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Stability of the West Antarctic ice sheet in a warming world

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Ice sheets are expected to shrink in size as the world warms, which in turn will raise sea level. The West Antarctic ice sheet is of particular concern, because it was probably much smaller at times during the past million years when temperatures were comparable to levels that might be reached or exceeded within the next few centuries. Much of the grounded ice in West Antarctica lies on a bed that deepens inland and extends well below sea level. Oceanic and atmospheric warming threaten to reduce or eliminate the floating ice shelves that buttress the ice sheet at present. Loss of the ice shelves would accelerate the flow of non-floating ice near the coast. Because of the slope of the sea bed, the consequent thinning could ultimately float much of the ice sheet's interior. In this scenario, global sea level would rise by more than three metres, at an unknown rate. Simplified analyses suggest that much of the ice sheet will survive beyond this century. We do not know how likely or inevitable eventual collapse of the West Antarctic ice sheet is at this stage, but the possibility cannot be discarded. For confident projections of the fate of the ice sheet and the rate of any collapse, further work including the development of well-validated physical models will be required.

Recent satellite observations show that mass loss from the West Antarctic ice sheet (WAIS) (Box 1) ranges from 100 to 200 Gt yr^{-1} (equivalent to 0.28 to 0.56 mm yr^{-1} sea-level rise), with the rate growing over at least the past two decades[1–4]. These observations intensify concerns about the ice sheet's stability that were first raised more than four decades ago[5]. Although early attention focused on the ice stream drainages feeding the large Ross and Filchner–Ronne ice shelves (Box 1), at present these areas are thickening at moderate rates (several centimetres per year) or are near balance[6–9]. Instead, net WAIS losses result predominately from the strong thinning (tens to hundreds of centimetres per year) of the glaciers draining the ice sheet's Amundsen Sea sector (Fig. 1), which in the 1980s was presciently described as the "weak underbelly" of the WAIS[10]. Although the large changes in the Amundsen Sea region have recently shifted focus from the Ross and Filchner–Ronne drainages, all sectors of the WAIS are subject to potential instabilities, although varied sensitivities to ice–ocean–atmosphere forcings may produce regional losses with differing onset times and durations.

WAIS history as a possible prelude to the future

Mercer[5] first raised concern over a possible future WAIS collapse by suggesting that the ice sheet was not present for an extended period during the last interglacial, corresponding to Marine Isotope Stage 5e (MIS 5e). This conclusion was largely based on geologic evidence of ice-marginal lakes that he assigned to MIS 5e at elevations (1,400 m) in the Transantarctic Mountains where present temperatures produce only negligible summer melt, but that remain poorly dated to this day[11,12]. From this finding he concluded that temperatures were at least 7 °C warmer than present. He argued that such warming was sufficient to eliminate the floating ice shelves needed to buttress the WAIS (see Box 2), resulting in the nearly total loss of the ice sheet and a consequent sea-level rise of several metres[12].

A further argument by Mercer for a past WAIS collapse was that it would have raised sea level by an amount consistent with palaeo-sea-level records[5]. Relative to the Holocene epoch, East Antarctic ice cores show anomalous warmth during MIS 5e, and to a lesser extent, during the previous three interglacials. Analysis of the magnitude and pattern of MIS 5e sea-level indicators provides a 95% probability that the Greenland and Antarctic ice sheets each contributed at least 2.5 m of sea-level rise to the MIS 5e global peak of at least 6.6 m above modern, with a 67% probability that it exceeded 8.0 m[13]. Although this analysis cannot distinguish between East and West Antarctic sources, deglaciation of the marine portions of the WAIS could have raised sea level by up to ~3.3 m[14] (Fig. 2).

Although limited observations make it difficult to determine the magnitude of the WAIS contribution to sea level during MIS 5e, there is substantial evidence that the ice sheet has shrunk considerably in the past, but with poor knowledge of precisely when. One such indication comes from diatom and beryllium-10 concentrations from glacial sediments retrieved from WAIS boreholes that point to open-water conditions well inland of the Ross ice shelf in the past 750,000 yr, probably indicating loss of all or most of the ice sheet[15]. Other evidence comes from the patterns of similarity of bryozoans found in the ocean surrounding the WAIS[16]. These similarities suggest that open seaways existed through what is now the centre of the ice sheet during the late Quaternary period, perhaps during one or more of the past few interglacials. Furthermore, sediment cores from near the front of the Ross ice shelf provide a discontinuous history of ice-sheet fluctuations, documenting notably warmer open-water conditions as recently as ~1 million years (Myr) ago and repeatedly more than ~3 Myr ago at times identified with global temperatures up to ~3 °C warmer than today, when CO_2 levels were up to ~400 ppmv (ref. 17). Ice-flow modelling indicates that such open-water conditions were associated with periods when the WAIS was largely absent[18]. Thus, the palaeorecord strongly suggests that the WAIS largely disappeared, perhaps during the past few hundred thousand years and more confidently during the past few million years, in response to warming similar to or less than that projected under business-as-usual CO_2 emission scenarios for the next few centuries[19].

Although the palaeorecord does not directly constrain the rapidity with which a collapse would occur, on many parts of the

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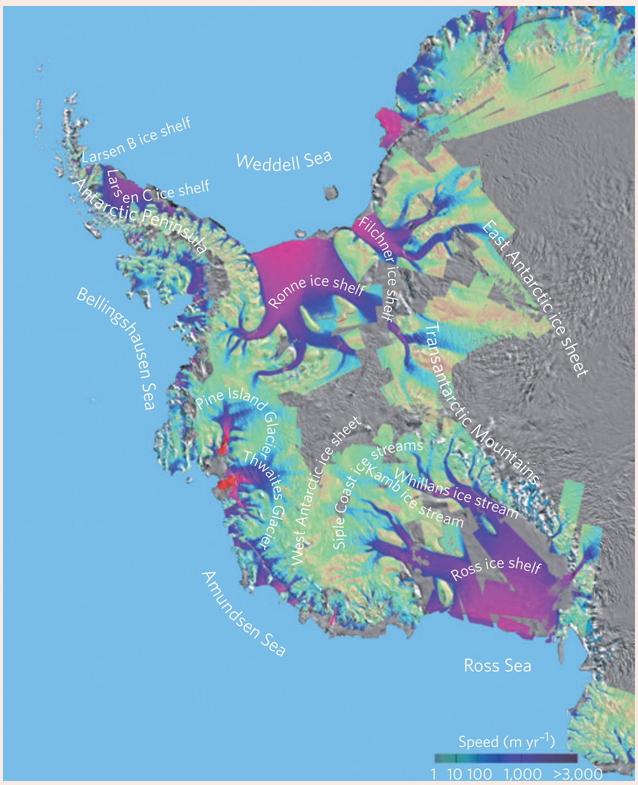
Box 1 | The West Antarctic ice sheet.

The Transantarctic Mountains geographically divide the large ice sheet covering Antarctica into two distinct parts, each of which is distinguished by its respective hemisphere. The far larger East Antarctic Ice Sheet lies largely on continental crust that is above sea level or would rebound above sea level if the ice sheet were removed[12,14] (see Fig. 2 for distribution of area above and below sea level). With only about 10% of the total Antarctic ice volume, the far smaller marine-based WAIS lies on a bed, much of which is many hundreds of metres below sea level (Fig. 2) and is draped by marine sediments[14,34], making it susceptible to the instabilities described throughout this Review. Although Mercer initially estimated that a collapse would produce a rise in sea level of 4–6 m (ref. 5), more recent analysis suggests collapse of only the unstable marine portions would produce a 3.3-m eustatic sea-level rise[14]. The main features of the WAIS are plotted on the map, which shows ice-flow speed (colour) over the MODIS-based Mosaic of Antarctica[104] (grayscale).

An ice sheet consists of thick, slow-moving inland ice feeding faster-moving (>100 m yr^{-1}), channelized ice streams (see map), which flow across grounding lines into floating extensions called ice shelves, from which icebergs calve. In West Antarctica, ice flow to the Weddell and Ross seas is largely concentrated through the Texas-sized Filcher–Ronne and Ross floating ice shelves that periodically (years to decades) calve Rhode-Island-sized icebergs. Along the less deeply embayed Amundsen Sea Coast, ice streams feed numerous smaller fringing ice shelves, which produce smaller icebergs more frequently. In addition to calving, most ice shelves lose much of their mass through basal melt where they are in contact with warm ocean water[54,62].



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continental shelf the retreat history during the last deglaciation included a series of rapid stepped retreats separated by periods when large sedimentary wedges were deposited, suggesting stillstands of centuries or longer[20]. The forcing, however, for such past events was slower than may occur in the future.

Playing out over a century, a WAIS collapse would produce severe economic consequences for many coastal communities, whereas a similar occurrence over a millennium would yield a far more manageable transition[21]. Thus, understanding the rate of ice loss is essential to developing an appropriate risk-mitigation strategy[22]. Barring new palaeo-evidence on past rates of collapse, projecting future rates will probably have to rely on models derived from a firm physical understanding of the processes leading to collapse. Major strides have been made, but such an understanding is far from complete at present, as evidenced by the recent lack of an upper bound on possible sea-level rise in the fourth Intergovernmental Panel on Climate Change (IPCC) Assessment[19].

Instability mechanisms

Removing the WAIS would leave broad, deep seaways that deepen towards the ice-sheet interior (Fig. 2). This bathymetry makes the ice sheet subject to the marine-ice-sheet instability (Box 2). Portions of the East Antarctic and Greenland ice sheets are also marine, and the discussion here applies to those regions as well, but the issue is quantitatively more important for the WAIS, with its extensive troughs extending to depths of more than 2 km[14] (Fig. 2). Despite its marine setting, the present ice sheet exists because various factors promote stability, including buttressing ice shelves and regions where local bathymetric slopes oppose the general trend. Climate forcing, in particular warming that affects ice-shelf viability, could undo this potentially fragile stability. Internal instabilities also have the potential to push the WAIS past a threshold where the marine-based instability may lead to irreversible retreat.

Internal instabilities. Through much of the Holocene the WAIS shrank with little or no climate or sea-level forcing[23,24]. In its present state, however, model results together with palaeoclimate evidence suggest that the WAIS can exist stably in a state similar to the modern[18], which in turn suggests that future changes are likely to be forced. Nonetheless, future internal instabilities cannot be excluded entirely, so we consider these before assessing possible forced instabilities.

More than three decades of research on the Siple Coast (Ross) ice streams has revealed a complicated flow history, much of which was driven by internal dynamics that, for the most part, seem to have acted independently of climate change. Ground-based radar observations of buried shear margins provided an early indication of this variability, by revealing that in around AD 1850 the now-stagnant Kamb ice stream was flowing at high (hundreds of metres per year) speeds comparable to those of nearby ice streams[25,26] (Box 1). Several other instances of past flow changes have been inferred from ground-based radar measurements[27,28]. Furthermore, velocity measurements on the adjacent Whillans ice stream indicate that a slowdown has been ongoing for more than three decades, which if sustained would cause this ice stream to stagnate in about 70 yr (ref. 29). Further evidence comes from satellite image mosaics of the entire Ross ice shelf, which provide a synoptic view of flow stripes (traces of former flow directions) that deviate strongly from the modern flow field[30]. Experiments with ice-shelf models reveal that these patterns are a result of the stagnations and re-activations of several ice streams over the past millennium[31,32]. Although the degree of variation is large, the switching on and off of ice streams may have had little net effect on grounding line retreat over the past millennium[31], suggesting little in the way of an ongoing unforced retreat.

Early WAIS research focused on internal instabilities that might lead to variable flow, including hypothesized ice-stream surges, which ultimately could produce ice-sheet collapse[33]. Later the discovery that the Siple Coast ice streams were underlain with weak deformable till[34–37] over much of their area[38,39] suggested other mechanisms for internal instability. One important consequence of weak beds is that much of the driving stress is not supported locally, and instead, resistance is concentrated at the stronger margins[40] or isolated sticky spots[41]. Where the margins dominate resistance, ice-stream speeds are proportional to the fourth power of the width[42]. As a consequence, a small degree of widening can produce large changes in speed. Thus, above a critical speed, shear heating may yield a positive feedback such that as an ice stream widens, the resulting increase in margin shear heating would lead to further widening and speedup[43].

Thinning caused by ice-stream widening and speedup also may introduce a stabilizing feedback at some locations on the Siple Coast and in other similar regions. In such locations, ice-stream thinning steepens basal temperature gradients enough to conduct more heat away from the bed than is supplied through the geothermal heat flux and friction from sliding, causing basal freezing to exceed the supply of meltwater from other areas[44,45]. The resulting withdrawal of water from the basal till can alter till porosity by a few per cent to increase basal resistance to levels greater than that of the extremely low ice-stream driving stresses (a few kilopascals), forcing ice-stream stagnation[44]. Thus, this sensitivity of till to water content may produce a thermal cycling of ice streams whereby thinning causes basal freezing that stops fast flow[31,45,46]. Following such a stagnation, the resulting thickening traps geothermal heat, increasing melting and reducing till strength to reactivate the ice stream and repeat the cycle.

Although other factors may play a role, thermally driven cycling of ice streams can account for much of the variability over the past millennium[30,31], with similar processes possibly having given rise to Heinrich events from the former Laurentide ice sheet[47]. Such internal instabilities will probably modulate sea level over the coming centuries, with present thickening now offsetting some of the thinning elsewhere[1,8,9]. The flowstripe record[30] from the past millennium indicates that these fluctuations are likely to have a relatively modest effect (tenths of a millimetre per year or less of sea-level variability) and will probably not lead to an ice-sheet collapse independent of climate or as part of an ongoing response to the last deglaciation.

Ocean-forced instabilities. The advance and retreat of the Antarctic ice sheet were long believed to have been driven largely by changes in sea level in response to the growth and decay of Northern Hemisphere ice sheets[48,49]. When forced with sea-level change following the Last Glacial Maximum (LGM), models simulate WAIS retreat at a rate roughly consistent with what is known about the retreat history[50]. Although > 100-m changes in sea level probably contributed to post-LGM retreat, it is unlikely that sea-level changes of several metres from non-WAIS sources will drive future WAIS losses, nor will WAIS losses introduce a positive feedback with sea level that could drive further retreat over the next few centuries[51,52].

With average summer surface temperatures well below freezing over most of the WAIS, the role of surface melt in ice loss is negligible at present[53]. Instead, the ice sheet sheds mass gained annually from snowfall by discharging ice to the ocean, where it either breaks off to form icebergs or melts *in situ* beneath the floating ice shelves[54]. Once ice crosses the grounding line and begins to float, its direct contribution to sea level is negligible. Nonetheless, the rate of discharge across the grounding line is strongly determined by the degree to which melting and iceberg calving influence the buttressing restraint provided by ice shelves[55,56].

Several processes control rates of sub-ice-shelf melt[54]. For many ice shelves, especially the largest ones, much of the basal melting is driven by high-salinity water produced during sea-ice formation over the adjacent continental shelf. This dense water sinks to the deep grounding lines, where it melts ice because the pressure melting point is lowered by roughly 0.75 °C km^{-1} below the surface[54,57]. The resulting fresh water from melting lowers the density, causing the water to rise buoyantly along the ice-shelf base. Turbulent mixing of heat into the boundary layer can drive additional melting, especially where the ice-shelf base is steep. The buoyant boundary

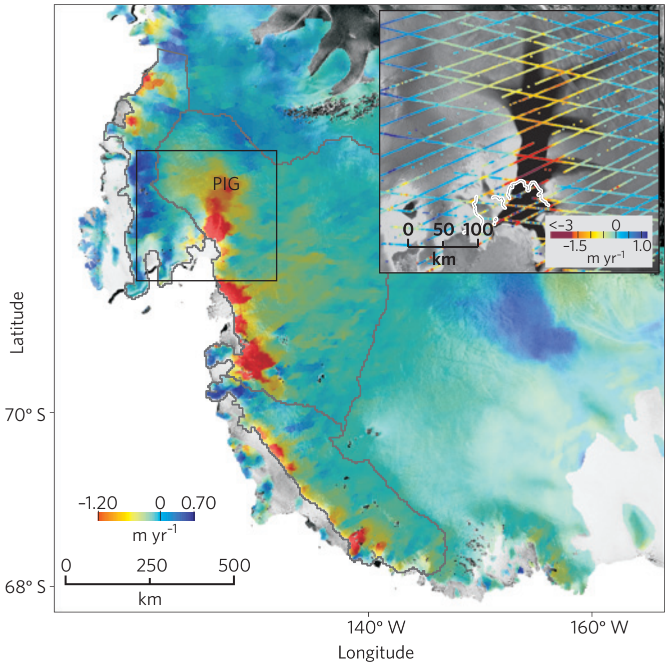

Figure 1 | Rates of elevation change along the Amundsen Coast of West Antarctica as determined from ICESAT by Pritchard *et al.*[9]. The results are gridded from individual tracks such as those shown in the inset for PIG. Figure reproduced from ref. 9, © 2009 NPG.

current may eventually supercool as it rises along the pressure-melting-point gradient, refreezing to form marine ice beneath the shallower parts of the shelf[58]. Such refreezing reduces the net melt, particularly beneath the Filchner–Ronne ice shelf where marine ice contributes nearly half the ice thickness in some parts of the central shelf[58–60]. A second contributor to sub-ice-shelf melting is mixing of shallow waters beneath the ice-shelf front by tides, long-period waves or other processes. In the case of tidal mixing, heat from seasonally warmed, near-surface waters causes melting even when water temperatures in most of the sub-ice-shelf cavity are near freezing[54,59–61]. Finally, warm (3–4 °C above the *in situ* melting point) Circumpolar Deep Water (CDW) produces large melt rates in the regions where it is able to access the sub-ice-shelf cavities[54].

At present, little heat from CDW makes its way beneath the Ross and Filchner–Ronne ice shelves, so the sea-ice-related circulation near the grounding lines and tidal mixing near the ice-shelf fronts cause most of the melting beneath these large ice shelves[54]. Combined, these processes yield ice-shelf-wide average melt rates of tens of centimetres per year, with rates near grounding lines and at ice-shelf fronts roughly an order of magnitude higher[54,59,60,62]. Because higher surface temperatures might reduce the sea-ice formation that produces the dense high-salinity waters that now cause melting beneath the large ice shelves, one hypothesis postulates that a warmer climate might actually lessen basal melt rates near the deep grounding lines[63]. More recent results, however, indicate that the production rate of high-salinity water may be less important to melting than the processes that control the flow of this dense water beneath the ice shelves[57]. Closer to the ice-shelf front, seasonally or annually higher near-surface ocean temperatures or faster mixing by waves or tides could enhance melting. Because this thinning is concentrated near the ice-shelf front, it should have less effect on ice flow than melt near the deep grounding lines[64].

In contrast to the far larger Ross and Filcher–Ronne ice shelves, substantial volumes of CDW do migrate beneath the Amundsen Sea ice shelves, producing high (> 5 m yr^{-1}) average melt rates[62], with modelled and estimated melt rates near the grounding line

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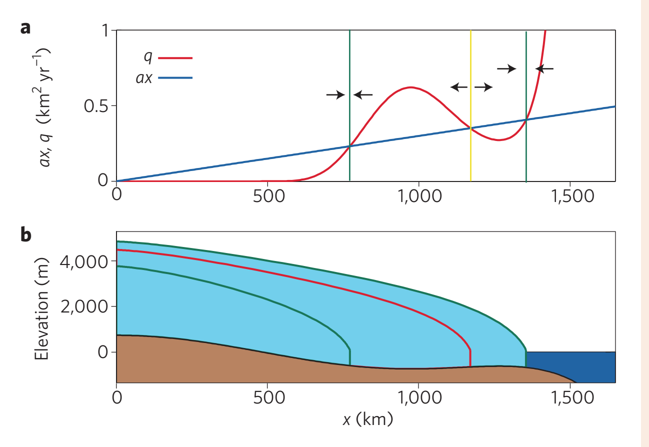
Box 2 | Marine-ice-sheet instability.

Like any pile, an ice sheet tends to spread under its own weight. This spreading tendency is restrained mainly by friction beneath inland ice, and by along-flow stresses in ice shelves, both of which are important restraints for ice streams. The difference between local basal restraint of inland ice and longitudinal stretching of ice shelves gives rise to the marine-ice-sheet instability, in which a freely spreading ice shelf cannot have a stable grounding line on a 'reversed' bed that deepens towards the ice-sheet centre[77,105].

The rate at which a grounding line discharges ice to an unconfined ice shelf increases nonlinearly with thickness (see figure), but the local frictional control of inland ice prevents immediate response to remote events at the grounding line. The basic principles of an ice sheet that feeds an unbuttressed ice shelf are illustrated with a simple model and figure by Schoof[77]. **a**, In steady state, the grounding-line discharge, q (red curve), which is dependent on the thickness of the grounding line, must match the balance flux, ax (blue line), which in this 2D example is the product of a spatially invariant accumulation rate, a, and the upstream catchment length, x. Steady state is achieved where $q = ax$, which occurs at three points in this example (indicated by green and red vertical lines), which also correspond to the ice-sheet steady-state profiles shown in **b**.

The arrows at the steady-state points indicate the direction of migration in response to an initial grounding-line perturbation. For a steady-state grounding line positioned on a forward slope (green profiles), a retreat reduces the discharge below the balance flux, causing the grounding line to thicken and re-advance. Similarly, advance down a forward slope causes $q > ax$, so that the ice sheet thins and retreats back to its steady-state position. The situation differs for a bed with a reverse slope (red profiles). In this case, an initial retreat increases discharge while reducing the balance flux, leading to grounding line thinning and further retreat. In this case, the retreat would continue until the upstream steady-state point is reached. Conversely, an advance slows discharge so that it exceeds the balance flux, promoting further advance.

The bed of a marine ice sheet steadily deepens towards the interior in many locations, with no stabilizing forward slope to stop a retreat once it is initiated. The analysis just described applies to a marine ice sheet with an unconfined ice shelf (that is, one that provides no buttressing). A marine ice sheet can be stabilized, however, where a confined ice shelf provides sufficient 'buttressing', arising from friction between its floating ice and its embayed walls[55]. Alternatively, highs or 'bumps' in the bed that locally offset the regional bed slope may stabilize the ice sheet[51,56]. Where both an ice shelf and a 'bump' act in concert to promote stability, a reduction in buttressing from warming-induced ice-shelf melting may cause threshold behaviour, with grounding-line migration initially restricted to the local bump and then jumping rapidly upstream to the next bump or farther. Figure reproduced with permission from ref. 77, © 2007 AGU.



of Pine Island Glacier (PIG) exceeding 100 m yr^{-1} (refs 65,66). For these shelves, melt rates are determined largely by the volume of CDW reaching their respective cavities[67], with ocean-circulation models suggesting much seasonal to intra-annual variability on the Amundsen Sea continental shelf[68]. In particular, the model estimates show a strong increase in the available heat for melting from the late 1990s onwards, which corresponds to a period over which the Amundsen Sea ice shelves thinned at rates ranging from 0.6 to 5.5 m yr^{-1} (ref. 69) and when many of the large glaciers feeding them sped up substantially[69,70] to produce most of Antarctica's present net contribution to sea-level rise[1,9].

Some of the recent ice-shelf thinning along the Amundsen Coast may be attributable to the ~0.2 °C warming of nearby ocean CDW[62,67,71]. Simulations of future warming have large uncertainties, but generally produce increases of at least a few tenths of a degree Celsius this century[72] and a few degrees Celsius by 3000[73] with continuing CO_2 emissions[74,75]. Thus, direct warming of CDW becomes an increasing threat to ice-sheet stability on greater than decadal timescales.

Much of the present variability in heat available for ice-shelf melting seems to be the result of seasonal to interannual variation in the transport of CDW from offshore into the subshelf troughs, rather than an actual warming of the CDW[67,68]. This variability is probably driven by shifts in wind patterns, including those linked to changes in position and strength of the Amundsen Sea low, which modulate onshore CDW flow[68]. Whether the recent increases in CDW transport represent multidecadal variability or a longer-term trend is unknown, though at least some models suggest enhanced winds increasing southward transport of warm waters perhaps reaching the ice shelves by 2100[73,74]. Thus, some combination of direct warming and enhanced transport of CDW seems likely to increase future rates of ice-shelf melt.

Increased basal melt can directly increase flow speed of inland ice by reducing the thickness, and thus, the buttressing restraint of an ice shelf[55,56,76]. Often CDW is limited to the deeper waters in the cavity[67,68], so the net effect may be to thin rather than eliminate an ice shelf, because a thinner shelf has less exposure to the deeper, warmer waters. Several decades of enhanced melting may be required to produce a substantial loss of buttressing. For example, on PIG, the large (nearly 4 m yr^{-1}) observed thinning rate reduced buttressing by at most a few per cent per year[69], making it hard to account for the observed speedup through recent loss of buttressing alone[77].

Reduced buttressing from increased melting may trigger other destabilizing feedbacks. For example, thinning at the grounding line may cause it to retreat, leading to a loss of basal friction beneath the newly ungrounded ice. This loss of resistive stress is similar to a loss of buttressing in that flow speed must increase to produce compensating resistive stresses elsewhere, yielding more thinning and retreat. For example, although it is probably a consequence of basal melting, much of the speedup of PIG can be accounted for by grounding-line retreat[78] rather than loss of resistive stress from the ice shelf[66,76,79]. This rapid response on PIG may have been preconditioned by the formation of a low-slope ice plain[80] over several decades or more[66,81]. A further feedback is that new ungrounding exposes more area to ocean-induced basal melting[67,81]. Such changes on an ice shelf and near its grounding line can produce strong thinning extending hundreds of kilometres inland in just a few years[66,79,82,83].

If the grounding line is perched at a stable point near a bed high where it is able to retreat into ever-deeper water, the processes and feedbacks just described may allow a small initial retreat to trigger much greater retreat through the marine-ice-sheet instability (Box 2). For example, several Amundsen Coast glaciers have deep troughs extending inland all the way to near the divide[84,85], so that once started, retreat may be self-sustaining. As an example, numerical simulations suggest that recent speedups on PIG may continue unabated at least through this century if high melt rates continue, but could slow if melt rates decline so that the shelf thickens and the grounding line readvances[66]. Although the recently accelerated ice losses, marine geometry and increased melt rates suggest that ice–ocean interactions pose a threat to this sector of the ice sheet, considerable uncertainty remains about whether the recent changes in CDW transport represent decadal scale variability or a longer-term trend.

Atmosphere-forced instabilities. Mercer's hypothesized sensitivity of the WAIS to warming was based largely on the observation that ice shelves along the Antarctic Peninsula exist only south of the 0 °C January isotherm[5,21], later determined to be approximately equivalent to the mean annual −9 °C isotherm[86]. Over the four decades since this initial hypothesis, the Antarctic Peninsula has warmed substantially (0.056 °C yr^{-1}), forcing a southward shift of both of these isotherms[87]. This shift in temperature coincided with the progressive retreat of several ice shelves over the newly warmed regions[87–92], further supporting the idea of a thermal limit on ice-shelf viability.

Several studies have noted that in addition to higher temperatures, high surface melting and ponded water are strongly associated with catastrophic ice-shelf breakup[21,88,91]. Although there are many ways that meltwater might affect an ice shelf[21], the most likely is through growth of crevasses by hydrofracturing[91,92]. Both theory[93] and observation[94] indicate that water's greater density relative to ice will overpressurize a sufficiently deep and water-filled crevasse, causing it to fracture through ice a kilometre or more thick as long as there is water available to keep the crack full[95]. For example, satellite imagery shows numerous melt ponds near surface crevasses on the Larsen B ice shelf just before its rapid collapse in 2002[89,91] (Fig. 3), which probably contributed to fracturing of the full ice-shelf thickness just before the breakup[87,89]. During the collapse, the toppling of the tall, narrow, hydro-fractured slabs like cascading dominos released a large store of potential energy, which seems to have contributed both to further breakup and rapid, widespread dispersal of the resulting ice-shelf detritus[90].

Although hydro-fracturing may have been the main contributor to breakup of the Larsen B and some other Antarctic Peninsula ice shelves, other factors probably contributed to the occurrence and rapidity of these events. Before it collapsed, sub-ice-shelf melting may have thinned the Larsen B ice shelf[96]. Furthermore, the dynamics of an embayed ice shelf are such that the interior parts act as a compressive arch, transferring stresses to the sides of the embayment, and this tends to produce instability once the calving front retreats into the region forming the arch[88]. Thus, initial ice-shelf losses produced by hydrofracturing might accelerate as the ice front retreats past a critical point[88].

Observations of ice-shelf retreat over the past few decades have served to strengthen Mercer's conjecture of vulnerability to warming[5,21]. Peninsula ice-shelf losses have already produced increased outflow from many of the formerly buttressed glaciers[97,98] (Fig. 3). Because of the high snowfall in the Peninsula, and because much of the bedrock is well above sea level, which precludes a marine-ice-sheet instability, it is likely that much of the ice in the region will survive the loss of remaining ice shelves, transitioning to a tidewater regime similar to much of Greenland. Losses would still increase in a warming climate, but not as part of a catastrophic response to ice-shelf loss.

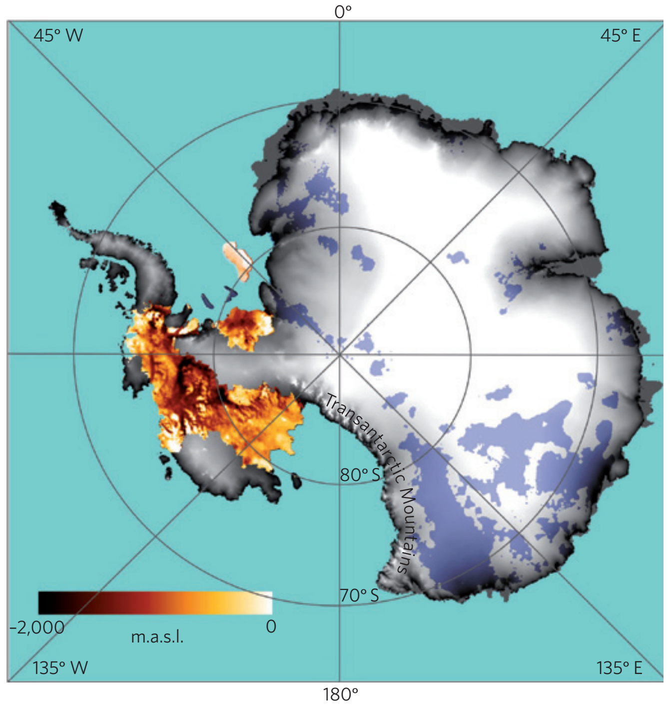


Figure 2 | Map of Antarctica with red-brown colours indicating the bed elevation of the marine portions of the WAIS[14]. The gray shaded regions show the East Antarctic ice sheet and the portions of West Antarctica likely to exist following a collapse of the marine ice sheet. Blue shaded regions show areas of East Antarctica where the bed is more than 200 m below sea level. Figure reproduced with permission from ref. 14, © 2009 AAAS.

Summertime melting is common on parts of the Ross and Filchner–Ronne ice shelves[99], but neither experiences notable ponding of meltwater[91]. Although the −9 °C mean annual and 0 °C January isotherms have been roughly equated in the peninsula, the equivalence may differ substantially from other regions owing to differing seasonality[87]. For example, January and mean annual temperatures differ by more than 9 °C on the Ross and Filchner–Ronne ice shelves[100]. Thus, the January isotherm should be a far more widely applicable predictor of ice-shelf viability along with other indicators of melt such as the number of positive-degree days[53].

The ability to pond water seems to be a stronger predictor of ice-shelf susceptibility than the actual volume of melt, with the extent of ponding depending on past melt history and the amount of firn available to absorb melt before ponding occurs[91]. As a result, the substantially lower accumulation rates on the Ross and Filchner–Ronne ice shelves[101] may mean that water can pond at slightly lower (for example, by ~1–2 °C) mean summer temperatures relative to the high-accumulation peninsula ice shelves. Even so, based on satellite estimates of temperature[100] and consistent with Mercer's initial analysis, substantial (~5–7 °C) summer warming probably would be required to threaten the largest WAIS shelves.

The WAIS over the next millennium

Projecting the contribution of the WAIS to sea-level change over the next few centuries has motivated much of the research focused on this ice sheet over the past few decades. Decades of research have narrowed many uncertainties, and heightened concern over others. Collectively, this body of research suggests that the ice sheet is not experiencing unforced mass loss due to internal instabilities or as part of an ongoing response to warming from the Last Glacial Maximum. Forcing by atmospheric or oceanic warming, however, could produce large future contributions of the WAIS to sea-level rise.

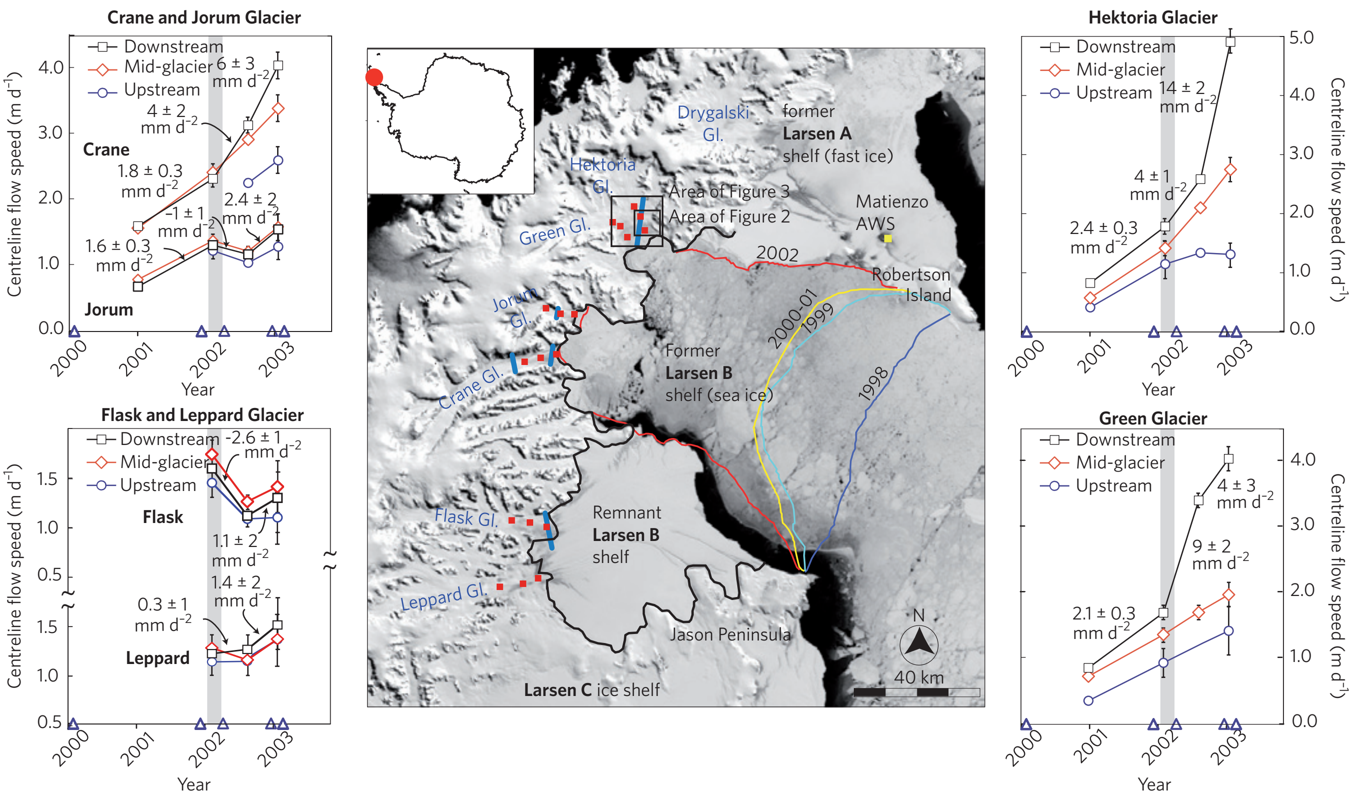


Figure 3 | MODIS image (centre panel) with history of retreat leading up to the collapse of the Larsen B ice shelf[98]. The side plots show the flow speed before and after the collapse. Note there was little or no speedup on Flask and Leppard glaciers, which flow into the remnant ice shelf and remain well buttressed. Figure reproduced with permission from ref. 98, © 2004 AGU.

Changes over the next century are of most immediate concern for policymakers[19], but few quantitative projections exist. The increasing rate of ice loss[4] from the WAIS seems to be largely due to changes in ice thickness near the grounding line that propagate inland. These changes are initially due to longitudinal stresses that produce strong thinning, and subsequently migrate inland by diffusion as slopes steepen[9,66,69,76,79]. For computational efficiency, models of the whole ice sheet tend to neglect the longitudinal stresses, and consequently do not adequately simulate the processes driving present change[102]. As a result, projections for the twenty-first century typically show a net decrease in sea level from Antarctica[19], despite present estimates of an increasingly negative mass balance[1,3,4].

In the absence of projections based on ice-sheet models that are in agreement with observations, other approaches have been employed to assess future ice-sheet mass loss. For example, using risk estimation protocols and an expert panel, Vaughan and Spouge[22] estimated a 5% chance of a WAIS contribution to sea-level rise in excess of 10 mm yr^{-1} over the next two centuries, although the elicitation was conducted before many of the recent changes in WAIS.

Pfeffer *et al.*[103] used a heuristic approach to estimate upper limits on sea-level rise in the twenty-first century. They assumed that Amundsen Coast glaciers accelerate to hypothesized maximum rates of flow over the next decade and remain at these elevated speeds for the remainder of the century. Using different scaling scenarios to determine maximum flow rates, they estimated an upper bound for twenty-first-century sea-level contributions from the WAIS ranging from 1.1 to 3.9 mm yr^{-1}. At present, PIG accounts for about 40% (0.13 mm yr^{-1} sea level rise) of the losses from the Amundsen Coast[70]. However, experiments with a numerical model of this glacier[66], which included longitudinal stresses and extreme oceanographic forcing (for example, melting), suggested twenty-first-century losses substantially less than 1.1 mm yr^{-1}.

Any future rapid losses, just like present losses[69,79], will most likely be initiated by a reduction in ice-shelf buttressing as warmer ocean waters flood ice-shelf cavities[68]. These losses can be amplified by additional feedbacks triggered once thinning commences (Box 2). Because Thwaites Glacier is more directly connected to the deep interior basins than the adjacent PIG, however, further studies will be required before contributions of the WAIS to future sea-level rise can be estimated with high confidence.

The bounds developed for the twenty-first century[103] exclude the loss of the large ice shelves that Mercer[5,21] envisioned as being responsible for the ice sheet's past and future demise. This omission is reasonable: simulated atmospheric warming in climate models[19,73] is much smaller than the rise in summer temperature, of about 5–7 °C, required to cause extensive meltwater ponding and hence fracturing of the Ross and Filchner–Ronne ice shelves. Numerical simulations often show that end-of-century warming and sea-ice reduction are especially small near the Ross ice shelf[19].

Nonetheless, end-of-century temperatures begin to approach thresholds of ice-shelf viability in many simulations[53]. Furthermore, large uncertainties remain about model skill at high southern latitudes, for atmospheric and oceanic circulation as well as temperature[22]. Ice-sheet simulations suggest that loss of the large ice shelves by atmospheric or oceanic forcing would probably presage collapse of the bulk of the marine ice sheet[18].

A collapse of the marine ice sheet in West Antarctica would raise sea level by more than three metres over the course of several centuries or less[14]. Such an event seems possible, but improved understanding of the expected atmospheric and oceanographic forcing and the ensuing ice-sheet response is required to quantify its likelihood. Precisely understanding the vulnerability of the West Antarctic ice sheet to a warming climate remains a grand challenge for the ice-sheet and climate-modelling communities.

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