Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

ANNUAL REVIEWS Further
Click here for quick links to Annual Reviews content online, including:
- Other articles in this volume
- Top cited articles
- Top downloaded articles
- Our comprehensive search

Oceanic Forcing of Ice-Sheet Retreat: West Antarctica and More

Richard B. Alley,[1] Sridhar Anandakrishnan,[1] Knut Christianson,[2,3] Huw J. Horgan,[4] Atsu Muto,[1] Byron R. Parizek,[5] David Pollard,[1] and Ryan T. Walker[6,7]

[1]Department of Geosciences and Earth and Environmental Systems Institute, The Pennsylvania State University, University Park, Pennsylvania 16802; email: rba6@psu.edu

[2]Department of Earth and Space Sciences, University of Washington, Seattle, Washington 98195

[3]Courant Institute of Mathematical Sciences, New York University, New York, NY 10012

[4]Antarctic Research Centre, Victoria University of Wellington, Wellington 6140, New Zealand

[5]Mathematics and Geoscience, The Pennsylvania State University, DuBois, Pennsylvania 15801

[6]Earth System Science Interdisciplinary Center, University of Maryland, College Park, Maryland 20742

[7]Cryospheric Sciences Laboratory, NASA Goddard Space Flight Center, Greenbelt, Maryland 20771

Annu. Rev. Earth Planet. Sci. 2015. 43:207–31

First published online as a Review in Advance on January 30, 2015

The *Annual Review of Earth and Planetary Sciences* is online at earth.annualreviews.org

This article's doi:
10.1146/annurev-earth-060614-105344

Copyright © 2015 by Annual Reviews.
All rights reserved

Keywords

Antarctica, Greenland, sea level, ice sheet, stability

Abstract

Ocean-ice interactions have exerted primary control on the Antarctic Ice Sheet and parts of the Greenland Ice Sheet, and will continue to do so in the near future, especially through melting of ice shelves and calving cliffs. Retreat in response to increasing marine melting typically exhibits threshold behavior, with little change for forcing below the threshold but a rapid, possibly delayed shift to a reduced state once the threshold is exceeded. For Thwaites Glacier, West Antarctica, the threshold may already have been exceeded, although rapid change may be delayed by centuries, and the reduced state will likely involve loss of most of the West Antarctic Ice Sheet, causing >3 m of sea-level rise. Because of shortcomings in physical understanding and available data, uncertainty persists about this threshold and the subsequent rate of change. Although sea-level histories and physical understanding allow the possibility that ice-sheet response could be quite fast, no strong constraints are yet available on the worst-case scenario. Recent work also suggests that the Greenland and East Antarctic Ice Sheets share some of the same vulnerabilities to shrinkage from marine influence.

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

INTRODUCTION

Almost five decades have passed since John Mercer's (1968) first warning that "industrial pollution of the atmosphere" or other causes could warm the "uniquely unstable and vulnerable" West Antarctic Ice Sheet enough to remove its ice shelves, causing "disintegration into the sea" of the marine portions (those resting on bedrock well below sea level) and raising sea level globally at a rate that "would be rapid, perhaps even catastrophic." Recent studies (Joughin et al. 2014a, Rignot et al. 2014) suggest that oceanic processes have already forced a threshold-crossing committing the ice to the deglaciation identified by Mercer. If this is correct, then despite the great progress in research since Mercer's time, the loss of the ice sheet was triggered before the threshold was properly quantified and thus before scientific knowledge could be used to design efficient policies relating to survival or loss of the ice sheet. The possibility remains, however, that the ice sheet has not been pushed past its survival threshold, and even if it has, the resulting lessons may inform work on the Greenland and East Antarctic Ice Sheets.

Here, we synthesize recent data and model results on ice-ocean interactions. We focus on Thwaites Glacier, the part of the West Antarctic Ice Sheet that is perhaps most likely to trigger a collapse, but attempt to provide perspective and draw on relevant related advances from the East Antarctic and Greenland Ice Sheets as well as past ice sheets. We show evidence that oceanic forcing has dominated past changes for the West and East Antarctic Ice Sheets and large parts of the Greenland Ice Sheet and older ice sheets. We also show that well-constrained projections will remain difficult, in part because of threshold behavior of the ice-sheet system.

MARINE ICE SHEETS CAN EXHIBIT A WELL-UNDERSTOOD INSTABILITY

Mass gain for an ice sheet occurs primarily by snowfall on its upper surface. Mass loss occurs primarily by melting of the upper surface in warmer regions (at lower elevation or latitude), by melting beneath attached floating ice shelves, and by calving of icebergs that drift away to melt elsewhere. Direct mass gain or loss by freezing or thawing beneath nonfloating ice is usually minor. An ice sheet spreads under gravity through deformation in the ice, and may be frozen to its substrate or slide over that substrate lubricated by meltwater or deformation in subglacial till (unconsolidated sediment) (see Cuffey & Paterson 2010 for additional background).

Although assessing and projecting changes in surface melting and snowfall involve notable uncertainties, possible changes in ice flow pose greater difficulties (e.g., Joughin & Alley 2011, NRC 2013). And, although changes in basal lubrication can be important, loss of ice-shelf buttressing is probably the dominant process triggering large, rapid shrinkage of ice sheets. Ice-shelf flow is almost always restrained by friction with fjord walls, adjacent ice, or local high spots in the seafloor, which in turn causes ice shelves to restrain the flow of their tributary ice streams. The upper surfaces of ice shelves are relatively low, warm, and prone to melting compared with other parts of an ice sheet, and sufficient meltwater can wedge open crevasses to cause iceberg calving, generating very rapid ice-shelf loss (e.g., Scambos et al. 2004, Cook et al. 2014).

More importantly in many cases, ice shelves are at the melting point at their base, and in general will thin in response to any oceanic warming. The melt-rate sensitivity depends on many factors with important dynamical controls. A linear approximation to observations gives a sensitivity of order 10 m/yr/°C for many existing ice shelves (e.g., Rignot & Jacobs 2002, Shepherd et al. 2004), although over a sufficiently broad range of conditions such melting is expected to increase as the square of ocean temperature (Holland et al. 2008b).

Ice flow from the grounding line (where flotation starts) to the calving front (where icebergs form) commonly takes years to centuries, and ice shelves are typically on the order of hundreds of meters thick at the front, so even a small temperature change can have a large influence on ice-shelf thickness. As an example, warming of 1°C that removes 10 m/yr from a 10-yr-long ice shelf would thin a 100-m ice front to zero. And, because thicker ice has more friction at the sides (e.g., Dupont & Alley 2005) and more interaction with local seafloor highs (pinning points), even a small warming can notably reduce this frictional buttressing, especially if applied near the grounding line (Walker et al. 2008), allowing faster flow of the nonfloating ice that feeds an ice shelf.

In the absence of friction from the sides or pinning points, the spreading rate of an ice shelf increases with approximately the third power of the ice thickness (Weertman 1957). The ice velocity at the front, and thus the calving flux for a steady ice shelf (Alley et al. 2008b), is the velocity crossing the grounding line plus the added speed from spreading of the floating ice; thus, a relatively deep grounding line can produce a very large iceberg flux.

The increase in spreading rate with ice thickness gives rise to the marine-ice-sheet instability for any freely spreading ice shelf with a grounding line on a bed that deepens toward the ice-sheet center (an overdeepened bed) (Weertman 1974, Schoof 2007). To understand this instability, consider the response if the grounding line is perturbed toward the center of the ice sheet (the logic is the same for the opposing perturbation, with all signs reversed). Ice that had been grounded becomes ungrounded, losing basal friction and allowing faster spreading for the thicker ice at the new grounding-line position. Because there is not an immediate and counteracting change in ice supply from the adjacent nonfloating ice, the enhanced spreading causes thinning and additional flotation from the sloping bed. Ice that had been too thick to float then progressively thins and floats, raising sea level. Ice-shelf buttressing can stabilize an otherwise unstable configuration, but then reduction in ice-shelf drag can cause unstable retreat across an overdeepened bed. Note also that thinning or loss of a constrained ice shelf is expected to contribute to sea-level rise by causing faster flow of nonfloating ice even on a horizontal bed or one that rises inland, but without the instability of the reversed bed.

Discovery of the very deep central basins beneath the West Antarctic Ice Sheet [to >2,500 m below sea level (Bentley & Ostenso 1961; also see Fretwell et al. 2013)] (**Figure 1**) quickly led Bentley & Ostenso (1961) to suggest that the ice sheet formed by merger and then thickening of ice shelves flowing from adjacent highlands; these authors did not provide modeling support for this conclusion but likely based it on the widespread occurrence of ice shelves and on the implausibility of sufficiently high ice cliffs advancing across deep basins. Such an origin for the West Antarctic Ice Sheet suggested that ice-shelf melting could reverse this process (Robin 1958, Hollin 1962, Mercer 1968). Numerical modeling (e.g., Thomas & Bentley 1978, Pollard & DeConto 2009) indicates that persistence of the modern West Antarctic Ice Sheet relies on the stabilizing influence of its ice shelves [aided by additional stabilizing influences (e.g., Gomez et al. 2010) as described below]. Portions of the East Antarctic and Greenland Ice Sheets also have marine character (e.g., Fretwell et al. 2013, Morlighem et al. 2014) (**Figure 1**), as did parts of former ice sheets, but the deeper, more extensive basins make the West Antarctic Ice Sheet more prone to marine influences than any other extant ice sheet and at least most paleo–ice sheets.

As discussed next, several lines of evidence show that ice-sheet stabilization by ice shelves has been highly variable over time. We consider history, processes, and possible future changes.

THE WEST ANTARCTIC ICE SHEET LIKELY DEGLACIATED FROM THE MARINE-ICE-SHEET INSTABILITY IN THE GEOLOGICALLY RECENT PAST

Many lines of evidence indicate deglaciation during the previous interglacial [the Eemian or Sangamonian, or Marine Isotope Stage (MIS) 5e, ∼130 ka], or at other times since ∼1 Ma or so.

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

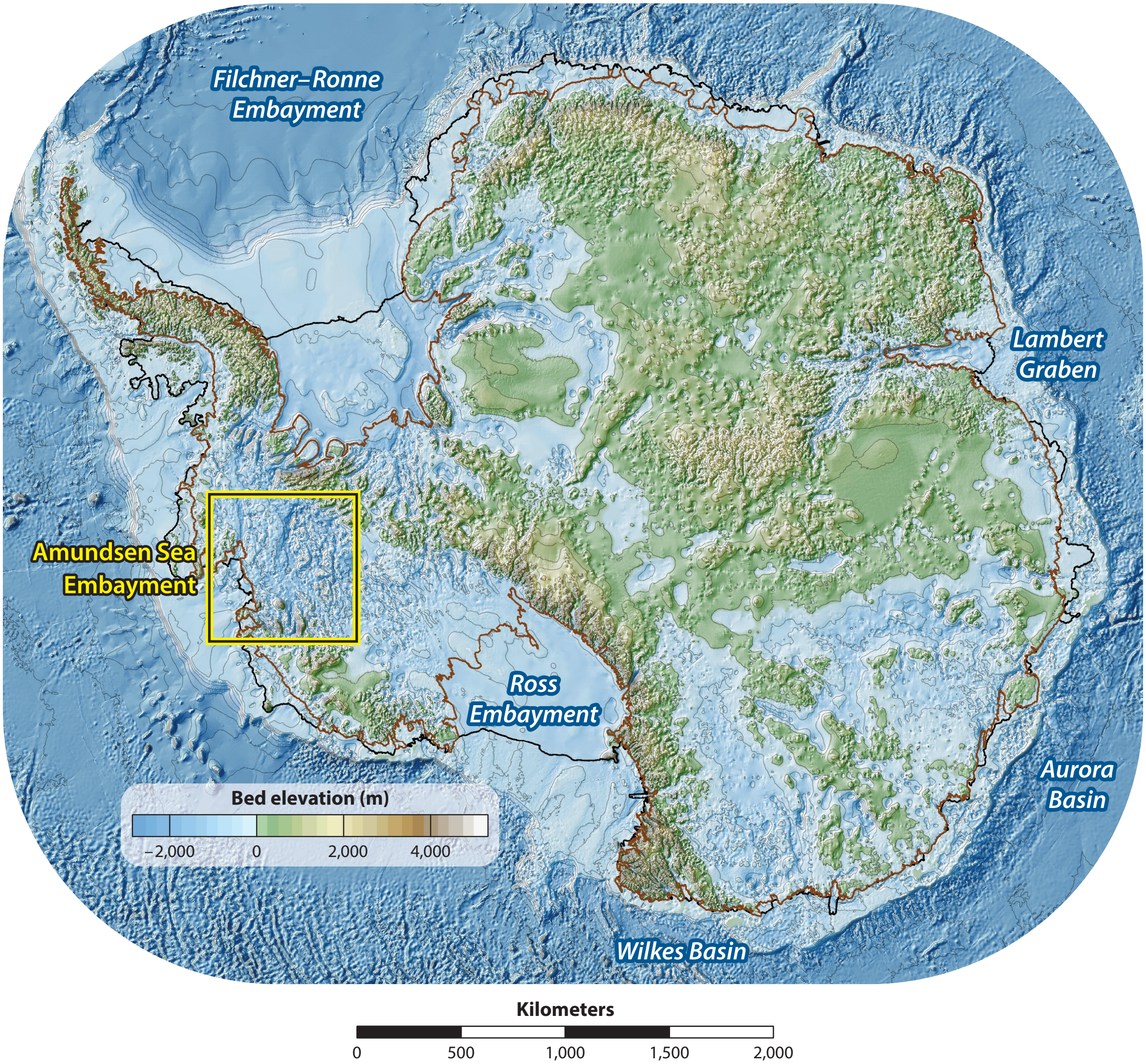

Figure 1

Bed topography (bathymetry) of Antarctica (contour interval is 500 m) with major marine basins labeled (Fretwell et al. 2013). Areas shaded in blue are marine based (below eustatic sea level). The ice-sheet grounding line (*brown*; Depoorter et al. 2013) and current ice-sheet extent (*black*; Depoorter et al. 2013) are plotted. The yellow box indicates the Amundsen Sea Embayment, shown in **Figure 2**. Data from Bedmap2 (see Fretwell et al. 2013); figure prepared using GMT and QGIS.

Unequivocal evidence of recent West Antarctic Ice Sheet/Thwaites Glacier deglaciation has not been found, however.

Mercer (1968) may have been the first to note that the volume of the West Antarctic Ice Sheet would explain much of the sea-level high stand from MIS 5e, which is recorded in many ways, including by old corals or other marine deposits found above modern sea level on islands. Numerous studies since then, especially globally and gravitationally consistent simulations of past sea-level change assimilating the available data, have greatly strengthened the evidence for West Antarctic Ice Sheet loss in MIS 5e. Sea level near an ice sheet is raised by meters or more because of the gravitational attraction between the ice and surrounding water. Loss of an ice sheet raises global average sea level, but this may be more than offset near the shrinking ice sheet by relaxation of the formerly attracted water; sea-level rise in general increases with increasing distance from a shrinking ice sheet. Fingerprinting of past sea-level high stands using a geographically dispersed suite of proxy records then can reveal the source(s) of the water. Kopp et al. (2009) found that the MIS 5e high stand required major contributions from both poles, with 95% confidence that each pole contributed at least 2.5 m of sea-level equivalent to a total rise of at least 6.6 m. The southern contribution is consistent with the ~3.3 m of sea-level equivalent in the marine portions of the West Antarctic Ice Sheet (Bamber et al. 2009b), although the fingerprinting lacked the spatial resolution to distinguish East from West Antarctic Ice Sheet sources. Additional studies using similar techniques also have found strong evidence of a large southern source consistent with West Antarctic Ice Sheet deglaciation (Dutton & Lambeck 2012, O'Leary et al. 2013).

Several other lines of evidence also indicate geologically recent deglaciation of the deep West Antarctic Ice Sheet basins, although some are less specific about timing. Diatoms recovered from sediments beneath Whillans Ice Stream, on the Siple Coast of the West Antarctic Ice Sheet inland of the Ross Ice Shelf, indicate open water there sometime since 750 ka (Scherer et al. 1998). Farther downstream at the ANDRILL site, sediments indicate open water multiple times in the geologically recent past (Naish et al. 2009). Coordinated ice-sheet modeling indicates that deglaciation there would have led to deglaciation of all the marine basins (Pollard & DeConto 2009). More generally, deglaciation of any large part of the West Antarctic Ice Sheet's deep basins likely would deglaciate all of them (Pollard & DeConto 2009); for example, deglaciating just the modern drainage into Pine Island Bay, which would raise modern sea level ~1 m, would leave a cliff that in places was >3 km high and extended >1 km above sea level, which likely would be very unstable (Bassis & Walker 2012; discussed further below).

Ice-core data from East Antarctica show anomalous warmth (compared with the current Holocene interglacial) during previous recent interglacials, especially during MIS 5e. This is most easily (although not uniquely) explained by a greatly reduced West Antarctic Ice Sheet during the anomalously warm times (Holden et al. 2011).

The isotopic composition of the deepest ice at Siple Dome is consistent with a source at high elevation on one of the nonmarine blocks of the West Antarctic Ice Sheet (Alley & White 2000). Grounding of an ice shelf is the most likely way (although not the only way) to have transported that ice to the site, with subsequent growth of the dome occurring above that grounded ice.

Further evidence for geologically recent deglaciation comes from the modern West Antarctic Ice Sheet coast, where closely related bryozoans (Barnes & Hillenbrand 2010, Vaughan et al. 2011) and octopuses (Strugnell et al. 2012) are geographically separated by less related types. However, the closely related species would be in proximity through seaways of a deglaciated West Antarctic Ice Sheet, suggesting ice loss within evolutionarily recent time [Vaughan et al. (2011) favor a loss during MIS 5e].

Farther seaward, the inner continental shelf of Pine Island Bay, in front of the Thwaites and Pine Island Glaciers, shows geomorphic features interpreted as channels, incised to >100 m local relief in water depths of more than 1,200 m, with some channels rising away from the ice sheet (Nitsche et al. 2013). Steady-state water drainage beneath an expanded West Antarctic Ice Sheet appears inadequate to explain such features. Instead, they have been interpreted as the products of outburst floods from water trapped subglacially following reestablishment of the ice sheet by ice-shelf bridging of the deep basins inland of Thwaites Glacier (Alley et al. 2006). Bentley &

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

Ostenso's (1961) proposal that the ice sheet formed by thickening of an ice shelf bridging the deep basins suggests the possibility that the trapped water formed saline subglacial lakes that escaped catastrophically rather than gradually; if the escape was through Pine Island Bay beneath grounded ice, features such as those observed could have been formed (Alley et al. 2006).

In contrast to the evidence for deglaciation, there do not seem to be strong arguments for long-term continuous persistence of the West Antarctic Ice Sheet in the deep marine basins. Cold ice did survive MIS 5e on Mt. Moulton (Korotkikh et al. 2011) on the Marie Byrd Land highland, but that region is not expected to deglaciate with the deep basins (Holschuh et al. 2014; also see Bradley et al. 2012). A Ross Ice Shelf collapse at MIS 5e is not recorded in the McMurdo Sound ANDRILL core; however, core recovery is poor in this interval, making this result unsurprising (Naish et al. 2009).

Thus, although additional information would be highly useful, the available data are most consistent with deglaciation of the deep basins of the West Antarctic Ice Sheet, with ~3 m of sea-level rise, during geologically recent time(s). In particular, deglaciation is suggested during MIS 5e, at ~130 ka.

Similar evidence of shrinkage notably below modern size is lacking for the East Antarctic Ice Sheet during MIS 5e, but there are suggestions in offshore sedimentary records of older ice loss from the East Antarctic Ice Sheet. As discussed below, discharge of debris-laden icebergs to the ocean likely indicates that no ice shelf is present, because ice shelves generally melt beneath, removing debris from basal ice. Pulses of iceberg-rafted debris during the Pliocene (between 5.3 and 3.3 Ma) and earlier along the East Antarctic coast, with sources in marine regions now buttressed by ice shelves, are suggestive of notable ice loss (e.g., Cook et al. 2013).

The review by Alley et al. (2010) and the newer results of Dahl-Jensen et al. (2013) summarize some of the extensive evidence that the Greenland Ice Sheet survived MIS 5e, but in a notably reduced state. Furthermore, limited evidence indicates complete or nearly complete loss of the ice sheet sometime after it formed, likely during MIS 11, at ~400 ka (Alley et al. 2010; also see Reyes et al. 2014). Far-field indications of sea level are consistent with loss of the West Antarctic and Greenland Ice Sheets, but little additional shrinkage of the East Antarctic Ice Sheet, during MIS 11 (Raymo & Mitrovica 2012). Still older data, from the Pliocene, include notably larger uncertainties, but point to enough sea-level rise (Rovere et al. 2014) to require complete loss of the Greenland Ice Sheet and the marine portions of the West Antarctic Ice Sheet, likely plus a few more meters of sea-level rise from the East Antarctic Ice Sheet. Thus, ice sheets shrank during past warm times.

AVAILABLE PALEOCLIMATIC DATA ARE CONSISTENT WITH PAST WEST ANTARCTIC ICE SHEET COLLAPSE AS WELL AS SLOWER DEGLACIATION

Mercer (1968) cited the geological evidence of West et al. (1960) (incorrectly cited as West & Sparks 1961) showing that "rise in sea level to above present levels took place during the Sangamon Hypsithermal [MIS 5e] and was very rapid," which "suggests possible catastrophic disintegration of the West Antarctic Ice Sheet at that time, but further evidence is needed." Above we summarize some of the more recent data on that rise above present levels, but newer data also provide insight on the rate of rise. Other than adding many better-dated and more widely distributed records to the citation list, a follower of Mercer could write the same sentence today. The available data allow, but do not require, sea-level rise then at rates much higher than the recent ~3 mm/yr (0.3 m/century), consistent with catastrophic disintegration or collapse (inferences from models and physical processes are discussed in later sections).

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

During at least some times in the past, sea-level rise clearly was much faster than the ongoing rate. Ice-age terminations involved sea-level rise of >100 m in ~10,000 yr—an average of 1 m/century, but with faster events reaching 4–5 m/century (e.g., Rohling et al. 2013). However, Earth had much more ice during those rapid rises than it does now, allowing processes not likely to act as rapidly today, such as widespread warming causing rapid surface melting of midlatitude ice sheets.

The MIS 5e sea-level records cited above provide more-useful constraints on rates of sea-level rise when ice volumes were similar to or smaller than modern. These records are interpreted with increasing confidence to indicate the large rapid rise during the main part of the deglaciation as the Laurentide and Fennoscandian Ice Sheets and other ice melted, as well as a second rapid rise late in the warm period from a baseline higher than modern. Focusing only on the latter part of MIS 5e with sea level above modern, Thompson et al. (2011) found a peak rate of rise >2.5 m over a 1-kyr interval (>0.25 m/century), and Kopp et al. (2013) similarly found 3–7 m/kyr (0.3–0.7 m/century). But, these studies lacked the time resolution to identify any shorter-lived, faster rise embedded within the 1-kyr interval of peak rise. O'Leary et al. (2013) found a rise of ~3 m over ~500 yr (0.6 m/century) early in MIS 5e and a later rise of ~6 m over ~1 kyr (also 0.6 m/century), but the dating uncertainties allow shorter as well as longer times for the rises. Rohling et al. (2008) used a smaller data set based on a novel sea-level indicator from the Red Sea with higher time resolution and, after applying a 750-yr Gaussian smooth, obtained for times with sea level above modern a peak rate of sea-level rise of 2.5 m/century averaged over ~300 yr. Again, in light of the uncertainties and the Gaussian smooth, faster or slower sea-level rise could be consistent with the data.

We do not know of any data that preclude with high confidence a West Antarctic Ice Sheet collapse in a century or less during MIS 5e or other, earlier interglacials. But, in light of Rohling et al.'s (2008) results, and their agreement with other records noted above, the most consistent interpretation is that large and rapid ice-sheet change from conditions similar to modern has caused peak sea-level rise much faster than recently observed. Further data testing this hypothesis are warranted.

OCEANIC FORCING IS ESPECIALLY IMPORTANT IN CAUSING RAPID ICE-SHEET CHANGES

The ice sheets have been losing mass at an accelerating rate over the past two decades, most recently contributing to sea-level rise at a rate slightly greater than 1 mm/yr (Vaughan et al. 2013). The loss is primarily from the Greenland Ice Sheet, the Antarctic Peninsula, and the West Antarctic Ice Sheet. In Greenland and the Antarctic Peninsula, the runoff produced by enhanced surface melting contributes to some of the mass loss (roughly half in Greenland; Cazenave & Le Cozannet 2014). Coastal thinning, from enhanced melting as well as other causes, tends to steepen the ice just inland, slightly enhancing the flow there in a fairly well characterized manner that diffuses slowly inland (e.g., Alley & Whillans 1984). Here we focus on additional accelerations of grounded ice due to the loss of ice-shelf buttressing.

We note that loss of friction beneath nonfloating ice through enhanced lubrication also can cause faster ice flow, contributing to faster iceberg calving and sea-level rise. The spectacular meltwater-lake drainage events through Greenland (e.g., Das et al. 2008) primarily feed channelized rather than distributed drainage systems, which have a relatively small effect on lubrication and ice flow (e.g., Parizek & Alley 2004, Shannon et al. 2013). The main potential for such lubrication to affect ice flow is if surface meltwater drainage to the bed migrates inland (Leeson et al. 2015) over regions where the ice is now frozen to its bed; the resulting rapid thawing would

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

notably enhance ice motion and sea-level rise, although without the sort of very rapid rates that seem possible from West Antarctica (Parizek & Alley 2004).

In contrast, several recent examples show the power of ice-shelf loss to enhance ice flow. At Jakobshavn Isbræ, Greenland, ∼1°C warming of water in front of the fjord led to ice-shelf thinning and then complete break-off, causing the flow velocity to more than triple (Holland et al. 2008a, Motyka et al. 2011, Joughin et al. 2014b). Along the Antarctic Peninsula, very rapid loss (over a few weeks) of the Larsen B Ice Shelf from a combination of increased basal melting (Shepherd et al. 2003) and increased surface melting forming lakes that wedged open crevasses led to acceleration of the nonfloating glaciers feeding it, by up to 6-fold (MacAyeal et al. 2003, Scambos et al. 2004). The acceleration and thinning of Pine Island Glacier, Thwaites Glacier (**Figure 2**), and nearby glaciers draining into the Amundsen Sea (Sutterley et al. 2014), which dominate the mass loss from the West Antarctic Ice Sheet, appear to have resulted from ice-shelf shrinkage in response to a wind-driven increase in the circulation of warm, and probably warming, Circumpolar Deep

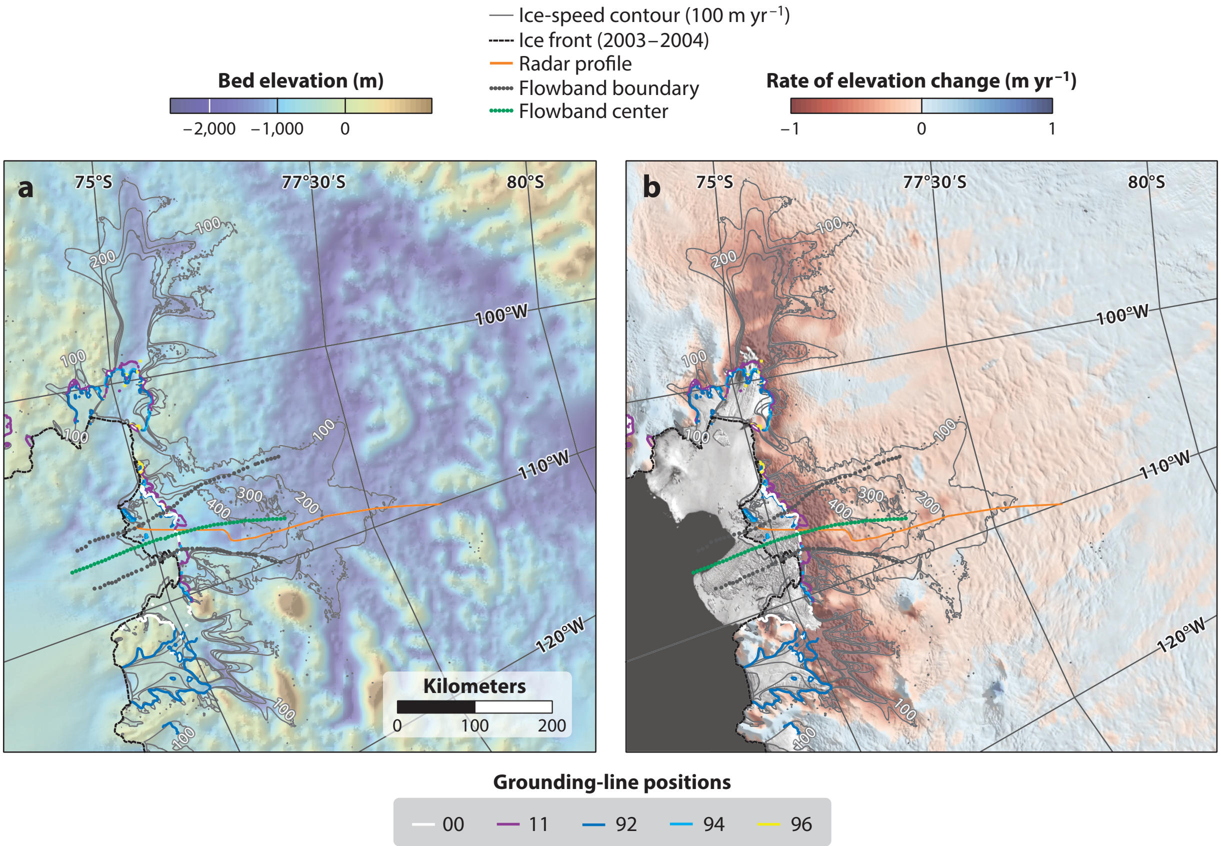

Figure 2

(*a*) Bed topography (bathymetry) of the Amundsen Sea Embayment (Fretwell et al. 2013) and (*b*) rate of ice-sheet elevation change (2003–2009) from ICESat GLAS laser altimetry. Grounding lines and ice-speed contours are from interferometric synthetic aperture radar data (Rignot et al. 2011, 2014). Ice-front position is from MODIS MOA (Haran et al. 2005). The modeled ice-sheet flowband (Parizek et al. 2013) is shown in **Figure 5**. The radar profile is shown in **Figure 4**. Data from Bedmap2 (see Fretwell et al. 2013); figure prepared using GMT and QGIS.

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

Water onto the continental shelf and beneath the ice shelves (e.g., Vaughan et al. 2013, Schmidtko et al. 2014).

As discussed by, for example, Alley et al. (2007), the biggest controls on ice-sheet extent are probably snowfall minus melting of the upper surface for land-ending glaciers, and temperature of ocean water at the grounding lines for marine-ending glaciers (also see Carlson & Winsor 2012). Warming increases saturation vapor pressure in air, and all else being equal this increases precipitation and thus snowfall in colder places; however, melting typically increases more than snowfall (e.g., Denton et al. 2005). Hence, for land-ending ice, and for some marine-ending ice, warming causes shrinkage because of increased melting on the upper surface. The old idea that warming grows ice (e.g., Scott 1905 for Antarctica, Ewing & Donn 1956 for northern ice) is counter to observations. Ice has difficulty advancing into water that is too deep, and sea-level fluctuations have been invoked as controlling ice-sheet behavior, with, for example, the Northern Hemisphere changes affecting Antarctica (Penck 1928). However, as discussed by Alley et al. (2007), the generally slow nature of large sea-level changes compared with the rapid impact of ocean-temperature changes, and the many stabilizing feedbacks that can offset sea-level rise, means that changing sea level is unlikely to exert a strong control on ice-sheet size in most cases (further discussed below). Simply as a scaling, above we showed how a 1°C warming can thin the 100-m front of an ice shelf to zero in a decade or less, greatly reducing its frictional resistance to ice flow, whereas the 100-m sea-level rise associated with ice-age terminations took ~10,000 yr.

The review by Alley et al. (2010) for Greenland noted that the ice sheet grew as precipitation decreased, and shrank as precipitation increased. Furthermore, some of the ice-sheet changes were counter to local sea-level forcing, with advance in a rising sea level and retreat in a falling sea level (many retreats were observed with rising sea level during warming, and advances with falling sea level during cooling, so in those cases one cannot separate the controlling climatic factor without model experiments). We know of no cases in which a sea-level change has been demonstrated with high confidence to control ice-sheet extent, whereas many cases of temperature control are known.

The reader should bear in mind that glaciers and ice sheets are sufficiently complex and diverse that there are likely to be exceptions to any generalization. If atmospheric warming causes a shift in storm tracks, some places may experience a greater rise in accumulation than in melting, especially if the warming is primarily in winter. Surging associated with changing basal lubrication can cause advance with no climatic forcing, and kinematic waves from changing surface mass balance can cause delayed marginal changes that may be counter to ongoing forcing. Especially rapid sea-level rise may exceed stabilizing feedbacks and force retreat; the question remains interesting whether melting of Northern Hemisphere ice sheets during the fastest large rise of the most recent deglaciation, Meltwater Pulse 1A, could have been fast enough to destabilize Antarctic grounding lines. Cuffey & Paterson's (2010) text is an excellent starting point for learning about these and additional complexities. Furthermore, these less common behaviors are often the most interesting research topics. Physical understanding, models, and ice-sheet history point to ice growth with cooling, ice shrinkage with warming, relatively slower changes controlled by air temperature for land-ending glaciers, and sometimes faster changes controlled by water temperature for marine-ending glaciers.

The enigmatic Heinrich (H) events of the North Atlantic might at first seem to provide a challenge to this simple summary, but they probably actually support it. During ice ages, millennial Dansgaard–Oeschger (DO) oscillations linked to sometimes-abrupt shifts in North Atlantic sea ice and ocean circulation had large effects on regional climates in much of the world (reviewed in Alley 2007). Within only some of the cold intervals that followed, a prominent layer of iceberg-rafted debris sourced from Hudson Strait was deposited on the seafloor of parts of the North Atlantic at the same time that widespread additional climate anomalies occurred (Hemming 2004). Surging of

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

the ice stream in Hudson Strait may have been involved (MacAyeal 1993); however, the occurrence of all of these H events after surface cooling suggests triggering.

Because ice shelves almost always experience net basal melting, an ice shelf serves to reduce iceberg-rafted debris, and ice-shelf loss allows icebergs to drift away before they lose their debris during sub-ice-shelf melting (e.g., Joughin et al. 2012a). Evidence from sediment cores and models (Marcott et al. 2011) shows that the surface cooling of DO oscillations led to widespread subsurface warming in the northwestern Atlantic, which would have increased melting beneath a Hudson Strait ice shelf. Loss of the ice shelf (see below) would have allowed both faster ice flow and delivery of debris-laden icebergs to the North Atlantic. If the ice stream froze to its bed following an interval of rapid flow, and then took several millennia to regrow and thaw at its bed, as in MacAyeal's (1993) model, then the behavior of skipping several DO surface coolings before another H event would be expected. Thus, although the H events first appeared to challenge our understanding of the causes of ice-sheet fluctuations, with cooling-induced ice-sheet mass loss, further research now suggests that warmer ocean waters at the grounding line were responsible. [The actual impact of H events on sea level remains difficult to determine owing to dating uncertainties (Siddall et al. 2008), although the most recent results suggest that sea level varied with northern temperature and thus that the H events were preceded by falling, not rising, sea level (Frigola et al. 2012).]

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

OCEANIC FORCING ACTS THROUGH MULTIPLE PROCESSES

The literature on ocean-ice interactions is growing very rapidly. Some of the brief overview that follows is based on a review by Joughin et al. (2012a).

The coldest waters produced in large amounts in the world ocean are the brines formed in contact with growing sea ice. The salt rejected during freezing makes the water dense enough to sink to the seafloor, which often deepens toward ice-sheet grounding lines because of past glacial erosion or glacial isostatic depression of the seafloor. Because of the pressure dependence of the melting point, water that started in equilibrium with ice at the surface will become warmer than equilibrium with increasing depth, more or less linearly with pressure by roughly 1°C for 1,300 m (Jenkins & Holland 1999). This water (often called High Salinity Shelf Water in Antarctica) thus causes melting where it contacts ice at depth. The meltwater released lowers the seawater density, often forming a buoyant plume that rises along the base of the ice shelf. Entrainment of adjacent waters into the buoyant plume typically drives further melting so long as the plume remains in contact with the base of the ice (a vigorous plume that entrains too much warmer and saltier water can lose its buoyancy). Under certain conditions, the pressure dependence can cause a plume that initially melts the ice shelf to supercool at shallower depth and grow ice that accretes to the base of the ice shelf, but even such ice shelves typically experience net basal melting (e.g., Jenkins & Doake 1991). However, under such High Salinity Shelf Water–type "cold" conditions, basal-melt rates of ice shelves are typically a few meters per year or less (Joughin et al. 2012a). Glaciological estimates of basal mass balance confirm this general pattern of melting and refreezing, with melting near grounding lines and freezing in the interior of large ice shelves (e.g., Rignot et al. 2013).

Almost all other waters in the world ocean are warmer than High Salinity Shelf Water, but often saltier and denser. Where these "warm" waters reach the grounding lines of ice sheets, melt rates can be tens of meters per year. Wind-driven processes may be highly important in controlling whether a sub-ice-shelf cavity is "cold" or "warm," and shifting winds or other forcing may cause a rapid switch from cold to warm (or warm to cold) (e.g., Dutrieux et al. 2014), which in turn can greatly and rapidly change the flow of tributary ice, thus affecting sea level.

This process understanding, plus recent results on ocean circulation, provides a consistent interpretation for ice-age control of the East Antarctic, West Antarctic, and perhaps Greenland

Ice Sheets by sub-ice-shelf melting. As argued by Adkins (2013), around Antarctica today the main "warm" source for ice-shelf melting is Circumpolar Deep Water, which is slightly modified from North Atlantic Deep Water. This water drives melting of ice shelves, and is freshened as a result, but because it is so cold, it remains dense enough to serve as a main source of the Antarctic Bottom Water that fills the deepest parts of the ocean. During the ice age, cooling or "shutdown" of North Atlantic Deep Water occurred, which in turn led to colder, saltier Antarctic Bottom Water and a colder ocean as a whole (Headly & Severinghaus 2007); however, cooling of Circumpolar Deep Water would have reduced melting beneath Antarctic ice shelves. In turn, some combination of warming North Atlantic Deep Water and shifting winds may bear much of the responsibility for postglacial East and West Antarctic Ice Sheet retreat by returning warm waters to ice-shelf grounding lines. The recent finding of an Antarctic iceberg-rafted debris event at or just after the abrupt northern warming at the start of Meltwater Pulse 1A (Weber et al. 2014) may provide a particularly good target for modeling studies addressing this hypothesis (because the iceberg-rafted debris event occurred at a time of rapid sea-level rise as well as North Atlantic Deep Water warming, correlation certainly does not prove causation).

Ice shelves also experience melting near their calving fronts from tidal or other mixing beneath them (e.g., Horgan et al. 2011). Even seasonally warmed waters can drive this melting. Because ice shelves often are pinned near their fronts by seafloor highs or islands, increasing such melting could contribute to faster ice flow and sea-level rise. However, the greater ability of deep, warm waters to melt ice at grounding lines continues to focus most attention there (Rignot & Jacobs 2002, Walker et al. 2008).

Much work now is addressing the details of the circulations bringing Circumpolar Deep Water or other warm waters to grounding lines, and the plumes or other circulations carrying waters away. Wind-driven and plume-driven circulations may interact in complex ways (e.g., Straneo et al. 2011). Melting may be localized in channels along the undersides of ice shelves (e.g., Rignot & Steffen 2008, Mankoff et al. 2012, Millgate et al. 2013, Stanton et al. 2013), potentially weakening ice shelves but also potentially limiting or localizing melting in ways that could leave more contact with pinning points or fjord walls. Many additional advances, and some surprises, seem likely, but the basics—warmer water thins ice shelves and accelerates sea-level rise—are highly solid.

One might at first suspect that ice-shelf buttressing would decrease smoothly with increasing warming; instead, observations and physical understanding both lead to the expectation that under sufficiently strong forcing a remnant ice shelf will calve away completely, producing a discontinuous jump to a grounded calving front with no buttressing from floating ice. Iceberg calving increases with along-flow spreading (e.g., Alley et al. 2008b, Levermann et al. 2012), and spreading is fastest at the grounding line if buttressing is sufficiently reduced, because ice is thickest there. Hence, the calving line can shift to the grounding line as ice-shelf thinning is followed by ice-shelf loss, as observed at Jakobshavn Isbræ (e.g., Joughin et al. 2012b) and elsewhere (Nettles & Ekstrom 2010; also see Meier & Post 1987). Ice-shelf loss can be further aided by meltwater wedging in crevasses (e.g., Scambos et al. 2004, Cook et al. 2012, Nick et al. 2013).

After removal of an ice shelf, frontal melting of the resulting grounded cliff may be significant, and may enhance loss of not-yet-floating ice. Melting is especially rapid where meltwater emerging from beneath the grounded ice creates buoyant plumes that entrain warm fjord waters; under those circumstances melting rates comparable to those beneath ice shelves are possible (e.g., Rignot et al. 2010; also see Le Brocq et al. 2013 for similar behavior beneath ice shelves). Rapid melting below the waterline can undercut the subaerial part, allowing calving. Fracturing patterns can extend this inland somewhat, such that the calving rate enabled by rapid submarine melting is potentially ten times the melt rate (Bartholomaus et al. 2013, O'Leary & Christoffersen 2013). Furthermore, notch formation at the waterline in response to seasonally warmed near-surface waters or other

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

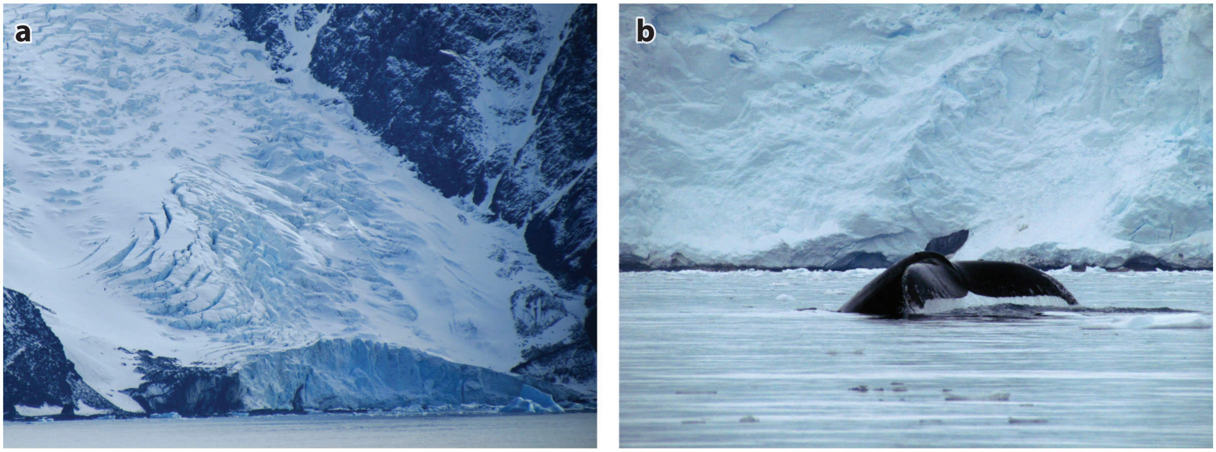

Figure 3

Grounded calving fronts of glaciers. Such features are caused by melting of ice that extends into the water, triggering calving. (*a*) Elephant Island, sub-Antarctic. (*b*) Neko Harbor, Antarctic Peninsula (a humpback whale tail is shown for scale). Photographs by R.B. Alley.

processes can allow subaerial ice to calve downward and submarine ice to calve upward by partially decoupling them from each other.

We note in passing that it is common in some regions to see calving fronts of glaciers that end at the coast but rest on bedrock at or just above sea level (e.g., **Figure 3**). Such common features are fully consistent with ocean melting undercutting and causing calving of ice that otherwise could advance. Under some circumstances, a glacier with a submarine grounding line may be forced to jump rapidly to such a coastal position in response to ocean warming. Moraines are often observed near and roughly parallel to modern fjord coasts in many deglaciated regions (e.g., Bentley et al. 2007 for South Georgia), and it is an interesting possibility that such moraines began forming very soon after a sudden oceanic warming.

In the absence of sufficiently large thermal undercutting or other mechanisms of destabilization (see below), grounded calving faces generally must thin close to flotation to allow calving (Joughin et al. 2012b). And in some cases, a mélange of sea ice and icebergs, or simply sea ice, may provide enough backstress to slow ice discharge somewhat (Joughin et al. 2008). Such a mélange can be weakened or removed by warming.

THWAITES GLACIER COULD EMPTY THE WEAK UNDERBELLY OF THE WEST ANTARCTIC ICE SHEET

Iceberg calving or melting in the ocean balances almost all of the ice accumulation on Antarctica, and much of that on Greenland, and all of this is subject to oceanic forcing. So far, changing winds, which may have resulted from natural variability, global warming, or the Antarctic ozone hole, appear to have played a major role in forcing warmer waters to grounding lines (see the review in Joughin et al. 2012a). Additional greenhouse-gas forcing appears likely to warm waters as well as to change circulation affecting all of the ice sheets (e.g., Hellmer et al. 2012, Straneo et al. 2012, Straneo & Heimbach 2013, Khan et al. 2014).

As discussed by Alley et al. (2008a, supporting material), snowfall on ice sheets is more or less 7 mm/yr of sea-level equivalent, mostly returned by ice flow, and the ice sheets were probably fairly

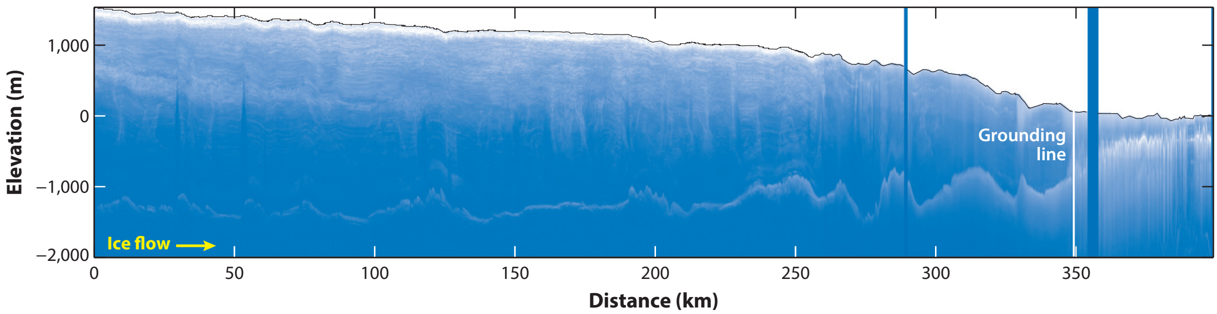

Figure 4

Radar profile (marked with an orange line in **Figure 2**) along Thwaites Glacier. Ice-flow direction is roughly from left to right. The grounding-line position (2011) is plotted in white (Rignot et al. 2014). Radar data from CReSIS (Gogineni 2012).

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

close to being in steady state before the recent human-related forcings. If all ice-sheet outlets were to experience a sustained doubling of ice loss, that would raise sea level by ~0.7 m in 100 yr. As discussed above, velocity increases of more than 2-fold have been observed, but most outlets so far have not experienced such acceleration, especially around the East Antarctic Ice Sheet. Most of the discharge from the ice sheets occurs in bedrock-bounded valleys with notable friction beneath or on the sides. Pfeffer et al. (2008) used physical-plausibility arguments about likely flow accelerations to argue that 2.0 m of sea-level rise by the year 2100 represents an upper limit from flow acceleration, with 0.8 m a more likely upper limit. Any such rise would be added to the contributions from melting of mountain glaciers and the upper surface of ice sheets (primarily the Greenland Ice Sheet) and from thermal expansion of ocean water, and the costs would be quite high.

However, we next focus on Thwaites Glacier in West Antarctica, because of the possibility that it could exceed the speed limits described by Pfeffer et al. (2008) and cause even faster sea-level rise (NRC 2013). Thwaites Glacier drains the so-called weak underbelly of the West Antarctic Ice Sheet, the region identified as early as 1981 by George Denton and Terry Hughes as the most likely route for ice-sheet collapse (see Hughes 1981). Thwaites Glacier now funnels both vertically and horizontally to discharge across a stabilizing sill between highlands that arose at least in part from former volcanism (**Figures 2** and **4**). Were Thwaites Glacier to retreat from this site of relative stability, it would develop a deeper, wider calving front than any observed today (NRC 2013), which might allow faster flow or breakage than considered by Pfeffer et al. (2008), as discussed below. And, because Thwaites Glacier is connected through the deep basins to enough marine ice to raise sea level by ~3.3 m, its loss likely would lead to loss of the marine portions of the West Antarctic Ice Sheet (**Figure 1**).

THWAITES GLACIER IS ACCELERATING AND THINNING RAPIDLY NOW, LIKELY BECAUSE OF OCEANIC FORCING

Thwaites Glacier, along with neighboring ice streams and outlet glaciers including Pine Island Glacier, has recently accelerated and thinned (e.g., McMillan et al. 2014, Mouginot et al. 2014, Sutterley et al. 2014) (**Figure 2**), making the Amundsen Sea/Pine Island Bay drainage the largest Antarctic source of sea-level rise (Velicogna & Wahr 2006, Shepherd & Wingham 2007, Rignot et al. 2008). The changes likely were triggered by reduced ice-shelf buttressing caused by increased sub-ice-shelf melting (e.g., Shepherd et al. 2004, Payne et al. 2007, Rignot et al. 2008, Pritchard et al. 2009, Joughin et al. 2010, Thomas et al. 2011, Park et al. 2013, Favier et al. 2014). The oceanic forcing may have involved warming of the Circumpolar Deep Water that can access the grounding lines (Rignot & Jacobs 2002; Jacobs et al. 2002, 2011; Schmidtko et al. 2014), and enhanced transport of this Circumpolar Deep Water onto the continental shelf and beneath ice shelves (Thoma et al. 2008, Steig et al. 2012, Dutrieux et al. 2014, Schmidtko et al. 2014). Additional warming of Circumpolar Deep Water and shifts in wind patterns affecting ocean circulation in the coming century are typically simulated under business-as-usual emissions of fossil-fuel CO_2 (Sen Gupta et al. 2009, Hattermann & Levermann 2010).

FUTURE THWAITES GLACIER CHANGES WILL LIKELY EXHIBIT THRESHOLD BEHAVIOR

The sedimentary record of former ice sheets on continental shelves shows that sustained stability typically alternates with uninterrupted, perhaps rapid retreat. The sedimentary record can reveal even short-lived pauses during retreat; repeated transverse moraines are occasionally observed that likely record wintertime pauses caused by backstress arising from sea ice or mélange (e.g., Dowdeswell et al. 2008). However, such features are not typical. Instead, the grounding line sits for an extended time (decades to millennia) at one site, depositing a large grounding-line wedge, and then migrates inland to a new site without pausing long enough to leave any record (e.g., Wellner et al. 2006, Dowdeswell et al. 2008, Jakobsson et al. 2011, Dowdeswell & Fugelli 2012). The rate of retreat is not tightly constrained, but there are at least suggestions of rapid changes (Jakobsson et al. 2011).

For the modern ice sheets, the retreats from ice-age maxima obviously did not "run away" to cause complete deglaciation. Instead, the grounding lines restabilized at local seafloor highs (Alley et al. 2007) or fjord narrowings (Jamieson et al. 2012). Once a retreating grounding line stabilizes, several processes serve to increase stability. These include sedimentation replacing water with sediment that provides more friction (Anandakrishnan et al. 2007), isostatic recovery as the seafloor rises in response to loss of the weight of the former ice sheet, and relaxation of self-gravitation as the local sea level falls because the shrinking ice mass exerts less pull on the surrounding ocean water (Gomez et al. 2010). These stabilizers are relatively slow, however; isostatic response and sedimentation have timescales of centuries to millennia, and self-gravitation depends on the ice volume inland and thus the time for response there to coastal thinning. Thus, faster forcing is expected to be more effective at destabilizing grounding lines, a potentially important issue given that a recent study of Arctic paleoclimates indicated that by some measures human forcing may become faster than any natural forcings over at least the most recent tens of millions of years (White et al. 2010).

New cosmogenic-isotope results on ice-sheet thinning in the Hudson Mountains, adjacent to Pine Island Glacier and not far from Thwaites Glacier, confirm the marine record of punctuated retreat. Johnson et al. (2014) found that the ice thinned >100 m at a rate of >1 m/yr about 8,000 yr ago, to a level near modern, and then apparently changed little before the onset of recent thinning. The prior thinning event occurred at a time of little sea-level change, and likely resulted from ice-shelf melting from warmer ocean waters (Johnson et al. 2014).

WHETHER THE THWAITES GLACIER STABILITY THRESHOLD HAS BEEN CROSSED, AND WHEN THE RAPID RETREAT WILL OCCUR, MAY DEPEND ON POORLY KNOWN CONDITIONS

Few modeling studies have specifically addressed the near-future stability of Thwaites Glacier using the highest-resolution data available. Many more modeling studies have focused on the

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

neighboring Pine Island Glacier, and these provide important lessons (e.g., Schmeltz et al. 2002, Payne et al. 2004, Joughin et al. 2010, Larour et al. 2012, Park et al. 2013, Favier et al. 2014). However, these studies do not substitute for work on Thwaites Glacier in projecting near-future ice-sheet changes, because Pine Island Glacier has less direct access to the deepest marine basins beneath the West Antarctic Ice Sheet, and because there is greater bedrock control of Pine Island Glacier, with shearing from side drag extending to the centerline (MacGregor et al. 2012). Pine Island and Thwaites Glaciers also likely differ oceanographically in some ways (see, e.g., references in Parizek et al. 2013).

The recent paper by Joughin et al. (2014a) found that Thwaites Glacier ice-shelf loss has likely already created conditions that will lead to retreat from the stabilizing sill, and thus to deglaciation of the marine portions of the West Antarctic Ice Sheet. The model runs indicate slow changes for some interval into the future (typically centuries) before a sudden acceleration as the grounding line shifts into the deeper basins, creating sea-level rise of >1 mm/yr from Thwaites Glacier alone. [Model runs generally were truncated when sea-level rise exceeded 1 mm/yr because the model used is not appropriate for assessing how much faster than 1 mm/yr Thwaites Glacier can contribute, although "much greater losses generally follow within a few years" (Joughin et al. 2014a, p. 737).] The runs reached this 1-mm/yr sea-level-rise threshold in 200 to 900 yr but do not span the full range of possibilities; both longer and shorter delays could be simulated. Additional uncertainty arises from a lack of data for Thwaites Glacier; Joughin et al. (2014a) used an empirically calibrated sub-ice-shelf melt function from Pine Island Glacier, for example.

Extensive simulations were conducted by Parizek et al. (2013) (**Figure 5**), using data sets from the SeaRISE project and building on the SeaRISE modeling (Bindschadler et al. 2013, Nowicki et al. 2013a). Detailed aerogeophysical and satellite data were used as inputs to a coupled model of the Thwaites Glacier ice stream, its ice shelf, and a sub-ice-shelf ocean plume, with the option of allowing ocean influences beneath the seaward-most grounded ice, reflecting recent discoveries of estuary-type water systems and tidal pumping near grounding lines (Christianson et al. 2013, Horgan et al. 2013, Walker et al. 2013). The SeaRISE data sets were collected at different times while Thwaites Glacier changed, and combining them probably had the effect of making the initial model glacier somewhat more stable than the actual glacier.

Parizek et al.'s (2013) simulations found that the main Thwaites Glacier ice shelf now provides limited stability, although the Eastern Shelf still buttresses that side of Thwaites Glacier (Rignot 2006, MacGregor et al. 2012; also see Joughin et al. 2014a), where the region of fastest flow could widen by outward stepping of the margin (MacGregor et al. 2013). Hence, warming beneath existing ice shelves could cause further flow acceleration. Warming is also expected to increase snow accumulation, but the effect was modeled to increase stability only slightly.

Both the SeaRISE simulations (Bindschadler et al. 2013, Nowicki et al. 2013a) and Parizek et al. (2013) found that under plausible forcings and parameter combinations, both long-term Thwaites Glacier stability and rapid retreat can occur (**Figure 5**). Results were especially sensitive to poorly known details of interactions in the grounding zone, and to the viscous versus plastic nature of the bed. Coastal thinning from reduced ice-shelf buttressing causes a wave of thinning to propagate inland, steepening the ice sheet and speeding ice flow to the grounding zone. The perturbation propagates rapidly on a nearly plastic bed, supplying much ice to the grounding zone in response to very small thinning and thus limiting further thinning, whereas coastal changes are larger with a viscous bed. Because retreat from the stabilizing sill requires thinning there, results are sensitive to the assumed bed character. Joughin et al. (2014a) used a nearly viscous bed, which they considered to be applicable to Thwaites Glacier, explaining some of the differences between their simulations and those of Parizek et al. (2013), but the full character of the bed extending inland is not known with high confidence (see, e.g., Tulaczyk et al. 2000, Rathbun et al. 2008,

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

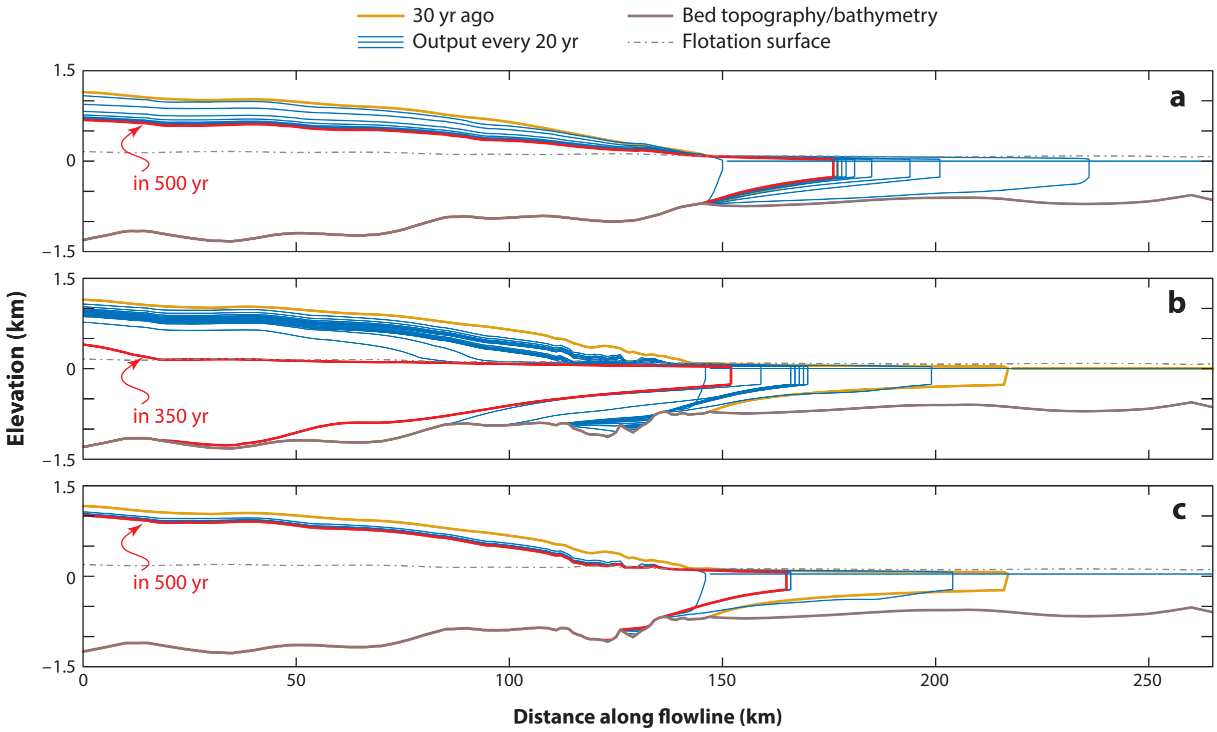

Figure 5

Thwaites Glacier flowline simulations. Flowline model runs for (*a*) SeaRISE, under a climate ensemble mean of 18 IPCC AR4 (Intergovernmental Panel on Climate Change Fourth Assessment Report) climate models for emissions scenario A1B, a 2× basal sliding amplification, and sub-ice-shelf melting from coupling with an ocean-plume model for 500 yr; (*b*) a 7-km-wide grounding zone, high-resolution bed topography, and linear-viscous substrate using the same climate and basal-melt forcing as above and no basal sliding amplification; and (*c*) a 7-km-wide grounding zone, high-resolution bed topography, and effectively plastic substrate for the same climate and basal-melt forcing as above and no basal sliding amplification. (Note the retreat across the model domain in 200 yr in panel *b* and the lack of retreat in panel *c*, for the linear-viscous and effectively plastic basal rheologies, respectively.) The brown line is the bedrock topography from (*a*) SeaRISE (Bamber et al. 2009a; Griggs & Bamber 2009; Le Brocq et al. 2010; Bindschadler et al. 2013; Nowicki et al. 2013a,b) and (*b*,*c*) a combination of SeaRISE and Operation Ice Bridge (Allen 2009, Leuschen 2010). Figure modified with permission from figures 6*f* and 10*d*,*e* in Parizek et al. (2013). Copyright © 2013 by Wiley.

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

Joughin et al. 2009, Sergienko & Hindmarsh 2013, Walker et al. 2014). And, although Joughin et al. (2014a) explored a wide range of parameter space representing the most likely values, they did not explore all possible parameter combinations, and some will likely yield either greater or less instability; parameterizations for basal melt of ice shelves may be especially important.

ANY FUTURE THWAITES GLACIER RETREAT MAY BE MUCH FASTER THAN GENERALLY MODELED, BECAUSE OF CLIFF INSTABILITY

Ice-shelf loss leaves a tidewater calving face (e.g., Meier & Post 1987). However, stress imbalances cause a tendency for cliff failure, as is well known in quarrying, mining, geomorphology, rock climbing, and elsewhere. Hollin (1962) noted that the margin of the ice sheet often forms a cliff that rarely exceeds 30 m, and that height "is presumably limited by mechanical factors" (p. 179). This topic was taken up in more detail by Hanson & Hooke (2003) and Bassis & Walker (2012), who suggested that the maximum height for a subaerial ice cliff is near 100 m. This is close to the greatest subaerial ice-cliff heights observed today [Jakobshavn Isbræ and Helheim Glacier in Greenland (Nick et al. 2013), and Crane Glacier after the loss of the Larsen B Ice Shelf in Antarctica (Scambos et al. 2011)] and corresponds to flotation for a bed almost 1,000 m below sea level. [The subaerial ice cliff at Jakobshavn Isbræ has sometimes slightly exceeded 100 m (e.g., Joughin et al. 2012b, using the bed elevation assuming near-flotation at calving), perhaps aided by mélange strength and frictional support from the walls of the narrow fjord.]

If Thwaites Glacier experiences a sustained Jakobshavn-type ice-shelf loss and retreats into the central basins beneath the West Antarctic Ice Sheet that reach >2,000 m below sea level, the resulting cliff would almost surely be highly unstable. If any mélange produced were sufficiently weak, the resulting cliff failure might cause the glacier to retreat much more rapidly than simulated by models lacking this process. Pollard et al. (2015) found that a parameterization for this process increased the instability and collapse rate of the West Antarctic Ice Sheet and of marine portions of the East Antarctic Ice Sheet, with this parameterization and the forcing adopted causing West Antarctic Ice Sheet collapse to occur on multidecadal timescales once initiated. No fully physical model now exists that includes this process, and given the dependence on poorly known fracture mechanics of ice, a quantitatively well-constrained model appears unlikely in the near future.

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

THWAITES GLACIER IS OF PARTICULAR CONCERN, BUT PORTIONS OF ALL THE ICE SHEETS MAY BE MORE VULNERABLE THAN PREVIOUSLY THOUGHT

The Pliocene history of iceberg-rafted debris around East Antarctica and the far-field sea-level record, noted above, suggest that the East Antarctic Ice Sheet lost significant mass, likely from the Wilkes Basin. Recent modeling by Mengel & Levermann (2014) and Pollard et al. (2015) exhibits the sort of threshold behavior described above, with sufficient warming causing retreat into deeper basins that speed retreat. [Similar behavior is modeled for Pine Island Glacier in the West Antarctic Ice Sheet by Favier et al. (2014).]

For the Greenland Ice Sheet, Morlighem et al. (2014) used a mass-conservation calculation and the spectacular new radar data from CReSIS and IceBridge to show that deep channels extend farther inland beneath the ice sheet in many more places than shown in older maps. Although still relatively narrow, and not nearly so deep as the West Antarctic Ice Sheet basins, these paths for iceberg discharge make the ice sheet more susceptible to mass loss than when modeled with the older data.

SUMMARY

Marine ice sheets—those with beds well below sea level—can change more rapidly than land-ending ice sheets by discharging icebergs that melt elsewhere, and marine ice sheets exhibit instability on beds that deepen toward the center of the ice sheet. Stability can be reestablished if floating ice remains attached in ice shelves that provide sufficient buttressing arising from ice-shelf friction with fjord walls or local seafloor highs. However, the upper surfaces of ice shelves are relatively low and warm compared with ice-sheet interiors, and melting on the ice shelves can fill lakes that drive crevasses through the shelves, triggering collapse. Lower surfaces and fronts of ice shelves in water are at the melting point, so oceanic changes that deliver warmer water to contact the ice shelves will increase melting, with strong sensitivity. Hence, marine ice sheets such as the West Antarctic Ice Sheet can be greatly and rapidly influenced by warming.

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

The available data indicate that the West Antarctic Ice Sheet deglaciated in the geologically recent past. Far-field data on rates of sea-level rise from ice-sheet configurations similar to today's lack the time resolution needed to assess the rate of deglaciation, but they allow for rates much higher than the recently observed sea-level rise. The climate above the West Antarctic Ice Sheet remained relatively cold during these intervals, so dynamic processes linked to the marine instability and likely forced by warmer ocean waters are implicated, acting through a range of complex but increasingly well understood processes.

Thwaites Glacier, in the weak underbelly of the West Antarctic Ice Sheet, is the most likely conduit for past and future ice-sheet collapse. Unlike retreat of other ice streams or outlet glaciers such as Pine Island Glacier, retreat of Thwaites Glacier from its current position would quickly and directly access the deep interior basins across a very broad iceberg calving front. Thwaites Glacier is now stabilized on a relatively high bedrock sill where topography narrows the flow, but the glacier is accelerating and thinning rapidly, likely because of oceanic forcing.

The geological record, and modeling targeting the modern setting, indicate that Thwaites Glacier will likely exhibit threshold behavior, with retreat off the sill (**Figures 2** and **4**) triggering much more rapid retreat that likely will be irreversible over human timescales of centuries or less. The stability threshold may already have been crossed, although the few modeling experiments to date do not provide full agreement on that. Whether the threshold has been crossed, and how rapidly the ongoing retreat may leave the stabilizing sill, may depend on processes and conditions that are not yet fully understood and measured, such that future modeling seems likely to leave substantial uncertainty for some time. Notably, retreat into the deep basins could create conditions unlike any seen on Earth today, with processes that generally are not accurately represented in the current generation of models. In particular, the possibility of cliff failure suggests that Thwaites Glacier retreat and West Antarctic Ice Sheet loss could be much faster than generally simulated.

And, although Thwaites Glacier and the West Antarctic Ice Sheet are of greatest concern, recent research shows that the Greenland Ice Sheet has greater marine character than previously thought, and portions of the East Antarctic Ice Sheet are vulnerable to marine instability from sufficiently large forcing. Vigorous research will be required to notably reduce the uncertainties, and the threshold nature of some of the behavior suggests that notable uncertainty will remain for some time even with vigorous research. The distribution of possible future rates and amounts of sea-level rise thus appears to be highly skewed, with a long tail to large, rapid, and costly rise, a phenomenon that is of potential interest to planners.

DISCLOSURE STATEMENT

The authors are not aware of any affiliations, memberships, funding, or financial holdings that might be perceived as affecting the objectivity of this review.

ACKNOWLEDGMENTS

Partial funding was provided by the US National Science Foundation, under PLR ANT grants 0424589 (the CReSIS Science and Technology Center), 1043018, 1043528, and AGS 1338832 (the FESD VOICE Project), and by NASA, under grants NNX10AI04G, NNX12AB69G, NNX12AD03A, and NNX12AP50G. We thank the members of Penn State Ice and Climate Exploration (PSICE), and numerous other colleagues who conducted the studies reviewed here or made them possible. We thank Eric Rignot for supplying grounding-line maps, and an anonymous referee for helpful suggestions.

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

LITERATURE CITED

Adkins JF. 2013. The role of deep ocean circulation in setting glacial climates. *Paleoceanography* 28:539–61

Allen C. 2009. *IceBridge MCoRDS L2 Ice Thickness*. Boulder, CO: Natl. Snow Ice Data Cent.

Alley RB. 2007. Wally was right: predictive ability of the North Atlantic "conveyor belt" hypothesis for abrupt climate change. *Annu. Rev. Earth Planet. Sci.* 35:241–72

Alley RB, Anandakrishnan S, Dupont TK, Parizek BR, Pollard D. 2007. Effect of sedimentation on ice-sheet grounding-line stability. *Science* 315:1838–41

Alley RB, Andrews JT, Brigham-Grette J, Clarke GKC, Cuffey KM, et al. 2010. History of the Greenland Ice Sheet: paleoclimatic insights. *Quat. Sci. Rev.* 29:1728–56

Alley RB, Dupont TK, Parizek BR, Anandakrishnan S, Lawson DE, et al. 2006. Outburst flooding and the initiation of ice-stream surges in response to climatic cooling: a hypothesis. *Geomorphology* 75:76–89

Alley RB, Fahnestock M, Joughin I. 2008a. Climate change: understanding glacier flow in changing times. *Science* 322:1061–62

Alley RB, Horgan HJ, Joughin I, Cuffey KM, Dupont TK, et al. 2008b. A simple law for ice-shelf calving. *Science* 322:1344

Alley RB, Whillans IM. 1984. Response of the East Antarctic ice sheet to sea-level rise. *J. Geophys. Res.* 89(C4):6487–93

Alley RB, White JWC. 2000. *Evidence of West Antarctic changes in the Siple Dome ice core*. Presented at 7th Annu. West Antarct. Ice Sheet Worksh., Sept. 28–30, Sterling, VA. **http://www.waisworkshop.org/pastmeetings/abstracts00/Alley.htm**

Anandakrishnan S, Catania GA, Alley RB, Horgan HJ. 2007. Discovery of till deposition at the grounding line of Whillans Ice Stream. *Science* 31:51835–38

Bamber JL, Gomez-Dans JL, Griggs JA. 2009a. A new 1 km digital elevation model of the Antarctic derived from combined satellite radar and laser data. Part 1: Data and methods. *Cryosphere* 3:101–11

Bamber JL, Riva REM, Vermeersen BLA, Le Brocq AM. 2009b. Reassessment of the potential sea-level rise from a collapse of the West Antarctic Ice Sheet. *Science* 324:901–3

Barnes D, Hillenbrand C. 2010. Faunal evidence for a late quaternary trans-Antarctic seaway. *Glob. Change Biol.* 16:3297–303

Bartholomaus TC, Larsen CF, O'Neel S. 2013. Does calving matter? Evidence for significant submarine melt. *Earth Planet. Sci. Lett.* 380:21–30

Bassis JN, Walker CC. 2012. Upper and lower limits on the stability of calving glaciers from the yield strength envelope of ice. *Proc. R. Soc. A* 468:913–31

Bentley CR, Ostenso NA. 1961. Glacial and subglacial topography of West Antarctica. *J. Glaciol.* 3:882–912

Bentley MJ, Evans DJA, Fogwill CJ, Hansom JD, Sugden DE, Kubik PW. 2007. Glacial geomorphology and chronology of deglaciation, South Georgia, sub-Antarctic. *Quat. Sci. Rev.* 26:644–77

Bindschadler RA, Nowicki S, Abe-Ouchi A, Aschwanden A, Choi H, et al. 2013. Ice-sheet model sensitivities to environmental forcing and their use in projecting future sea level (the SeaRISE project). *J. Glaciol.* 59:195–224

Bradley SL, Siddall M, Milne GA, Masson-Delmotte V, Wolff E. 2012. Where might we find evidence of a Last Interglacial West Antarctic ice sheet collapse in Antarctic ice core records? *Glob. Planet. Change* 88–89:64–75

Carlson AE, Winsor K. 2012. Northern Hemisphere ice-sheet responses to past climate warming. *Nat. Geosci.* 5:607–13

Cazenave A, Le Cozannet G. 2014. Sea level rise and its coastal impacts. *Earth's Future* 2:15–34

Christianson K, Parizek BR, Alley RB, Horgan HJ, Jacobel RW, et al. 2013. Ice sheet grounding zone stabilization due to till compaction. *Geophys. Res. Lett.* 40:5406–11

Cook CP, van de Flierdt T, Williams T, Hemming SR, Iwai M, et al. 2013. Dynamic behaviour of the East Antarctic ice sheet during Pliocene warmth. *Nat. Geosci.* 6:765–69

Cook S, Rutt IC, Murray T, Luckman A, Zwinger N, et al. 2014. Modelling environmental influences on calving at Helheim Glacier in eastern Greenland. *Cryosphere* 8:827–41

Cook S, Zwinger T, Rutt IC, O'Neel S, Murray T. 2012. Testing the effect of water in crevasses on a physically based calving model. *Ann. Glaciol.* 53:90–96

Cuffey KM, Paterson WSB. 2010. *The Physics of Glaciers*. Oxford, UK: Butterworth-Heinemann. 4th ed.

Dahl-Jensen D, Albert MR, Aldahan A, Azuma N, Balsley-Clausen D, et al. 2013. Eemian interglacial reconstructed from a Greenland folded ice core. *Nature* 493:489–94

Das SB, Joughin I, Behn MD, Howat IM, King MA, et al. 2008. Fracture propagation to the base of the Greenland Ice Sheet during supraglacial lake drainage. *Science* 320:778–81

Denton GH, Alley RB, Comer GC, Broecker WS. 2005. The role of seasonality in abrupt climate change. *Quat. Sci. Rev.* 24:1159–82

Depoorter MA, Bamber JL, Griggs JA, Lenaerts JTM, Lightenberg SRM, et al. 2013. Calving fluxes and basal melt rates of Antarctic ice shelves. *Nature* 502:89–92

Dowdeswell JA, Fugelli EMG. 2012. The seismic architecture and geometry of grounding-zone wedges formed at the marine margins of past ice sheets. *Geol. Soc. Am. Bull.* 124:1750–61

Dowdeswell JA, Ottesen D, Evans J, Cofaigh CÓ, Anderson JB. 2008. Submarine glacial landforms and rates of ice-stream collapse. *Geology* 36:819–22

Dupont TK, Alley RB. 2005. Assessment of the importance of ice-shelf buttressing to ice-sheet flow. *Geophys. Res. Lett.* 32:L04503

Dutrieux P, De Rydt J, Jenkins A, Holland PR, Ha HK, et al. 2014. Strong sensitivity of Pine Island ice-shelf melting to climatic variability. *Science* 343:174–78

Dutton A, Lambeck K. 2012. Ice volume and sea level during the last interglacial. *Science* 337:216–19

Ewing M, Donn WL. 1956. A theory of ice ages. *Science* 123:1061–65

Favier L, Durand G, Cornford SL, Gudmundsson GH, Gagliardini O, et al. 2014. Retreat of Pine Island Glacier controlled by marine ice-sheet instability. *Nat. Clim. Change* 4:117–21

Fretwell P, Pritchard HD, Vaughan DG, Bamber JL, Barrand NE, et al. 2013. Bedmap2: improved ice bed, surface and thickness datasets for Antarctica. *Cryosphere* 7:375–93

Frigola J, Canals M, Cacho I, Moreno A, Sierro FJ, et al. 2012. A 500 kyr record of global sea-level oscillations in the Gulf of Lion, Mediterranean Sea: new insights into MIS 3 sea-level variability. *Clim. Past* 8:1067–77

Gogineni P. 2012. *CReSIS radar depth sounder data*. Center for Remote Sensing of Ice Sheets (CReSIS), Lawrence, KS, retrieved July 2014. **http://data.cresis.ku.edu/**

Gomez N, Mitrovica JX, Huybers P, Clark PU. 2010. Sea level as a stabilizing factor for marine-ice-sheet grounding lines. *Nat. Geosci.* 3:850–53

Griggs JA, Bamber JL. 2009. A new 1 km digital elevation model of the Antarctic derived from combined satellite radar and laser data. Part 2: Validation and error estimates. *Cryosphere* 3:113–23

Hanson B, Hooke RLeB. 2003. Buckling rate and overhang development at a calving face. *J. Glaciol.* 49:577–86

Haran T, Bohlander J, Scambos T, Painter T, Fahnestock M. 2005. *MODIS Mosaic of Antarctica 2003–2004* (*MOA2004*) *Image Map*. Boulder, CO: Natl. Snow Ice Data Cent. Updated 2013. doi: 10.7265/N5ZK5DM5

Hattermann T, Levermann A. 2010. Response of Southern Ocean circulation to global warming may enhance basal ice shelf melting around Antarctica. *Clim. Dyn.* 35:741–56

Headly MA, Severinghaus JP. 2007. A method to measure Kr/N_2 ratios in air bubbles trapped in ice cores and its application in reconstructing past mean ocean temperature. *J. Geophys. Res.* 112:D19105

Hellmer HH, Kauker F, Timmermann R, Determann J, Rae J. 2012. Twenty-first-century warming of a large Antarctic ice-shelf cavity by a redirected coastal current. *Nature* 485:225–28

Hemming SR. 2004. Heinrich events: massive late Pleistocene detritus layers of the North Atlantic and their global climate imprint. *Rev. Geophys.* 42:RG1005

Holden PB, Edwards NR, Wolff EW, Valdes PJ, Singarayer JS. 2011. The Mid-Brunhes Event and West Antarctic Ice Sheet stability. *J. Quat. Sci.* 26:474–77

Holland DM, Thomas RH, de Young B, Ribergaard MH, Lyberth B. 2008a. Acceleration of Jakobshavn Isbræ triggered by warm subsurface ocean waters. *Nat. Geosci.* 1:659–64

Holland PR, Jenkins A, Holland DM. 2008b. The response of ice shelf basal melting to variations in ocean temperature. *J. Clim.* 21:2558–72

Hollin JT. 1962. On the glacial history of Antarctica. *J. Glaciol.* 4:173–95

Holschuh N, Pollard D, Alley RB, Anandakrishnan S. 2014. Evaluating Marie Byrd Land stability using an improved basal topography. *Earth Planet. Sci. Lett.* 408:362–69

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

Horgan HJ, Alley RB, Christianson K, Jacobel RW, Anandakrishnan S, et al. 2013. Estuaries beneath ice sheets. *Geology* 41:1159–62

Horgan HJ, Walker RT, Anandakrishnan S, Alley RB. 2011. Surface elevation changes at the front of the Ross Ice Shelf: implications for basal melting. *J. Geophys. Res.* 116:C02005

Hughes TJ. 1981. The "weak underbelly" of the West Antarctic Ice Sheet. *J. Glaciol.* 27:518–25

Jacobs SS, Giulivi CF, Mele PA. 2002. Freshening of the Ross Sea during the late 20th century. *Science* 297:386–89

Jacobs SS, Jenkins A, Giulivi CF, Dutrieux P. 2011. Stronger ocean circulation and increased melting under Pine Island Glacier ice shelf. *Nat. Geosci.* 4:519–23

Jakobsson M, Anderson JB, Nitsche FO, Dowdeswell JA, Ryllencreutz R, et al. 2011. Geological record of ice shelf breakup and grounding line retreat, Pine Island Bay, West Antarctica. *Geology* 39:691–94

Jamieson SSR, Vieli A, Livingstone SJ, Cofaigh CÓ, Stokes C, et al. 2012. Ice-stream stability on a reverse bed slope. *Nat. Geosci.* 5:1–4

Jenkins A, Doake CSM. 1991. Ice-ocean interaction on Ronne Ice Shelf, Antarctica. *J. Geophys. Res.* 96(C1):791–813

Jenkins A, Holland DM. 1999. Modeling thermodynamic ice-ocean interactions at the base of an ice shelf. *J. Phys. Oceanogr.* 29:1787–800

Johnson JS, Bentley MJ, Smith JA, Finkel RC, Rood DH, et al. 2014. Rapid thinning of Pine Island Glacier in the early Holocene. *Science* 343:999–1001

Joughin I, Alley RB. 2011. Stability of the West Antarctic Ice Sheet in a warming world. *Nat. Geosci.* 4:506–513

Joughin I, Alley RB, Holland DM. 2012a. Ice-sheet response to oceanic forcing. *Science* 338:1172–76

Joughin I, Howat I, Alley RB, Ekstrom G, Fahnestock M, et al. 2008. Ice-front variation and tidewater behavior on Helheim and Kangerdlugssuaq Glaciers, Greenland. *J. Geophys. Res.* 113:F01004

Joughin I, Smith BE, Holland DM. 2010. Sensitivity of 21st century sea level to ocean-induced thinning of Pine Island Glacier, Antarctica. *Geophys. Res. Lett.* 37:L20502

Joughin I, Smith BE, Medley B. 2014a. Marine ice sheet collapse potentially under way for the Thwaites Glacier Basin, West Antarctica. *Science* 344:735–38

Joughin I, Smith BE, Shean DE, Floricioiu D. 2014b. Further summer speedup of Jakobshavn Isbræ. *Cryosphere* 8:209–14

Joughin I, Smith I, Howat D, Floricioiu D, Alley RB, et al. 2012b. Seasonal to decadal scale variations in the surface velocity of Jakobshavn Isbrae, Greenland: observation and model-based analysis. *J. Geophys. Res.* 117:F02030

Joughin I, Tulaczyk S, Bamber JL, Blankenship D, Holt JW, et al. 2009. Basal conditions for Pine Island and Thwaites Glaciers, West Antarctica, determined using satellite and airborne data. *J. Glaciol.* 55:245–57

Khan SA, Kjaer KH, Bevis M, Bamber JL, Wahr J, et al. 2014. Sustained mass loss of the northeast Greenland ice sheet triggered by regional warming. *Nat. Clim. Change* 4:292–99

Kopp RE, Simons FJ, Mitrovica JX, Maloof AC, Oppenheimer M. 2009. Probabilistic assessment of sea level during the last interglacial stage. *Nature* 462:863–67

Kopp RE, Simons FJ, Mitrovica JX, Maloof AC, Oppenheimer M. 2013. A probabilistic assessment of sea level variations within the last interglacial stage. *Geophys. J. Int.* 193:711–16

Korotkikh EV, Mayewski PA, Handley MJ, Sneed SB, Introne DS, et al. 2011. The last interglacial as represented in the glaciochemical record from Mount Moulton Blue Ice Area, West Antarctica. *Quat. Sci. Rev.* 30:1940–47

Larour E, Morlighem M, Seroussi H, Schiermeier J, Rignot E. 2012. Ice flow sensitivity to geothermal heat flux of Pine Island Glacier, Antarctica. *J. Geophys. Res.* 117:F04203

Le Brocq AM, Payne AJ, Vieli A. 2010. An improved Antarctic dataset for high resolution numerical ice sheet models. *Earth Syst. Sci. Data* 2:247–60

Le Brocq AM, Ross N, Griggs JA, Bingham RG, Corr HFJ, et al. 2013. Evidence from ice shelves for channelized meltwater flow beneath the Antarctic Ice Sheet. *Nat. Geosci.* 6:945–48

Leeson AA, Shepherd A, Briggs K, Howat I, Fettweis X, et al. 2015. Supraglacial lakes on the Greenland ice sheet advance inland under warming climate. *Nat. Clim. Change* 5:51–55

Levermann A, Albrecht T, Winkelmann R, Martin MA, Haseloff M, Joughin I. 2012. Kinematic first-order calving law implies potential for abrupt ice-shelf retreat. *Cryosphere* 6:273–86

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

MacAyeal DR. 1993. Binge/purge oscillations of the Laurentide Ice Sheet as a cause of the North Atlantic's Heinrich events. *Paleoceanography* 8:775–84

MacAyeal DR, Scambos TA, Hulbe CL, Fahnestock MA. 2003. Catastrophic ice-shelf break-up by an ice-shelf-fragment-capsize mechanism. *J. Glaciol.* 49:22–36

MacGregor JA, Catania GA, Conway HB, Schroeder DM, Joughin IR, et al. 2013. Weak bed control of the eastern shear margin of Thwaites Glacier, West Antarctica. *J. Glaciol.* 59:900–12

MacGregor JA, Catania GA, Marskowski MS, Andrews AG. 2012. Widespread rifting and retreat of ice-shelf margins in the eastern Amundsen Sea Embayment between 1972 and 2011. *J. Glaciol.* 58:458–66

Mankoff KD, Jacobs SS, Tulaczyk SM, Stammerjohn SE. 2012. The role of Pine Island Glacier ice shelf basal channels in deep-water upwelling, polynyas and ocean circulation in Pine Island Bay, Antarctica. *Ann. Glaciol.* 53:123–28

Marcott SA, Clark PU, Padman L, Klinkhammer GP, Springer SR, et al. 2011. Ice-shelf collapse from subsurface warming as a trigger for Heinrich events. *PNAS* 108:13415–19

McMillan M, Shepherd A, Sundal A, Briggs K, Muir A, et al. 2014. Increased ice losses from Antarctica detected by CryoSat-2. *Geophys. Res. Lett.* 41:3899–905

Meier MF, Post AS. 1987. Fast tidewater glaciers. *J. Geophys. Res.* 92(B9):9051–58

Mengel M, Levermann A. 2014. Ice plug prevents irreversible discharge from East Antarctica. *Nat. Clim. Change* 4:451–55

Mercer JH. 1968. Antarctic ice and Sangamon sea level. *Int. Assoc. Sci. Hydrol. Publ.* 79:217–25

Millgate T, Holland PR, Jenkins A, Johnson HL. 2013. The effect of basal channels on oceanic ice-shelf melting. *J. Geophys. Res. Oceans* 118:6951–64

Morlighem M, Rignot E, Mouginot J, Seroussi H, Larour E. 2014. Deeply incised submarine glacial valleys beneath the Greenland Ice Sheet. *Nat. Geosci.* 7:418–22

Motyka RJ, Truffer M, Mortensen J, Rysgaard S, Howat I. 2011. Submarine melting of the 1985 Jakobshavn Isbræ floating tongue and the triggering of the current retreat. *J. Geophys. Res.* 116:F01007

Mouginot J, Rignot E, Scheuchl B. 2014. Sustained increase in ice discharge from the Amundsen Sea Embayment, West Antarctica, from 1873 to 2013. *Geophys. Res. Lett.* 41:L059069

Naish T, Powell R, Levy R, Wilson G, Scherer R, et al. 2009. Obliquity-paced Pliocene West Antarctic Ice Sheet oscillations. *Nature* 458:322–28

Nettles M, Ekstrom G. 2010. Glacial earthquakes in Greenland and Antarctica. *Annu. Rev. Earth Planet. Sci.* 38:467–91

Nick FM, Vieli A, Andersen ML, Joughin I, Payne A, et al. 2013. Future sea-level rise from Greenland's main outlet glaciers in a warming climate. *Nature* 497:235–38

Nitsche FO, Gohl K, Larter RD, Hillenbrand CD, Kuhn G, et al. 2013. Paleo ice flow and subglacial meltwater dynamics in Pine Island Bay, West Antarctica. *Cryosphere* 7:249–62

Nowicki S, Bindschadler RA, Abe-Ouchi A, Aschwander A, Bueler E, et al. 2013a. Insights into spatial sensitivities of ice mass response to environmental change from the SeaRISE ice sheet modeling project I: Antarctica. *J. Geophys. Res. Earth Surf.* 118:1002–24

Nowicki S, Bindschadler RA, Abe-Ouchi A, Aschwander A, Bueler E, et al. 2013b. Insights into spatial sensitivities of ice mass response to environmental change from the SeaRISE ice sheet modeling project II: Greenland. *J. Geophys. Res. Earth Surf.* 118:1025–44

NRC (Natl. Res. Counc.). 2013. *Abrupt Impacts of Climate Change: Anticipating Surprises*. Washington, DC: Natl. Acad. Press

O'Leary M, Christoffersen P. 2013. Calving on tidewater glaciers amplified by submarine frontal melting. *Cryosphere* 7:119–28

O'Leary MJ, Hearty PJ, Thompson WG, Raymo ME, Mitrovica JX, Webster JM. 2013. Ice sheet collapse following a prolonged period of stable sea level during the last interglacial. *Nat. Geosci.* 6:796–800

Parizek BR, Alley RB. 2004. Implications of increased Greenland surface melt under global-warming scenarios: ice-sheet simulations. *Quat. Sci. Rev.* 23:1013–27

Parizek BR, Christianson K, Anandakrishnan S, Alley RB, Walker RT, et al. 2013. Dynamic (in)stability of Thwaites Glacier, West Antarctica. *J. Geophys. Res. Earth Surf.* 118:638–55

Park JW, Gourmelen N, Shepherd A, Kim SW, Vaughan DG, Wingham DJ. 2013. Sustained retreat of the Pine Island Glacier. *Geophys. Res. Lett.* 40:2137–42

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

Payne AJ, Holland PR, Shepherd AP, Rutt IC, Jenkins A, Joughin I. 2007. Numerical modeling of ocean-ice interactions under Pine Island Bay's ice shelf. *J. Geophys. Res.* 112:C10019

Payne AJ, Vieli A, Shepherd A, Wingham DJ, Rignot E. 2004. Recent dramatic thinning of largest West Antarctic ice stream triggered by oceans. *Geophys. Res. Lett.* 31:L23401

Penck A. 1928. The causes of the ice-age. *Sitz. Preuss. Akad. Wiss. Phys. Math. Kl.* 1928:76–85

Pfeffer WT, Harper JT, O'Neel S. 2008. Kinematic constraints on glacier contributions to 21st-century sea-level rise. *Science* 321:1340–43

Pollard D, DeConto RM. 2009. Modelling West Antarctic ice sheet growth and collapse through the past five million years. *Nature* 458:329–32

Pollard D, DeConto RM, Alley RB. 2015. Potential Antarctic Ice Sheet retreat driven by hydrofracturing and ice cliff failure. *Earth Planet. Sci. Lett.* 412:112–21

Pritchard H, Arthern R, Vaughan D, Edwards L. 2009. Extensive dynamic thinning on the margins of the Greenland and Antarctic ice sheets. *Nature* 461:971–75

Rathbun AP, Marone C, Alley RB, Anandakrishnan S. 2008. Laboratory study of the frictional rheology of sheared till. *J. Geophys. Res.* 113:F02020

Raymo ME, Mitrovica JX. 2012. Collapse of polar ice sheets during the stage 11 interglacial. *Nature* 483:453–56

Reyes AV, Carlson AE, Beard BL, Hatfield RG, Stoner JS, et al. 2014. South Greenland ice-sheet collapse during Marine Isotope Stage 11. *Nature* 510:525–28

Rignot E. 2006. Changes in ice dynamics and mass balance of the Antarctic ice sheet. *Philos. Trans. R. Soc. A* 364:1637–55

Rignot E, Bamber J, van den Broeke M, Davis C, Yonghong L, et al. 2008. Recent Antarctic mass loss from radar interferometry and regional climate modelling. *Nat. Geosci.* 1:106–10

Rignot E, Jacobs S. 2002. Rapid bottom melting widespread near Antarctic ice sheet grounding lines. *Science* 296:2020–23

Rignot E, Jacobs S, Mouginot J, Scheuchl B. 2013. Ice-shelf melting around Antarctica. *Science* 341:266–70

Rignot E, Koppes M, Velicogna L. 2010. Rapid submarine melting of the calving faces of West Greenland glaciers. *Nat. Geosci.* 3:187–91

Rignot E, Mouginot J, Morlighem M, Seroussi H, Scheuchl B. 2014. Widespread, rapid grounding line retreat of Pine Island, Thwaites, Smith, and Kohler glaciers, West Antarctica, from 1992 to 2011. *Geophys. Res. Lett.* 41:3502–9

Rignot E, Mouginot J, Scheuchl B. 2011. Antarctic grounding line mapping from differential satellite radar interferometry. *Geophys. Res. Lett.* 38:L10504

Rignot E, Steffen K. 2008. Channelized bottom melting and stability of floating ice shelves. *Geophys. Res. Lett.* 35:L02503

Robin GdQ. 1958. Glaciology III: Seismic shootings and related investigations. In *Norwegian-British-Swedish Antarctic Expedition, 1949–52: Scientific Results*, Vol. 5, pp. 1–134. Oslo: Nor. Polarinst.

Rohling EJ, Grant K, Hemleben CH, Siddall M, Hoogakker BAA, et al. 2008. High rates of sea-level rise during the last interglacial period. *Nat. Geosci.* 1:38–42

Rohling EJ, Haigh ID, Foster GL, Roberts AP, Grant KM. 2013. A geological perspective on potential future sea-level rise. *Sci. Rep.* 3:3461

Rovere A, Raymo ME, Mitrovica JX, Hearty PJ, O'Leary MJ, Inglis JD. 2014. The Mid-Pliocene sea-level conundrum: glacial isostasy, eustasy and dynamic topography. *Earth Planet. Sci. Lett.* 387:27–33

Scambos TA, Berthier E, Shuman CA. 2011. The triggering of subglacial lake drainage during rapid glacier drawdown: Crane Glacier, Antarctic Peninsula. *Ann. Glaciol.* 52:74–82

Scambos TA, Bohlander JA, Shuman CA, Skvarca P. 2004. Glacier acceleration and thinning after ice shelf collapse in the Larsen B embayment, Antarctica. *Geophys. Res. Lett.* 31:L18402

Scherer RP, Aldahan A, Tulaczyk S, Possnert G, Engelhardt H, Kamb B. 1998. Pleistocene collapse of the West Antarctic ice sheet. *Science* 281:82–85

Schmeltz M, Rignot E, Dupont TK, MacAyeal DR. 2002. Sensitivity of Pine Island Glacier, West Antarctica, to changes in ice-shelf and basal conditions: a model study. *J. Glaciol.* 48:552–58

Schmidtko S, Heywood KJ, Thompson AF, Aoki S. 2014. Multidecadal warming of Antarctic waters. *Science* 346:1227–31

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

Schoof C. 2007. Ice sheet grounding line dynamics: steady states, stability, and hysteresis. *J. Geophys. Res.* 112:F03S28

Scott RF. 1905. Results of the National Antarctic Expedition. I. Geographical. *Geogr. J.* 25:353–70

Sen Gupta A, Santoso A, Taschetto A, Ummenhofer C, Trevena J, England M. 2009. Projected changes to the Southern Hemisphere ocean and sea ice in the IPCC AR4 climate models. *J. Clim.* 22:3047–78

Sergienko OV, Hindmarsh RCA. 2013. Regular patterns in frictional resistance of ice-stream beds seen by surface data inversion. *Science* 342:1086–89

Shannon SR, Payne AJ, Bartholomew ID, van den Broeke MR, Edwards TL, et al. 2013. Enhanced basal lubrication and the contribution of the Greenland ice sheet to future sea-level rise. *PNAS* 110:14156–61

Shepherd A, Wingham D. 2007. Recent sea-level contributions of the Antarctic and Greenland ice sheets. *Science* 315:1529–32

Shepherd A, Wingham D, Payne T, Skvarca P. 2003. Larsen Ice Shelf has progressively thinned. *Science* 302:856–59

Shepherd A, Wingham D, Rignot E. 2004. Warm ocean is eroding West Antarctic Ice Sheet. *Geophys. Res. Lett.* 31:L23402

Siddall M, Rohling EJ, Thompson WG, Waelbroeck C. 2008. Marine isotope stage 3 sea level fluctuations: data synthesis and new outlook. *Rev. Geophys.* 46:RG4003

Stanton TP, Shaw WJ, Truffer M, Corr HFJ, Peters LE, et al. 2013. Channelized ice melting in the ocean boundary layer beneath Pine Island Glacier, Antarctica. *Science* 341:1236–39

Steig EJ, Ding Q, Battisti DS, Jenkins A. 2012. Tropical forcing of Circumpolar Deep Water inflow and outlet glacier thinning in the Amundsen Sea embayment, West Antarctica. *Ann. Glaciol.* 53:19–28

Straneo F, Curry RG, Sutherland DA, Hamilton GS, Cenedese C, et al. 2011. Impact of fjord dynamics and glacial runoff on the circulation near Helheim Glacier. *Nat. Geosci.* 4:322–27

Straneo F, Heimbach P. 2013. North Atlantic warming and the retreat of Greenland's outlet glaciers. *Nature* 504:36–43

Straneo F, Sutherland DA, Holland D, Gladish C, Hamilton GS, et al. 2012. Characteristics of ocean waters reaching Greenland's glaciers. *Ann. Glaciol.* 53:202–10

Strugnell JM, Watts PC, Smith PJ, Allcock AL. 2012. Persistent genetic signatures of historic climatic events in an Antarctic octopus. *Mol. Ecol.* 21:2775–87

Sutterley TC, Velicogna I, Rignot E, Mouginot J, Flament T, et al. 2014. Mass loss of the Amundsen Sea Embayment of West Antarctica from four independent techniques. *Geophys. Res. Lett.* 41:8421–28

Thoma M, Jenkins A, Holland D, Jacobs S. 2008. Modelling Circumpolar Deep Water intrusions on the Amundsen Sea continental shelf, Antarctica. *Geophys. Res. Lett.* 35:L18602

Thomas R, Frederick E, Li J, Krabill W, Manizade S, et al. 2011. Accelerating ice loss from the fastest Greenland and Antarctic glaciers. *Geophys. Res. Lett.* 38:L10502

Thomas RH, Bentley CR. 1978. A model for Holocene retreat of the West Antarctic Ice Sheet. *Quat. Res.* 10:150–70

Thompson WG, Curran HA, Wilson MA, White B. 2011. Sea-level oscillations during the last interglacial highstand recorded by Bahamas corals. *Nat. Geosci.* 4:684–87

Tulaczyk S, Kamb WB, Engelhardt HF. 2000. Basal mechanics of Ice Stream B, West Antarctica: 1. Till mechanics. *J. Geophys. Res.* 105(B1):463–81

Vaughan DG, Barnes DKA, Fretwell PT, Bingham RG. 2011. Potential seaways across West Antarctica. *Geochem. Geophys. Geosyst.* 12:Q10004

Vaughan DG, Comiso JC, Allison I, Carrasco J, Kaser G, et al. 2013. Observations: cryosphere. In *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*, ed. TF Stocker, D Qin, GK Plattner, M Tignor, SK Allen, et al. pp. 317–382. Cambridge, UK: Cambridge Univ. Press

Velicogna I, Wahr J. 2006. Measurements of time-variable gravity show mass loss in Antarctica. *Science* 311:1754–56

Walker RT, Dupont TK, Parizek BR, Alley RB. 2008. Effects of basal-melting distribution on the retreat of ice-shelf grounding lines. *Geophys. Res. Lett.* 35:L17503

Walker RT, Parizek BR, Alley RB, Anandakrishnan S, Riverman KL, Christianson K. 2013. Ice-shelf tidal flexure and subglacial pressure variations. *Earth Planet. Sci. Lett.* 361:422–28

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

Walker RT, Parizek BR, Alley RB, Brunt KM, Anandakrishnan S. 2014. Ice shelf flexure and tidal forcing of Bindschadler Ice Stream, West Antarctica. *Earth Planet. Sci. Lett.* 395:184–93
Weber ME, Clark PU, Kuhn G, Timmermann A, Sprenk D, et al. 2014. Millennial-scale variability in Antarctic ice-sheet discharge during the last deglaciation. *Nature* 510:134–38
Weertman J. 1957. Deformation of floating ice shelves. *J. Glaciol.* 3:38–42
Weertman J. 1974. Stability of the junction of an ice sheet and an ice shelf. *J. Glaciol.* 13:3–11
Wellner JS, Heroy DC, Anderson JB. 2006. The death mask of the Antarctic ice sheet: comparison of glacial geomorphic features across the continental shelf. *Geomorphology* 75:157–71
West RG, Sparks BW, Sutcliffe AT. 1960. Coastal interglacial deposits of the English Channel. *Philos. Trans. R. Soc. B* 243:95–133
White JWC, Alley RB, Brigham-Grette J, Fitzpatrick JJ, Jennings AE, et al. 2010. Past rates of climate change in the Arctic. *Quat. Sci. Rev.* 29:1716–12

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

Annual Review
of Earth and
Planetary Sciences

Volume 43, 2015

Contents

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.

Indexes

Errata

An online log of corrections to *Annual Review of Earth and Planetary Sciences* articles may be found at http://www.annualreviews.org/errata/earth

Annu. Rev. Earth Planet. Sci. 2015.43:207-231. Downloaded from www.annualreviews.org
Access provided by Harvard University on 03/07/16. For personal use only.