few floating ice shelves at high northern latitudes, most ice discharge from Greenland occurs through narrow, fast-moving tidewater glaciers that calve directly to the ocean from termini with little (few km) or no floating extent (Fig. 3). Over the past two decades, many of these glaciers have sped up by more than 50% (*2*, *29*), coinciding with and likely linked to warmer waters in their respective fjords and nearby offshore locations (*30*–*33*) (Fig. 2D).

Both observation and modeling indicate that increases in speed include a major response to the loss of grounded and floating ice as glacier termini have retreated (*34*, *35*). Similar to loss of ice-shelf buttressing, retreat of a grounded terminus

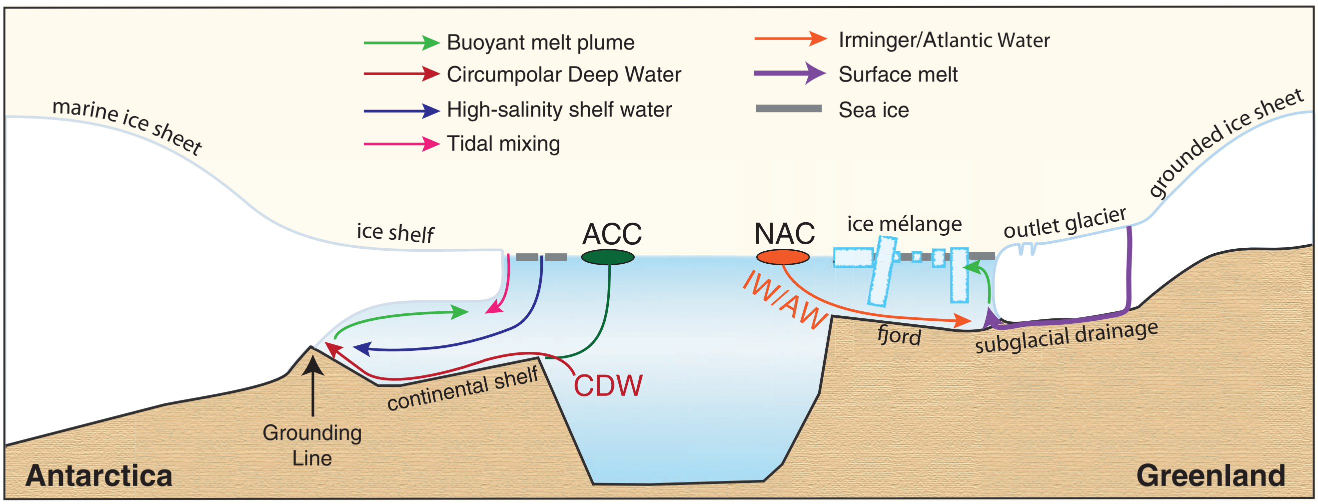

Fig. 1. Schematic illustration of the ocean processes influencing ice shelves and outlet glaciers that are described in the main text. Melting beneath Antarctic ice shelves occurs through a combination of three processes (*6*). The first occurs where dense, high-salinity shelf water is formed near the ice-shelf front during winter sea-ice growth. Although this water is at the surface freezing point, it can melt ice when it sinks to depths because it is above the local pressure melting point. The second occurs where tidal mixing moves seasonally warmed near-surface ocean water beneath the shelf front. Both of these processes are active for ice shelves with cold cavities. In contrast for warm ice shelf cavities, melting is dominated by the presence of a subsurface, warm water mass (CDW), originating from the Antarctic Circumpolar Current (ACC). Ocean melting of Greenland outlet glaciers is driven by analogous warm waters, namely the IW along the western and southeastern coasts and AW elsewhere. Both of these subsurface water masses originate with the North Atlantic Current (NAC). Where melting occurs, the buoyancy of the resulting meltwater plume produces positive feedback driving further melt, which may be enhanced further where subglacial meltwater is present (*38*).

removes downstream resistive stresses (e.g., ice-bed and sidewall traction) that must be redistributed upstream via increased strain rates (faster flow). Once a retreat begins, additional positive feedbacks such as steepening of surface slopes and reduction in effective pressure (the excess of the weight of the ice over basal water pressure) as ice thins may amplify the response (*36*).

How the ocean triggers initial retreat of a tidewater glacier front, with little or no ice shelf, is still poorly understood, but several potential contributing processes have been identified, many of which relate to enhanced melt (*37*). Where warmer ocean waters reach glacier termini, summer melt may occur at rates of up to 3 to 4 m/day (*38*, *39*). Rapid melting of ice cliffs requires vigorous turbulent flows to prevent formation of stable boundary layers that otherwise would partially insulate the ice from the warm ocean water. Such flows are enhanced in the summer by meltwater from the upper surface of the ice sheet, which drains to the bed to reemerge beneath the terminus and then rise buoyantly as depicted in Fig. 1 (*38*). Such meltwater-driven plumes also may affect any floating ice tongues that develop in regions of rapid surface melting. Thus, although water temperatures in fjords may vary little seasonally (*31*), melt rates may approximately double in the summer when driven by freshwater runoff (*38*). It is important to note that many fjords are choked with a mélange of sea ice and icebergs (Fig. 4) that can extend deep (>100 m) into the water column. Melting of such ice must consume some poorly known fraction of the oceanic heat that might otherwise produce melt at the terminus.

The degree to which melt influences a glacier depends on the extent to which its geometry exposes it to the ocean. For example, in western Greenland in the early 2000s (see location in Fig. 2), the collapse of Jakobshavn Isbræ's ~15-km-long floating ice tongue may have been the result of thinning brought on by high (~1 m/day) melt rates (*30*, *40*). In contrast, where glacier termini are grounded with little or no floating extension, far less area is exposed to warm waters. In such cases, melt rates (up to 4 m/day) are moderate relative to mean calving rates (~10 to 40 m/day). Whereas increased melting (by ~1 m/year) could cause gradual retreats of up to a few hundred meters per year, much of the observed retreat has occurred rapidly in association with enhanced calving rather than by gradual melt back (*35*, *41*). In such cases, increased melt may play a role in producing excess calving, perhaps through melting from tidally driven flow of warm water beneath the ice.

Another way that warmer fjord waters may contribute to glacier retreat and speed-up is through the influence of melt on the ice mélange choking many glacier fjords (Fig. 4), especially in winter. In summer, the mélange's constituent icebergs typically float freely, but in winter sea ice bonds them together to form a rigid mass that is pushed down the fjord by the advancing glacier terminus (*42*). Although the strength of this amalgam should have only a small effect on the back stress, observation and theory both indicate that it can suppress wintertime calving (*42*, *43*), which picks up again in spring and summer when the mélange remobilizes and clears the fjord. On Jakobshavn Isbræ, this seasonal variation of calving rates allows the terminus to advance over the winter and retreat during the summer, causing its speed to vary annually by 20 to 30% near the terminus. Warmer water in the fjord may shorten the period when the mélange is frozen, causing retreat by lengthening the duration of when strong calving occurs. For example, the breakup of Jakobshavn Isbræ's ice tongue coincided with a period of reduced sea-ice concentration in nearby Disko Bay (*42*).

Although warmer ocean temperatures likely influence glacier stability through the processes just described, the extent to which each of these or other unidentified processes contribute remains poorly characterized. Collecting oceanic observations near the calving front presents a major challenge, because at any moment skyscraper-size blocks could tumble over or rise from below, releasing seismic energy equivalent to a magnitude-5.0 earthquake (*44*). Furthermore, much of the area is perennially ice covered, and deep-draft icebergs scour fjord bottoms, making it difficult to emplace oceanographic instruments. Such factors have greatly impeded progress. As a result, a solid causal understanding of the strong correspondence between warmer ocean and atmospheric temperatures around Greenland and glacier retreat, calving, and acceleration is lacking, presenting a

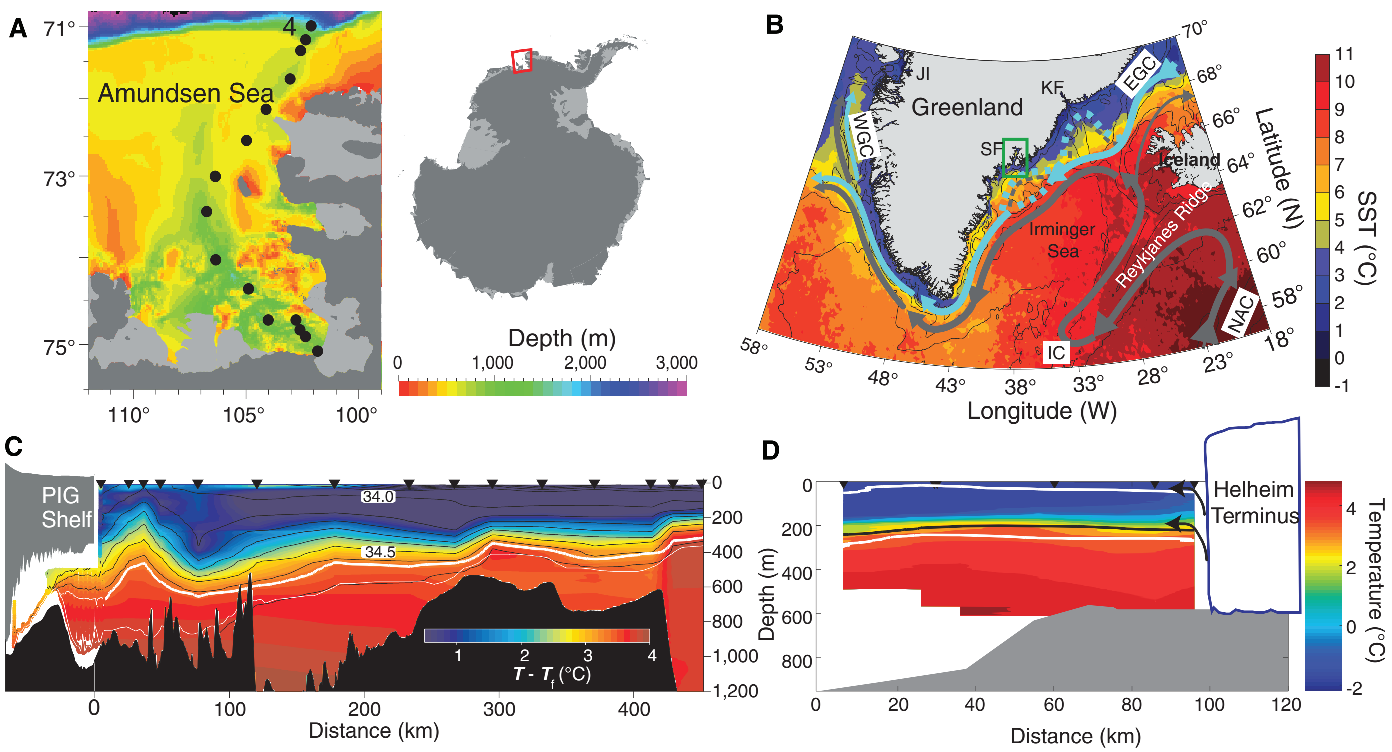

Fig. 2. Locations of temperature observations from (**A**) the Amundsen Sea, Antarctica, and (**B**) Sermilik Fjord (SF), Greenland. Also shown are the locations of Greenand's three largest glaciers along with their respective fjords: Helheim (SF), Kangerdlugssuaq (KF), and Jakobshavn Isbræ (JI). Color indicates water temperature (different scale for each location) in the proximity of (**C**) Pine Island Glacier and (**D**) Helheim Glacier. [Credit: Adapted by permission from Macmillan Publishers Limited: *Nature Geosciences* (*23*, *31*, *32*), copyright (2010 and 2011)].

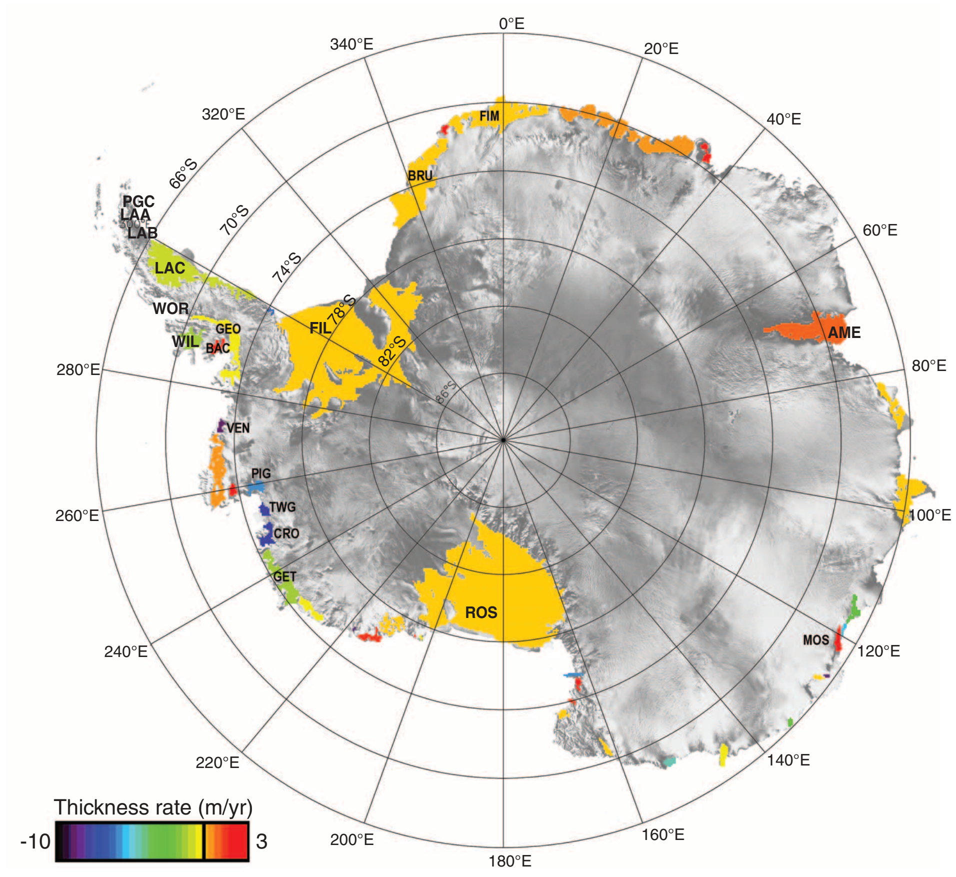

Fig. 3. Rate of thickness change of Antarctic ice shelves, with red indicating growth; green-purple, shrinkage; and yellow, stability. The rates were derived by Shepherd *et al.* (*20*) using radar altimetry. AME, Amery; MOS, Moscow University; ROS, Ross; GET, Getz; CRO, Crosson/Dotson; TWG, Thwaites Glacier; PIG, Pine Island Glacier; VEN, Venable Ice Shelf; FIL, Filchner; GEO, George IV; BAC, Bach; WIL, Wilkins; WOR, Wordie; LA A to C, Larsen A to C ice shelves; and PGC, Prince Gustav Channel. [Credit: Copyright 2010 American Geophysical Union (*20*). Reproduced by permission of American Geophysical Union.]

major interdisciplinary research challenge. Future progress likely will rely on the development of a new suite of innovative oceanographic and glaciological instrumentation, with an emphasis on automated and smart-sensing capabilities notably beyond what is currently available.

Paleo Perspective

The history of ice-sheet fluctuations confirms the strong influence of oceanic conditions on land-ice extent and volume. In particular, advance and retreat of the Greenland and Antarctic ice sheets appear to have been driven in important ways by the ocean, which at times may have been more important than atmospheric forcing. In Antarctica, it is clear that surface melting played almost no role in advance and retreat, and ice loss occurred as precipitation increased. Furthermore, differences in behavior between marine and terrestrial margins suggest that surface melting was not the only control in Greenland [e.g., (*45*)].

Sea level exerts an oceanic control on ice sheets. As often assumed, growth and shrinkage of the North American and Eurasian ice sheets may have contributed to long-term changes in Antarctica through sea-level–forced shifts in grounding-line position. Sea-level forcing, however, is generally quite slow compared with potential rates of forcing by warmer water [e.g., (*46*)], and several feedbacks tend to stabilize ice margins against rising sea level. One such feedback is sedimentation at grounding lines, which often can keep up with rates of sea-level rise (*46*). Furthermore, thinning of grounded ice in response to sea-level rise reduces the gravitational attraction of the ice for ocean water, as well as unloading the crust to cause isostatic rebound, reducing or reversing local sea-level rise (*47*). For example, extensive late-glacial raised beaches around Greenland document relative land uplift despite global sea-level rise (*48*). Thus, continued retreat during periods with local sea-level fall suggests that neither global nor local sea level was dominant in controlling the Greenland Ice Sheet's size. Similarly, trends in Greenland measured by Global Positioning Systems and other methods reveal little recent response to global sea-level rise (*49*). Thus, paleoclimatic data suggest that ice-sheet margins may have responded primarily to changes in ocean temperature or circulation.

Sea-floor sedimentary records on the Greenland and Antarctic continental shelves document a range of retreat behaviors from the Last Glacial Maximum that may involve ice-ocean interaction. The ice-flow response to such oceanic forcing likely is complex and may exhibit threshold behavior, with similar forcing producing very different short-term behavior even for rather similar ice margins (*46*). For example in some places, steady retreat left limited deposits that form numerous small moraines, which may be annual. Elsewhere, long-term stability (decades to millennia?) created large grounding-zone sedimentary wedges, punctuated by very rapid retreat that left essentially no grounding-zone deposits [e.g., (*50*)]. Retreat styles varied in different parts of Antarctica's Ross Sea [e.g., (*51*)] and among Greenland's drainages [e.g., (*52*)]. For example, adjacent drainages in the Ross Sea produced different numbers of grounding-zone wedges between intervals of rapid retreat (*53*).

A likely way that ice-ocean interaction has contributed to some threshold behavior is through its impact on ice-shelf thickness and extent. For example, warming-induced thinning of Jakobshavn Isbræ's floating ice tongue eventually led to its complete loss and faster terminus flow with no sustained regrowth of the tongue (*42*), despite the fact that the potential melting rates even with Irminger Water in the fjord would allow persistence of a reduced ice shelf capable of at least some buttressing. This is consistent with expectations from at least one proposed iceberg calving law (*54*) and suggests hysteresis with easier retreat than advance (*55*). The sedimentary record indicates that such behavior has occurred in the past, with an odd set of ridges in Pine Island Bay, Antarctica, likely recording the break-off of an extensive ice shelf during retreat across the continental shelf (*56*).

The enigmatic Heinrich events of the North Atlantic also may document ice-shelf break-off resulting from ice-ocean interaction. Because of the near-ubiquitous basal melting close to grounding lines, ice shelves tend to serve as debris filters, delaying (decades to centuries) iceberg formation while some or all of the entrained debris is melted free and deposited beneath the shelf (*57*). Accordingly, the sudden onset of extensive ice-rafted debris at distant locations such as occurred during Heinrich events suggests ice-shelf loss. Under this hypothesis, a rapid ice-shelf breakup, perhaps in response to warming of waters in the sub–ice-shelf cavity, creates icebergs from debris-rich areas near the grounding line that drift with much of their debris-load intact to the North Atlantic (*58*).

Although far from complete, the paleoclimatic record offers numerous indications that ocean-ice interaction has been important over the longer term and probably dominant in controlling marginal fluctuations of ice sheets through glacial

Fig. 4. Ice mélange (middle of photo) in front of the roughly 10-km-wide calving terminus of Jakobshavn Isbræ (see Fig. 2 for location), which lies between the two rock outcrops (just above the middle of the photo). Until the early 2000s, a several-hundred-meter-thick ice tongue existed in this now ~14-km-long mélange-covered area. [Photo credit: I. Joughin]

cycles as well as in present times. Gradual retreats have been observed, but stability punctuated by rapid retreat has been common, with hysteresis likely occurring, much as may be the case for many recent changes.

Outlook

Numerous observations indicate that ice-ocean interaction drives much of the recent increase in mass loss from both the Greenland and Antarctic ice sheets. Much less well understood are the details of the processes by which warmer waters in contact with glacial ice lead to greater ice loss, particularly for tidewater glaciers in Greenland. Although ice-ocean processes are at least qualitatively better understood in Antarctica, there are still numerous challenges to understanding the ice-sheet response to warmer ocean waters, in particular problems related to grounding line migration (*28*) and iceberg calving. Furthermore, little is known about future ocean forcing. Physical understanding indicates that bulk ocean warming would speed ice loss, but the recent observed changes have been driven less by such warming than by transport of already warm water into ice-shelf cavities and fjords with dimensions well below the resolution of most GCMs, indicating the need for finer-scale regional models. Where regional-scale ocean models have been coupled to GCMs, the results indicate the potential for far more extreme changes within this century than had been anticipated (*26*).

Although large uncertainties remain, tremendous progress has been made over the past two decades in identifying the role ice-ocean interaction plays in ice-sheet stability. The remaining challenges require a coordinated and sustained effort by glaciologists, oceanographers, and climate modelers before reliable projections of future sea level can be made. Until that time, Greenland and Antarctica will remain the "wild cards" in sea-level projections.

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Acknowledgments: We acknowledge the contributions from the papers cited in this manuscript and, just as important, the immense body of research that could not be cited because of space constraints. Comments by M. Maki and the anonymous reviewers improved the manuscript. The U.S. NSF supported I.J.'s (ANT-0424589), R.B.A's (ANT-0424589, ANT-1043528, ANT-0944286, and ANT-0909335), and D.M.H.'s (ARC-080639 and ANT-073286) effort. Additional support was provided by NASA for I.J. (NNX06A103G and NNX08AD64G), R.B.A. (NNX10AI04G), and D.M.H. (NNX08AN52).

10.1126/science.1226481