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The Greenland and Antarctic ice sheets under 1.5 °C global warming

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Even if anthropogenic warming were constrained to less than 2 °C above pre-industrial, the Greenland and Antarctic ice sheets will continue to lose mass this century, with rates similar to those observed over the past decade. However, nonlinear responses cannot be excluded, which may lead to larger rates of mass loss. Furthermore, large uncertainties in future projections still remain, pertaining to knowledge gaps in atmospheric (Greenland) and oceanic (Antarctica) forcing. On millennial timescales, both ice sheets have tipping points at or slightly above the 1.5–2.0 °C threshold; for Greenland, this may lead to irreversible mass loss due to the surface mass balance–elevation feedback, whereas for Antarctica, this could result in a collapse of major drainage basins due to ice-shelf weakening.

Projections of future sea-level rise (SLR, Box 1) are primarily hampered by our incomplete knowledge of the contributions of the Greenland and the Antarctic ice sheets (GrIS and AIS, respectively), Earth's largest ice masses. In this Review we consider the potential contribution of both ice sheets under a strongly mitigated climate change scenario that limits the rise in global near-surface temperature to less than 2 °C above pre-industrial levels (targeting 1.5 °C), as agreed at the Twenty-first Conference of the Parties to the UNFCCC in Paris. We base the evaluation on both present-day observed/modelled changes and future forcings according to the Representative Concentration Pathway (RCP) RCP2.6 scenario. We use RCP2.6, the most conservative of the four RCPs of GHG concentration trajectories adopted by the IPCC for its Fifth Assessment Report (AR5), because it is the scenario in the published literature that best approximates the above warming range. Ice-sheet mass balance is defined as the net result of all mass gains and losses, and surface mass balance (SMB) as the net mass balance at the ice-sheet surface (where a negative mass balance means mass loss), including the firn layer. Hence, SMB does not include dynamical mass loss associated with ice flow at the ice-sheet margin or melting at the ice–ocean interface. Increased ice flow accounts for about one-third of the recent GrIS mass loss[1]. For Antarctica, where mass lost through ice discharge past the grounding line (the limit between the grounded ice sheet and floating ice shelf) is relatively evenly shared between oceanic basal melt before reaching the ice front and iceberg calving, increased ice flow accounts for all of the recent mass loss[2,3].

In the following sections we synthesize: (1) the latest available evidence of GrIS and AIS mass balance changes together with possible climate forcings from the atmosphere/ocean; and (2) the expected responses of the ice sheets under conditions of limited (1.5 °C) global warming by 2100. In the concluding section, we highlight outstanding issues that require urgent attention by the research community to improve projections.

Greenland forcing and mass balance changes

Greenland has warmed by ~5 °C in winter and ~2 °C in summer since the mid-1990s[4], which is more than double the global mean warming rate in that period. The GrIS has also been losing mass at an increasing rate since the 1990s[5] with 0.65–0.73 mm yr^{-1} of mean SLR equivalent (sle) for 2012–2016[6]. Since 2000, both SMB decrease and ice discharge increase contributed to mass loss, but the relative contribution of SMB decrease to the total mass loss went up from 42% to 68% between 2000 and 2012[1]. The current observed SMB decrease is mainly driven by increased melt and subsequent runoff[7] and is in part attributed to anthropogenic global warming and concurrent Arctic amplification (exacerbated Arctic warming due to regional feedbacks of global warming), but also to recent atmospheric circulation changes in summer observed since the 2000s[8]. The occurrence of a negative North Atlantic Oscillation (NAO) and a concurrent positive phase of the East Atlantic Pattern since 2000 can be interpreted as a weakening and southward displacement of the jet stream[9,10], allowing for anomalous high pressure[8] and enhanced atmospheric blocking[11] over the GrIS. These circulation changes in summer have favoured the advection of warm southerly air masses[12] and increased incoming solar radiation[13], leading to more melt, which is further enhanced by the melt–albedo feedback. The relative contribution of global warming and natural

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Box 1 | Projections of ice-sheet mass loss

Projections of ice-sheet contributions to SLR are established using ice flow models that compute the evolution of ice sheets under given climate scenarios. Many of these models were constructed to study the evolution of ice sheets across glacial–interglacial cycles, and are not therefore ideally suited to making projections for this century. Accordingly, the past decade has seen the modelling community repurpose these models, increasing confidence in the skill of ice-sheet models (particularly in the interactions with boundary conditions, such as ice–ocean and ice–bedrock), but they still lag behind other areas of the climate system.

Atmospheric and oceanic forcings are the primary drivers of ice-sheet change, and knowledge of the evolution of precipitation and surface melt is obtained from regional or global circulation models or parameterizations, whereas ocean circulation models or parameterizations are used to provide melt at the front of marine-terminating glaciers and the underside of floating ice shelves. Accurate information on the properties of the substrates underlying ice sheets (such as bedrock elevation and sediment rheology) are also important in determining reliable estimates of ice-sheet evolution.

For low-emissions scenarios and the near term, the initial state used by ice-sheet models is a key control on the reliability of their projections, because the anticipated mass loss is relatively small in comparison to the total mass of the ice sheets. Two main families of initialization strategies are employed at present. The first is spin-up of the model over glacial–interglacial periods, which ensures that the internal properties of the ice sheet are consistent with each other but may provide an inaccurate representation of the ice sheets' contemporary geometry and velocity. The alternative is the assimilation of satellite data, which may lead to inconsistencies in flow properties but has a greatly improved representation of the current geometry and surface velocity. These two approaches lead to large differences in the initial conditions from which projections are made and therefore create a significant spread in projected contributions to future SLR — even when forced with similar datasets[29,94]. Disentangling the impacts of natural variability and forced climate change is also more difficult for these low-emissions scenarios, but new model intercomparisons tend to focus on this aspect[95].

climate variability to the recent atmospheric circulation changes in Greenland remains an open question[14]. However, the Coupled Model Intercomparison Project Phase 5 (CMIP5) models do not exhibit such circulation changes, either in future warming scenarios or in present-day simulations[12]. This explains why the recent observed SMB is lower and runoff higher than predicted by these models (Fig. 1a, b).

The fact that climate models have limited skill in representing future changes in the North Atlantic jet stream[9] also affects how well clouds and precipitation over Greenland are simulated in future scenarios. The general relation between precipitation and temperature (+5% K^{-1}) derived using CMIP5 future projections[12] is subject to modification by structural changes in the North Atlantic atmospheric polar jet stream. Moreover, model (mis-)representation of clouds has a major effect on projected melt and runoff[15]. In one CMIP5-forced regional climate model, runoff depends linearly on temperature for low-warming scenarios (Fig. 1b). In this model, runoff from the GrIS at the end of the twenty-first century is estimated at around 1 mm yr^{-1} sle (360 Gt yr^{-1}) for the +1.5 °C scenario. These end-of-century temperature and runoff values are close to what is currently observed, which may be attributed to the recent circulation changes mentioned above.

A decrease in SMB lowers the ice-sheet surface, which in turn lowers SMB because at lower elevations, near-surface air temperature is generally higher[16,17]. Additional SMB changes due to the SMB–surface elevation feedback are small for limited warming: in a coupled SMB–ice dynamical simulation, the feedback contributes 11% to the GrIS runoff rate in an RCP2.6 scenario, or ~3 mm of additional SLR by 2100[17].

Apart from SMB, changes in the discharge of ice from iceberg calving and melt from the fronts of marine-terminating outlet glaciers have the potential to increase the rate at which the GrIS contributes to future SLR, and many of these processes are starting to be included in state-of-the-art GrIS models[18]. Calving and frontal melt has already led to ice-front retreat along most of the GrIS and acceleration of marine-terminating glaciers since about 2000[19]. Discharge from the GrIS increased from 1960 to 2005 but stabilized thereafter, although with large interannual fluctuations[1,20]. These recent changes in discharge are thought to be linked in part to fluctuations in the North Atlantic ocean circulation[21,22]. There is evidence that the increase in ice discharge from the 1970s to early 2000s, as measured by changes in iceberg numbers, is also closely related to increasing runoff[20] — from increased melting of ice fronts by upwelling freshwater plumes and the filling and hydrofracturing of crevasses[23], for example.

Increased runoff, percolation of meltwater to the base of the ice sheet and subsequent basal lubrication has also been proposed as a mechanism for general ice flow acceleration in the ablation zone (the Zwally effect)[24], but has since been shown to result in only moderate speed-up at the beginning of the melt season, which can be counteracted by the development of an efficient drainage system[25]. Modelling studies indicate that on decadal to centennial timescales, the Zwally effect has a very limited contribution to global SLR[26,27].

Future SMB and discharge components of the mass budget cannot be separated entirely because of the SMB–elevation feedback and, more importantly, the interaction between the two components as more negative SMB removes ice before it can reach the marine margins[27,28]. However, both of these effects become more important with stronger climate forcing and therefore remain limited for the low-emissions scenario considered here. Modelling studies indicate that the partitioning between mass losses from SMB and ice discharge and their spatial distribution are likely to remain similar to the present day[17,27], although these studies do not account for the full range of uncertainty associated with outlet glacier changes. However, given that recent SMB changes dominate the recent GrIS mass loss[14], the largest source of uncertainty in future SLR is likely to be linked to SMB.

Expected Greenland response

Modelling studies of the GrIS, according to RCP2.6, report a large spread in ice-sheet volume change of 14–78 mm sle by 2100[17,27], with uncertainty arising mainly from differences between climate models. The largest discrepancies between different climate projections and ice-sheet models occur over the fast-flowing outlet glaciers[29]. Recent advances in high-resolution model simulations[30] highlight the importance of bed topography in controlling ice-front retreat for a given amount of ocean warming. However, capturing the dynamics of outlet glaciers remains difficult for several reasons: (1) outlet glacier flux is not always determined at a sufficient resolution due to limited knowledge of the subglacial topography[31] (despite the significant progress made through mass conservation algorithms[32]); (2) the impact of ocean temperature on ice discharge at the margin is poorly constrained; (3) understanding of iceberg calving remains limited[33], yet such mechanisms drive most of the dynamic changes in marine-terminating glaciers[34].

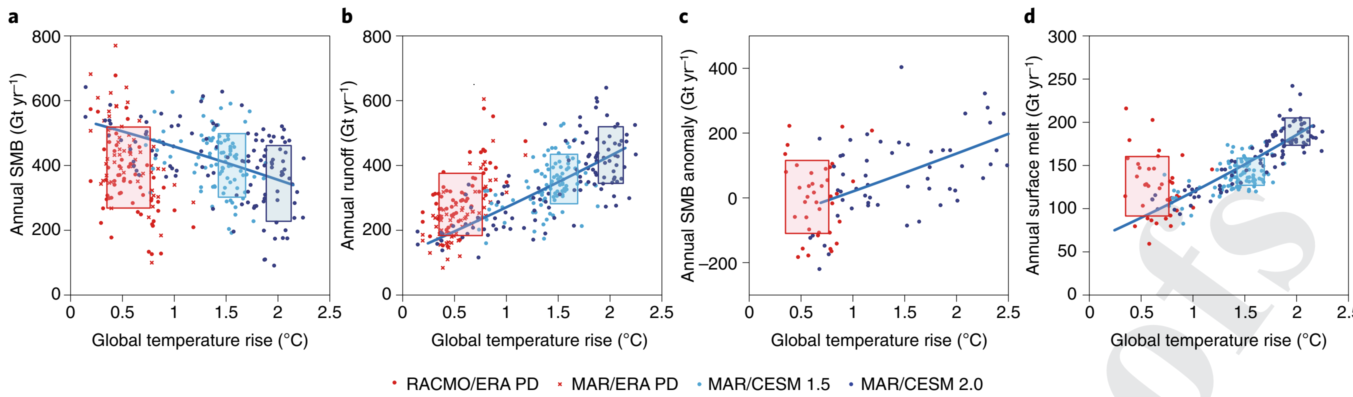

Fig. 1 | Annual mean surface mass fluxes as a function of global mean temperature anomalies. Temperature anomalies are referenced to the pre-industrial era (1850–1920). **a**, GrIS SMB. **b**, GrIS runoff. **c**, Antarctic SMB. **d**, Antarctic surface melt. Red colours indicate model realizations of present-day ice sheets (RACMO2 and MAR forced by ERA reanalysis data). Blue colours indicate model realizations of future ice sheets. In **a** and **b**, MAR is forced with CESM-CAM5 1.5 and 2.0 future scenarios (+1.5 and 2.0 °C). In **c**) RACMO2 is forced with a HadCM3 A1B scenario. In **d**, CESM-CAM5 1.5 and 2.0 future scenarios include surface melt parameterized in terms of near-surface temperature[48]. Trend lines are shown for future (blue) model realizations. Boxes delimit 2 s.d. in temperature and SMB components over the present-day period (red boxes) and the stationary climate over 2061–2100 in the CESM-CAM5 1.5 (light blue boxes) and 2.0 (dark blue boxes) scenarios. None of these simulations include coupling to an ice dynamical model.

Box 2 | Climate commitment and tipping points

For the long-term evolution of the ice sheets, on multicentennial to multimillennial timescales, feedbacks with the atmosphere and ocean increase in importance. When subjected to perturbed climatic forcing over this timescale, the ice sheets manifest large changes in their volume and distribution. These changes typically occur with a significant lag in response to the forcing applied, which leads to the concept of climate commitment: changes that will occur in the long-term future are committed to at a much earlier stage[96]. Because of the long residence time of CO_2 in the atmosphere, climate change in coming decades will most probably last long enough to dictate ice-sheet evolution over centuries and millennia[41,58,81,97]. Furthermore, the ice sheets are subject to threshold behaviour in their stability, as a change in boundary conditions such as climate forcing can cause the current ice-sheet configuration to become unstable. Crossing this tipping point leads the system to equilibrate to a qualitatively different state[98] (by melting completely, for example). The existence of a tipping point implies that ice-sheet changes are potentially irreversible — returning to a pre-industrial climate may not stabilize the ice sheet once the tipping point has been crossed. A key concept here is the timeframe of reversal, because many ice-sheet changes may only be reversible over a full glacial–interglacial cycle with natural rates of changes in climatic variables. For both Greenland and Antarctica tipping points are known to exist for warming levels that could be reached before the end of this century[58,81,99]. The unprecedented rate of increase in GHGs over the Anthropocene leaves the question of irreversible crossing of tipping points unresolved. For example, it is possible that the expected future increase in GHGs will prevent or delay the next ice-sheet inception[100].

On longer timescales (Box 2), a tipping point (when the ice sheet enters a state of irreversible mass loss and complete melting is initiated) exists as part of the coupled ice sheet–atmospheric system. This consists of two interrelated feedback mechanisms: the SMB–elevation feedback, as described above, and the melt–albedo feedback[35–37]. The latter acts on the surface energy balance, by allowing more absorption of solar radiation from a melting and darkening snow surface, or removal of all snow leading to a darker ice surface. This feedback may be enhanced by ice-based biological processes, such as the growth of algae[38]. Thus, the activation of these feedbacks can lead to self-sustained melting of the entire ice sheet, even if the anomalous climatic forcing is removed.

It is clear that if the tipping point is crossed, a complete disappearance of the GrIS would occur on a multimillennial timescale[39–41]. However, further work is urgently needed to diagnose how close the GrIS is to this tipping point. Figure 2 shows results from an ensemble of simulations using one model and varying key parameters related to precipitation changes and melt rates[40]. Simulations were performed with slowly increasing climatic forcing, allowing the ice sheet to maintain a state of quasi-equilibrium. Each simulation in the ensemble reached a tipping point, when the ice sheet could no longer sustain itself. Figure 2a compares this equilibrium threshold with the diagnosed SMB of the GrIS given its present-day distribution, which can roughly be used as a proxy for stability. SMB is spatially inhomogeneous, however, with high accumulation and melt rates in the south, and cold, desert-like conditions in the north. These simulations show that the northwest sector of the ice sheet is particularly sensitive to small changes in SMB, given the relatively low accumulation rates and associated slower flow of ice from inland compared to the south. Thus, in this model, a negative SMB in the northwest sector is a good predictor for the estimated threshold for complete melting of the ice sheet.

The 95% confidence interval for the regional summer temperature threshold leading to GrIS decline ranges from 1.1–2.3 °C above pre-industrial, with a best estimate[40] of 1.8 °C. This level of warming is well within the range of expected regional temperature changes if global warming is limited to 1.5 °C, as CMIP5 models predict that Greenland near-surface air temperatures increase more than the global average and current levels of summer warming already reach this limit. This means that the threshold will probably be exceeded, even for aggressive anthropogenic carbon emissions reductions. However, in some peak-and-decline scenarios of CO_2 levels, full retreat can probably be avoided despite the threshold having been temporally crossed.

The committed SLR after 1,000, 5,000 or 15,000 years — that is, how much the ice sheet will melt for a given climatic perturbation today (assumed constant in time) — increases nonlinearly for

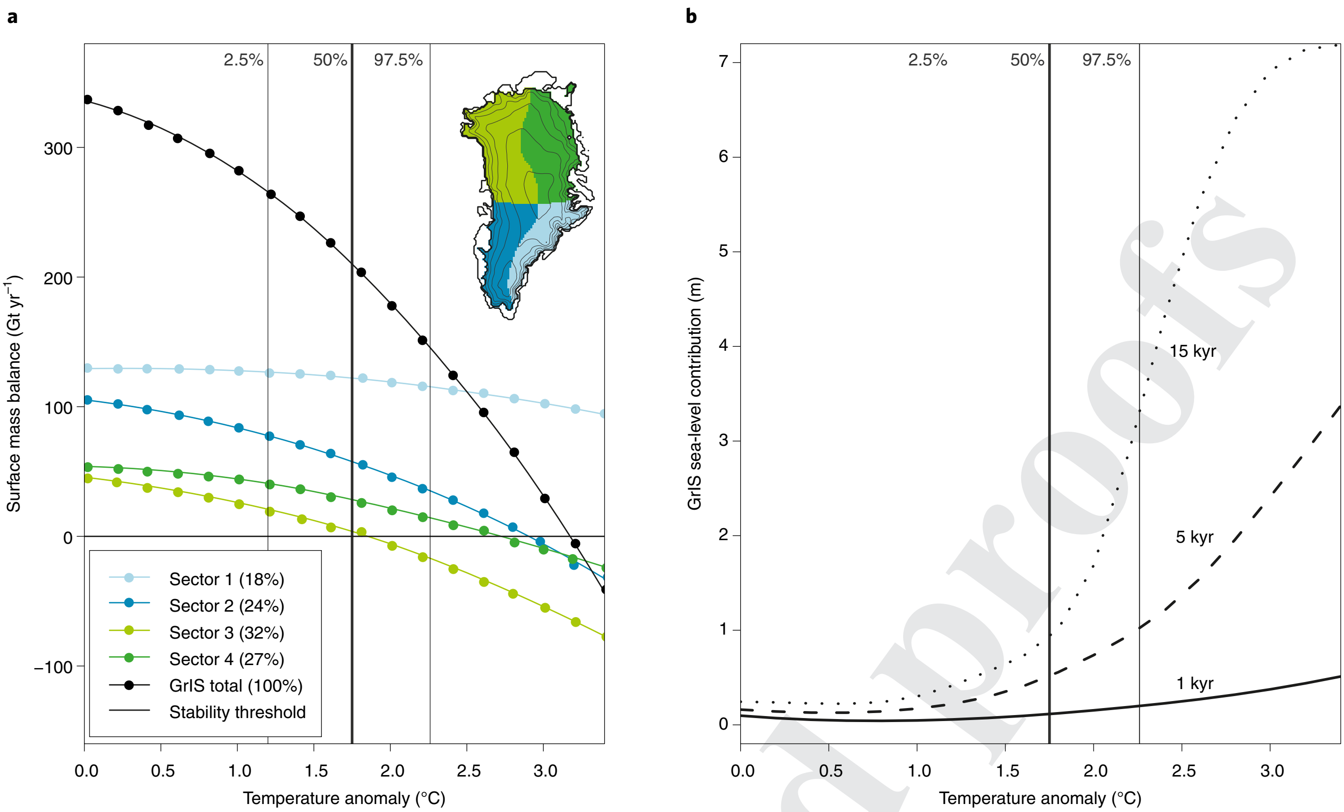

Fig. 2 | GrIS stability as a function of the imposed regional summer temperature anomaly with best-estimate model parameter values. a, GrIS surface mass balance by sector, diagnosed from regional climate model simulations with a fixed, present-day ice-sheet topography. **b**, Expected SLR contribution of the GrIS after 1, 5 and 15 kyr versus constant temperature. The vertical lines in both panels show the probability of crossing the tipping point for melting the ice sheet (2.5%, 50% and 97.5% credible intervals) to 10% of its current volume or less, as estimated by an ensemble of dynamic quasi-equilibrium simulations of the GrIS under a slowly warming climate[40].

higher levels of warming (Fig. 2b). The lag in response implies that such a retreat would be set in motion much sooner, on timescales of the order of decades to centuries (see Box 2). Thus, crossing the limit of 1.5 °C global warming this century may impose a commitment to much larger and possibly irreversible changes in the far future[40,41].

Antarctic forcing and mass balance changes

The AIS has been losing mass since the mid-1990s, contributing 0.15–0.46 mm yr^{-1} sle on average between 1992 and 2017, accelerating to 0.49–0.73 mm yr^{-1} between 2012 and 2017[42]. Observations over the past five years show that mass loss mainly occurs in the Antarctic Peninsula and West Antarctica (0.42–0.65 mm yr^{-1} sle), with no significant contribution from East Antarctica (−0.01–0.16 mm yr^{-1} sle)[42]. The mass loss from the West Antarctic Ice Sheet (WAIS) is primarily caused by the acceleration of outlet glaciers in the Amundsen Sea Embayment (ASE), where the ice discharge of large outlet glaciers such as the Pine Island and Thwaites glaciers increased threefold since the early 1990s[42]. However, this ASE mass loss is not a recent phenomenon, as ocean sediment records indicate that Pine Island Glacier experienced grounding-line retreat since approximately the 1940s[43].

Antarctic SMB is projected to increase under atmospheric warming, governed by increased snowfall due to increased atmospheric saturation water vapour pressure, the availability of more open coastal water and changing cloud properties[44]. Ice cores suggest that on centennial timescales SMB has increased, especially in the Antarctic Peninsula, representing a net reduction in sea level of ~0.04 mm per decade since 1900 CE[45]. According to CMIP5 model means for RCP2.6, increased snowfall mitigates SLR by 19 mm by 2100 and by 22 mm if only those CMIP5 models that best capture CloudSat-observed Antarctic snowfall rates are used[46]. Under atmospheric warming, Antarctic surface melt (estimated at ~0.3 mm yr^{-1} sle[47]) is projected to increase approximately twofold by 2050, independent of the RCP forcing scenario[48]. Recent studies show that meltwater in Antarctica can be displaced laterally in flow networks[49], and sometimes even enters the ocean[50]. However, further research is needed to assess whether these processes can challenge the present view that almost all surface meltwater refreezes in the cold firn[47].

Major ice loss from the AIS stems from an increased discharge of grounded ice into the ocean, with ice shelves (the floating extensions of the grounded ice sheet) playing a crucial role. The buttressing provided by ice shelves can affect inland ice hundreds of kilometres away[51], and hence controls grounding-line retreat and associated ice flow acceleration. Ice shelves are directly affected by oceanic and atmospheric conditions, and any change in these conditions may alter their buttressing effect and impact the glaciers feeding them. For instance, increased sub-shelf melting causes ice shelves to thin, increasing their sensitivity to mechanical weakening and fracturing. This causes changes in ice shelf rheology and reduces buttressing of the inland ice, leading to increased ice discharge[52]. Warming of the atmosphere promotes rainfall and surface melt on the ice shelves and causes hydrofracturing as water present at the ice-sheet surface propagates into crevasses[53,54] or by tensile stresses induced by lake drainage[55]. Anomalously low sea-ice cover and the associated increase in ocean swell has also been identified as an important precursor of Antarctic Peninsula ice shelf collapse[56]. These mechanisms were probably involved in the rapid breakup of Larsen B ice shelf in 2002[55]. Although ice cores show that surface melting in the Antarctic Peninsula is now greater than ever recorded in recent history[57], for low-emissions scenarios, the presence of significant rainfall and surface runoff is unlikely to spread far south of the Antarctic Peninsula by 2100[48,54]. Assessment of future surface melt-induced ice-shelf collapse is therefore highly uncertain for mitigated scenarios, with largely diverging estimates in recent literature. Parts of Larsen C, George VI and Abbot ice shelves may become susceptible to hydrofracturing by 2100 under a mitigated climate scenario[54], but most studies identify significant potential ice-shelf collapse by 2100 under only the unmitigated scenarios[48,58].

Major recent dynamic ice loss in the ASE is associated with high melt rates at the base of ice shelves that result from inflow of relatively warm Circumpolar Deep Water (CDW) in ice shelf cavities[59,60], which led to increased thinning of ice shelves in the area and to reduced buttressing of the grounded ice. Evidence from East Antarctica, as well as along the southern Antarctic Peninsula, also links glacier thinning and grounding-line retreat to CDW reaching the deep grounding lines[61,62].

However, the link between CDW upwelling and global climate change is not yet clearly demonstrated, and decadal variability (such as the El Niño/Southern Oscillation) may dominate ice-shelf mass variability in this sector[63]. This variability may increase as interannual atmospheric variability increases in a warming climate[63]. The CMIP5 ensemble also shows a modest mean warming of Antarctic Shelf Bottom Water (ASBW), the ocean water masses occupying the seafloor on the Antarctic continental shelf that provide the heat for basal melting of Antarctic ice shelves, of 0.25 ± 0.5 °C by 2100 under RCP2.6[64]. Given that present-day biases in ASBW in CMIP5 models are of the same order or larger than this warming, and that the main limitation is the ability of these models to resolve significant features in both bedrock topography and the ocean flow[65], RCP2.6 projections of future sub-ice shelf melt remain poorly constrained[64]. Moreover, the link between increased presence of warm deep water on the continental shelf and higher basal melt rates is not always clear; simulations of strengthened westerly winds near the western Antarctic Peninsula showed an increase in warm deep water on the continental shelf but a coincident decrease in ice-shelf basal melt[66].

Increasing the wind forcing over the Antarctic Circumpolar Current has been shown to have little effect on ice shelf basal melting[67]. Ocean-sea ice projections that include ice-shelf cavities have indicated the possibility that significant amounts of warm deep water could gain access to the Filchner-Ronne ice-shelf cavities in the coming century, increasing melt rates by as much as two orders of magnitude[68,69]. This process was seen with forcing from only one of two CMIP3 models and was more dependent on the model that produced the forcing than on the emissions scenario[69], suggesting that this scenario has a low probability.

Reduction of buttressing of ice shelves via the processes described above may eventually lead to the so-called marine ice sheet instability (MISI; Fig. 3). For the WAIS, where the bedrock lies below sea level and slopes down towards the interior of the ice sheet, MISI may lead to a (partial) collapse of this marine ice sheet. This process, first hypothesized in the 1970s, was recently theoretically confirmed[70] and demonstrated in numerical models[71]. It arises from thinning and eventually flotation of the ice near the grounding line, which moves the latter into deeper water where the ice is thicker. Thicker ice results in increased ice flux, which further thins (and eventually floats) the ice, resulting in further retreat into deeper water (and thicker ice) and so on. The possibility that some glaciers, such as Pine Island Glacier and Thwaites Glacier, are already undergoing MISI has been suggested by numerical simulations using state-of-the-art ice-sheet models[72,73]. The past retreat (up to 2010) of Pine

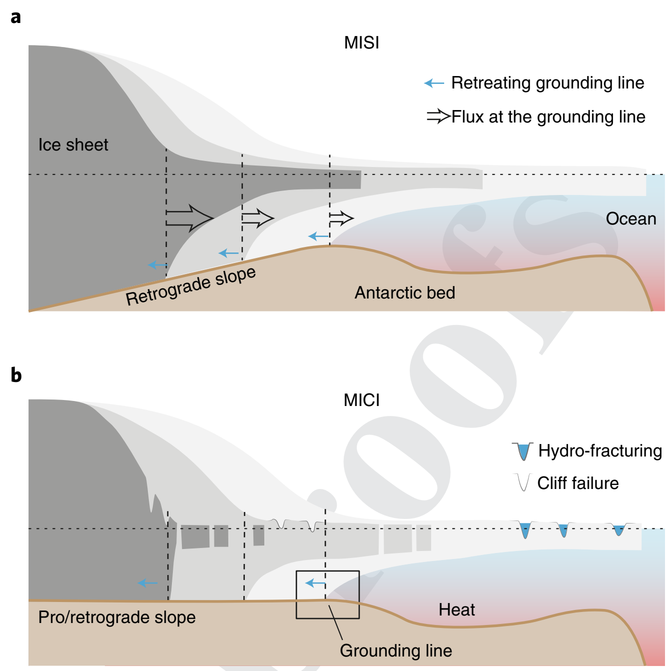

Fig. 3 | MISI and MICI as main drivers for potential (partial) collapse of the AIS. a, MISI can lead to unstable retreat of grounding lines resting on retrograde bed slopes, a very common situation in Antarctica. MISI stems from a positive feedback loop between the increased flux and ice thickness at the grounding line after the latter starts to retreat. **b**, MICI is the result of the collapse of exposed ice cliffs (after the ice shelf collapses due to hydrofracturing) under their own weight. MISI applies for a retrograde slope bed, whereas MICI can also apply for prograde slopes. Both MISI and MICI are thus superimposed for retrograde slopes[58,87]. The red shading qualifies the heat forcing exerted by the ocean on the basal surface of the ice shelf.

Island Glacier has been attributed to MISI[72,74] triggered by oceanic forcing, although its recent slowdown may be due to a combination of abated forcing[75] and a concomitant increase in glacier buttressing. Thwaites Glacier is currently in a less-buttressed state, and several simulations using state-of-the-art ice sheet models indicate continued mass loss and possibly MISI even under present climatic conditions[73,76,77].

Evidence from the observed Larsen B collapse, and rapid front retreat of Jakobshavn Isbrae in Greenland, suggests that hydrofracturing could lead to the rapid collapse of ice shelves and potentially produce high ice cliffs with vertical exposure above 90 m rendering the cliffs mechanically unsustainable, possibly resulting in what has been termed marine ice cliff instability (MICI; Fig. 3)[78]. This effect, if triggered by a rapid disintegration of ice shelves due to hydrofracturing could lead to an acceleration of ice discharge in Antarctica, but is unlikely in a low-emissions scenario[58,79]. However, this process has not yet been observed in Antarctica, and may be prevented or delayed by refreezing of meltwater in firn[54] or if efficient surface drainage exists[50].

Expected Antarctic response

A major limiting factor in projecting the future AIS response is how global warming relates to ocean dynamics that bring CDW onto and across the continental shelf, potentially increasing sub-shelf melt. Because of this uncertainty, several studies apply linear extrapolations of present-day observed melt rates, while focusing on unmitigated scenarios (RCP8.5). Mass loss according to mitigated scenarios are essentially limited to dynamic losses in the ASE

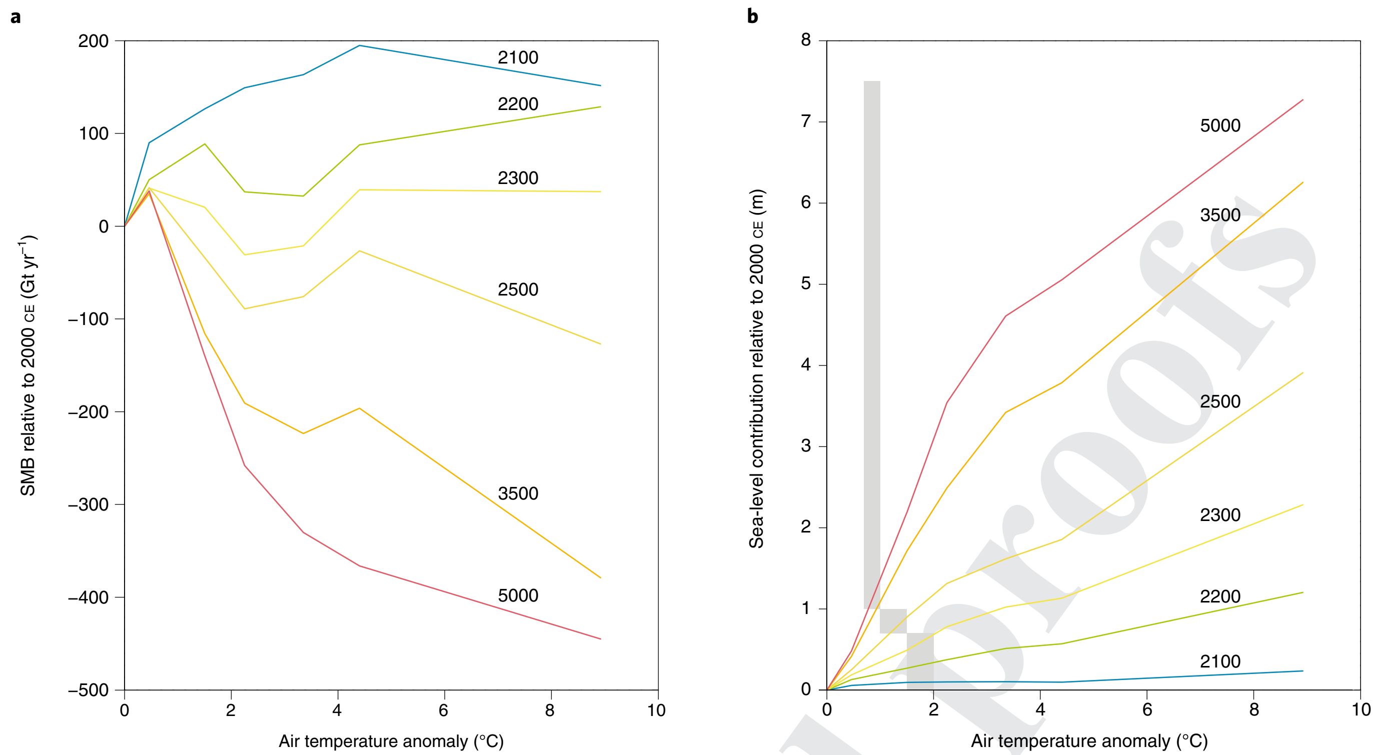

Fig. 4 | AIS stability as a function of the imposed regional annual mean temperature anomaly. a,**b**, Changes in SMB (**a**) and SLR contribution (**b**) for the AIS relative to 2000 CE as simulated under spatially uniform temperature increases that follow RCP trajectories to 2300 CE and then stabilize[81]. Coloured lines denote different years (CE); data are averages of high and low scenarios, denoting two different grounding-line parameterizations. Grey shading shows the approximate equivalent global mean temperature anomaly for an Antarctic mean temperature anomaly of 1.5–2.0 °C, accounting for polar amplification.

of up to 0.05 m sle by 2100. This is not very different from a linear extrapolation of the present-day mass losses[76,77,80] and in contrast with the observed acceleration of mass loss over the last decade[42]. For the whole AIS, mass loss between 0.01 and 0.1 m by 2100 is projected according to RCP2.6[81], which is not dissimilar (−0.11 to 0.15 m by 2100) from model simulations based on Pliocene sea level (5–15 m higher than today) tuning[58], associated with a different melt parameterization at the grounding line (Fig. 4). As the value of sea level at the Pliocene is still debated[82], tuning the model with a higher Pliocene sea-level target (10–20 m) increases the model sensitivity, with an upper bound of 0.22 m by 2100 according to the same scenario[58].

Because ocean heat supply is the crucial forcing for sub-shelf melting, oceanic forcing has the potential to modulate the retreat rate. Significant regional differences exist between Antarctic drainage basins in terms of oceanic heat fluxes and the topographic configuration of the ice-sheet bed[83]. Consequently, the ice-sheet response to ocean thermal forcing, even for small temperature anomalies, may be governed by bed geometry as much as by environmental conditions[83,84]. Observations and modelling show that surface melt occurs on some smaller ice shelves[44,47,48], but also that this may not be a recent phenomenon[49]. According to global and regional atmospheric modelling, under intermediate emissions scenarios, Antarctic ice shelf surface melt will probably increase gradually and linearly[48]. It should be noted, however, that while surface melt is not the major present-day forcing component, the high-end SLR contributions reached for RCP8.5 scenarios[58] stem from increased surface melting rather than oceanic forcing.

The projected long-term (500-year) SLR contribution of the AIS for warming levels associated with the RCP2.6 scenario is limited to well below one metre, although with a probability distribution that is not Gaussian and presents a long tail towards high values due to potential MICI[58], with the caveats listed above. Importantly, substantial future retreat in some basins (such as TG) cannot be ruled out and grounding-line retreat may continue even with no additional forcing[73,77,85,86]. The long-term SLR contribution of the AIS therefore crucially depends on the behaviour of individual ice shelves and outlet glacier systems and whether they enter into MISI for the given level of warming. Under sustained warming, a key threshold for survival of Antarctic ice shelves, and thus the stability of the ice sheet, seems to lie between 1.5 and 2 °C mean annual air temperature above present (Figs. 1d and 4)[81]. The activation of several larger systems, such as the Ross and Ronne-Filchner drainage basins, and onset of much larger SLR contributions are estimated[81] to be triggered by global warming between 2 and 2.7 °C. This implies that substantial Antarctic ice loss can be prevented only by limiting GHG emissions to RCP2.6 levels or lower[58,81]. Crossing these thresholds implies commitment to large ice-sheet changes and SLR that may take thousands of years to be fully realized and be irreversible on longer timescales.

Need for improvement

Considerable progress has been made over the past decade with respect to understanding processes at the interface between ice sheets, atmosphere and ocean, but significant uncertainties in both forcing and the response of the ice sheets remain[18,87]. For the AIS, for instance, the majority of present-day mass loss (essentially the ASE) is driven by changes in ocean circulation. Our ability to simulate those changes into the future is limited, leading to large remaining uncertainties for any projection of AIS mass balance. Similar challenges remain in modelling changes in regional atmospheric circulation that affect GrIS mass loss. Therefore, it is not clear to what degree global warming must be limited to reduce future ice-sheet-related SLR contributions. Other challenges in climate and ice-sheet modelling concern model resolution, initialization and coupling. Model resolution is a key issue, as climate and ocean models tend to be too diffusive. Higher model resolutions increase eddy activity and advective heat transfer more readily than at lower resolution[88]. Recent work[89] uses high-resolution, non-hydrostatic atmospheric and detailed SMB models to better represent surface physical processes at scales finer than 10 km. Likewise, to resolve grounding-line dynamics, ice-sheet models need high spatial resolution across the grounding line[90] and new numerical techniques, such as adaptive meshing, have been developed in recent years to achieve this[91]. Model initialization relies on two distinct, but often combined approaches (spin-up versus data assimilation; Box 1), the latter technique improving for centennial projections with the increasing access to high-resolution satellite products.

Further developments include the need for two-way coupling of ice sheets with coupled atmosphere–ocean models, meaning that climate models not only force ice-sheet models but that the reverse is also true. This calls for closer collaborations across disciplines, which is exemplified by ice sheet model intercomparisons (such as ISMIP6[92]) within CMIP6. A similar intercomparison exercise for SMB and ocean models is urgently needed, given remaining uncertainties in absolute SMB values and sub-shelf melting, with the former particularly relevant for Greenland[7,14,93] and the latter for Antarctica. For instance, if a possible link is found between global warming and the current circulation changes observed in summer over Greenland, this could significantly amplify the melt acceleration projected for the future via a newly recognized positive feedback. To achieve this, it will therefore be critical to further understand and improve the representation of changes in the atmosphere and ocean global circulation in global and regional climate model simulations.

Data availability

Data from CESM-CAM5 is available at: https://www.earthsystemgrid.org/dataset/ucar.cgd.ccsm4.lowwarming.html

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Author contributions
F.P. and C.R. coordinated the study. F.P., C.R. and E.H. led the writing, and all authors contributed to the writing and discussion of ideas. J.T.M.L., P.K.M. and L.D.T. contributed the data that led to Fig. 1. L.F. designed Fig. 3. N.R.G. provided the data that led to Fig. 4. Q21

Competing interests
The authors declare no competing interests.

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