The uncertain future of the Antarctic ice sheet

Frank Pattyn [1*] and Mathieu Morlighem [2]

[1]Laboratoire de Glaciologie, Université libre de Bruxelles, Brussels, Belgium
[2]Department of Earth System Science, University of California, Irvine, CA, USA

*To whom correspondence should be addressed; E-mail: fpattyn@ulb.ac.be

The Antarctic ice sheet is losing mass at an accelerating pace and ice loss will likely continue over the coming decades and centuries. Some regions of the ice sheet may reach a tipping point, potentially leading to rates of sea level rise at least an order of magnitude larger than those currently observed, due to strong positive feedbacks in the ice-climate system. How fast and how much Antarctica will contribute to sea level remains uncertain, but multimeter sea level rise is likely for a mean global temperature increase of around two degrees above pre-industrial levels on multicentennial time scales, or sooner for unmitigated scenarios.

Major uncertainties in predicting and projecting future sea-level rise are due to the contribution of the Antarctic ice sheet (*1*). These uncertainties essentially stem from the fact that some regions of the ice sheet may reach tipping points, defined as (regionally) irreversible mass loss, with a warming climate. The exact timing of when these tipping points are reached remains difficult to assess, allowing for a large divergence in timing of onset and mass loss in model projections. The instability mechanisms responsible for these tipping points are closely related to the shape of the bed under the ice sheet (Fig. 1). The West Antarctic ice sheet, which has the potential to raise sea level by 5.3 m (*2*), has its current base grounded well below sea level and the bed deepens from the periphery of the ice sheet towards the interior (a so-called retrograde bed slope). Marine basins are also present in certain areas of the East Antarctic ice sheet (Fig. 1), which has a far greater sea level potential of 52.2 m (*2*). Marine ice sheets are in direct contact with the ocean under floating ice shelves around the coast, and changes in ocean circulation or heat content may lead to rapid ice loss on timescales of decades to centuries. The uncertainty in the timing and extent of potential tipping points also stems from our poor knowledge of both drivers of change and mechanisms that operate in the dynamics of marine ice sheets. Despite these shortcomings, multi-model comparisons like ISMIP6 allow for a more standardised approach that enable outliers to be more clearly identified. Hence, uncertainties in future projections have since been reduced and more robust projections of sea level contributions from the Antarctic ice sheet are to be expected.

Observations and drivers of dynamical mass change

Recent satellite observations indicate that the contribution of the Antarctic ice sheet to sea-level rise has significantly increased in recent years (*3*). Antarctica has been contributing 0.15–0.46 mm yr^{-1} to sea level on average between 1992 and 2017, accelerating to 0.49–0.73 mm yr^{-1} between 2012 and 2017 (*4*). Most ice loss is concentrated in West Antarctica, where the thinning of floating ice shelves is causing glacier flow to accelerate and grounding lines (the contact between the grounded ice sheet and the ice shelf floating on the ocean) to retreat.

The acceleration and thinning of Pine Island Glacier, Thwaites Glacier, and nearby glaciers draining into the Amundsen Sea (Fig. 2), which dominate the mass loss from the West Antarctic Ice Sheet (WAIS), results from ice-shelf thinning and shrinkage, and associated grounding line retreat. This is thought to be a response to a wind-driven increase in the circulation of warm Circumpolar Deep Water (CDW) onto the continental shelf reaching ice shelf cavities and grounding lines (*5*). The strengthening of the regional westerly winds that have forced warmer waters to the grounding zones are attributed primarily to remote changes occurring in the tropics (*6*). However, changes in larger-scale circulation owing to the recent stratospheric cooling due to ozone depletion and increased concentration of greenhouse gases have also been identified as potential drivers (*7*). Thwaites Glacier is today undergoing the largest changes of any ice-ocean system in Antarctica (*8*). This ongoing mass loss will be modulated but likely not reversed by variability in the ocean (*9*).

The East Antarctic Ice Sheet (EAIS) is closer to a balanced state, but this remains poorly constrained in terms of surface mass balance (essentially precipitation-evaporation) and glacial isostatic adjustment (GIA) in response to volume change stemming from the last glacial-interglacial period. Recent studies reveal that some ice shelves in East Antarctica, once thought to be stable, are also exposed to ocean heat and are experiencing high rates of basal melt (*10*), hence the discharge of EAIS may increase if the atmospheric and oceanic conditions change.

Antarctic surface mass balance derived from reconstructions of ice core records show large but opposing trends across West Antarctica, especially for recent decades, while precipitation changes are less pronounced in East Antarctica (*11*). A key attribute of precipitation events is the penetration of warm, moist air masses over the ice sheet, which may dominate the annual total precipitation, and make such events primarily responsible for most interannual variations in precipitation (*12*).

Dynamics of the marine ice sheet

The mass balance of the Antarctic ice sheet, and therefore its contribution to sea level, is determined by the balance between mass gain and mass loss. The ice sheet gains mass from snowfall on its surface and loses mass primarily by ocean-induced melting beneath its floating ice shelves along the coast, and by calving icebergs that drift away and melt in the ocean. While the surface mass balance has been relatively stable over the past decades, ice flow in several sectors of the ice sheet has accelerated, thereby increasing ice discharge. The dominant process triggering these large, rapid changes is the loss of ice-shelf buttressing. This is initiated by changes in ocean circulation and to a lesser extent atmospheric drivers that control summer surface-melt rates (*13, 14*). In particular, the warmer waters of the CDW move toward the ice fronts and ice-shelf grounding zones along troughs in the bathymetry, causing increased melting at the ice-ocean interface. This process thins the ice shelves, reducing drag along their sides and at local pinning points on sea-floor highs, which in turn reduces the buttressing i.e., the resistive stress that the ice shelves exert on the grounded ice (*8*). Thinning ice shelves lead to faster grounded-ice flow, which in turn leads to further thinning, causing previously grounded ice to float as the grounding zone retreats farther inland. This process can be particularly fast and unstable along retrograde slopes (i.e., the bed deepens inland), as more ice crossing the grounding zone and a smaller accumulation area (*15, 16*) creates a positive feedback process known as the marine ice sheet instability (MISI; Fig. 3). The process may halt when the bedrock rises upward, i.e., when a prograde bed slope or pronounced ridge at the bed is encountered, or when ice shelves exert enough buttressing to stop further grounding-line retreat.

The retreat up to 2010 of Pine Island Glacier has been attributed to enhanced ocean-induced melt, although its recent slowdown may be due to a combination of reduced forcing and a concomitant increase in glacier buttressing (*17*). It is possible that some glaciers, such as Pine Island Glacier and Thwaites Glacier, may already be undergoing MISI (*9*). Thwaites Glacier is currently in a less-buttressed state as its ice shelf is mostly unconfined, and several simulations using state-of-the-art ice sheet models indicate continued mass loss and possibly MISI or MISI-like behaviour even under present climatic conditions (*18–20*).

More recently, the hypothesis of Marine Ice Cliff Instability has emerged (MICI) (*14, 21*), postulating that ice cliffs become unstable and collapse if higher than $\sim$90 m above sea level, facilitating the rapid retreat of ice sheets. This process may have been significant in Antarctica during past warms periods (*14*) by enhancing MISI (Fig. 3). During Pliocene warm periods, sea level was 10 to 20 m higher than present (*22*), requiring extensive retreat or collapse of the Greenland, West Antarctic and marine-based sectors of the East Antarctic ice sheets. The MICI mechanism enables to increase the model sensitivity to reach such high sea-level stands during that period (*14*). However, contrary to the MISI hypothesis, MICI is not supported by a formal linear stability analysis (*16*), which hampers an adequate representation in marine ice sheet models. Furthermore, MICI has not been observed at such a scale in Antarctica and so it remains unclear how rapid an ice cliff would retreat as a function of its height (*23*). So far, models including MICI parameterized the rate of retreat based on observed retreat rate of Jakobshavn Isbræ in West Greenland, which reached 3 km yr^{-1} when its ice shelf collapsed in the early 2000s.

Cliff instability requires an a priori collapse of ice shelves, and is favoured by, among others, hydro-fracturing through the increase of water pressure in surface crevasses, which widens and deepens them (*21, 24, 25*). Contrary to MISI, MICI could also occur on prograde bed slopes. Evidence from the Larsen B collapse, and rapid front retreat of Jakobshavn Isbræ, suggest that hydrofracturing could lead to the rapid collapse of ice shelves and potentially produce high, mechanically unsustainable, ice cliffs (*21, 24*). However, its current impact is limited as only few Antarctic ice shelves have collapsed by now. Moreover, recent work shows that the critical cliff height increases with timescale (the longer the timescale, the taller the cliff needs to be before collapse is possible), and therefore, ice shelf buttressing must be removed on timescales of less than one day to produce rapid brittle fracturing of a subaerial ice cliff at heights attainable in ice sheets (*23*). Compelling evidence from the Ross Sea from observations show that there has been no immediate grounding line retreat after cliff collapse in the past (*26*). More research into the dynamics of ice cliffs is needed and the existence of MICI remains today controversial.

Projecting the future of the Antarctic ice sheet

A major factor that limits reliable projections of the future Antarctic ice sheet 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 or simple parameterizations of ice-ocean melting rates, mostly focusing on unmitigated climate scenarios, such as RCP8.5. Numerous large-scale modelling studies conducted in the last decade have simulated future collapse of WAIS under various climate-warming scenarios (*13, 14, 27–30*). These studies find that future grounding-zone retreat into the central WAIS region is expected on timescales of a few centuries to a millennium, contributing several meters to global mean sea level rise. However, while the time of onset of collapse is quite different across models and scenarios, all models produce WAIS collapse under unmitigated emission scenarios on multi-centennial timescales.

Whole Antarctic simulations for unmitigated emission scenarios (RCP8.5) show a large scatter on centennial and multi-centennial timescales (Fig. 4). However, the introduction of MICI in one ice-sheet model (*14*) results in future sea-level rise estimates of almost one order of magnitude larger compared to other studies (Fig. 4). While projected contributions of the Antarctic ice sheets to sea-level rise by the end of this century for recent studies hover between 0 and 0.45 m (5%–95% probability range), the MICI model occupies a range of 0.2–1.7 m (Fig. 4). The discrepancy is even more pronounced for 2300, where the MICI results and other model estimates no longer agree within uncertainty bounds. Given the uncertainty range on Pliocene sea-level stands, MICI is not necessarily required to lead to rapid multi-meter sea level rise (*31*) and other mechanisms related to basal conditions may well be able to accelerate mass loss on shorter timescales (*30, 32*).

Not all feedbacks in marine ice sheets enhance ice loss and collapse. Several mechanisms may slow down rapid ice retreat. For instance, as glaciers thin, the pressure that they exert on the Earth crust decreases and so the bed rises in response to the reduction in ice mass. The lithosphere is a viscoelastic material and the rate of uplift has two distinct response times: the elastic response is instantaneous but limited in magnitude, while the viscous response is slow but larger in magnitude. A low-viscosity asthenosphere and a thin lithosphere (known as a weak Earth structure) as observed under WAIS will produce a faster and more localised viscoelastic response of the solid Earth on decadal rather than millennial timescales (*33*). When the bedrock rises, the grounding line retreat may slow down as the height above hydrostatic equilibrium increases inland. Simulations that account for this negative feedback show bedrock uplift delays the collapse of WAIS, leading to slower mass loss (*34*) compared to models that keep a fixed bedrock geometry. While this mechanism has a strong impact on model simulations on multicentennial to millennial time scales, it is not yet clear whether it is significant on the scale of decades.

Sea-level commitment and tipping points

On multicentennial to multimillennial timescales, feedbacks with the atmosphere and ocean increase in importance. When subjected to perturbed climatic forcing over these timescales, 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 sea-level commitment, i.e., ice mass losses that will occur in the long-term future are committed to that loss at a much earlier stage. Ice sheets are subject to threshold behaviours in their stability, as a change in boundary conditions such as climate forcing can cause the current ice-sheet configuration to become unstable through, for instance, MISI. Crossing these tipping points leads the system to equilibrate to a qualitatively different state (a complete collapse of WAIS, for example). The existence of a tipping point implies that ice-sheet changes are potentially irreversible. In other words, returning to a pre-industrial climate may not necessarily stabilize the ice sheet once the tipping point has been crossed. Reversibility, however, may be possible over large climate cycles, such as a glacial-interglacial cycle.

The projected long-term sea level rise contribution of the Antarctic ice sheet for warming levels associated with the high-mitigation RCP2.6 scenario is limited to well below one metre, although with a probability distribution that is not Gaussian, but skewed with a long tail towards high values due to potential MICI (*1*). However, substantial future retreat in some basins (such as Thwaites Glacier) cannot be ruled out, as grounding-line retreat may continue even with no additional forcing (*18–20, 32*). The long-term sea level rise contribution of the Antarctic ice sheet therefore crucially depends on the behaviour of individual ice shelves and outlet glacier systems and whether they enter MISI for a given level of warming. Under sustained warming, a threshold for the 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 (*28*). Crossing these thresholds implies commitment to large ice-sheet changes and sea level rise that may take thousands of years to be fully realized and be irreversible on longer timescales (*1*).

Understanding key physical processes

Considerable progress has been made over the past decade with respect to understanding fundamental processes at the interface between ice sheets, atmosphere and ocean and mechanisms of ice sheet instability. However, along with missing knowledge on the drivers of change, some key physical processes inherent to the dynamics of retreating marine ice sheets are still poorly understood. These processes include (i) ice-ocean interface processes responsible for sub-shelf melt, (ii) calving and (hydro)fracture processes, (iii) ice-sheet basal sliding and subglacial sediment deformation, and (iv) GIA. This missing knowledge reduces our capability to accurately predict the timing and magnitude of the onset of enhanced mass loss or define potential tipping points of the Antarctic ice sheet.

As discussed above, increased sub-shelf melting (i) has triggered the observed acceleration of large Antarctic outlet glaciers in the Amundsen Sea sector during the last decade (*3, 4, 8*), and it is therefore critical that numerical ice sheet models represent the processes governing sub-shelf melt accurately. Sub-shelf melting is either parameterized or computed through the coupling with an ocean model. Parameterizations typically relate sub-shelf melting to ocean temperature and/or ice-shelf depth, either in a linear or a quadratic fashion, which leads to higher melting close to the grounding line (*35*). Other parameterizations relate sub-shelf melting to the distance to the grounding line, the ice-shelf and cavity depths, or more recently by using melt rates from a plume model that are extended spatially using physically motivated scalings depending upon local slope and ice draft (*35*). More accurate representations of sub-shelf melting can be achieved by the coupling to an ocean model, which should lead to significant improvements compared to simple parameterizations, since it accounts for the transfer of heat, freshwater and momentum between the two bodies.

Iceberg calving (ii) is responsible for the other part of the ice mass loss at the margins of the Antarctic ice sheet. Calving occurs when ice chunks break off from the edge of floating ice shelves in Antarctica. The rate at which icebergs detach from the ice shelf, or calving rate, determines the dynamics of the ice front. When the ice front is stationery, the calving rate is equal to the flow velocity of the ice. The calving rate therefore modulates buttressing induced by ice shelves and hence indirectly controls upstream grounded ice speed and subsequent sea level rise contribution. The large amount of ice lost through calving is common for Antarctica, but its representation and quantification in models is hampered by the difficult access to field sites, a high variability in time and space, and its inherent discontinuous nature, as opposed to the continuum approach used in most models. Until recently, calving rates were essentially either assumed to be equal to ice velocity (i.e. by keeping ice front fixed in space) or based on empirical relationships that are not well constrained by observations. Recent studies apply continuum damage mechanics to simulate crevasse formation. This approach represents initial ice microfractures and their vertical development as crevasses, which in turn weakens the ice through damage and decreases ice viscosity, and which can be advected with the ice flow (*36*). Hydrofracturing, based on the surface meltwater widening and deepening crevasses is also ubiquitously parameterized in ice sheet models, and forms the precursor for MICI (*21, 24*). Calving remains one of the grand challenges of ice sheet modelling and no general calving law exists yet, which profoundly limits our ability to model catastrophic calving events.

Basal conditions (iii) and glacial isostatic adjustment (iv) both have an impact on how ice sheets respond to forcing. While the physics of GIA is well understood, the upper-mantle viscosity under the Antarctic ice sheet is poorly constrained. Similarly, the mechanics of basal friction and how it varies spatially remains largely unknown. Models typically rely on simple friction laws that depend on the basal velocity linearly or non-linearly (*37*), which is generally a good approximation for a hard bedrock. Many Antarctic ice streams are however known to be lying on soft beds that have a layer of deformable till. Recent studies and laboratory experiments suggest that the rheology of the till is plastic at large strain, and new parameterizations are being developed to account for both soft and hard beds (*37*). The development and validation of these new friction laws is critical to further improve the predictive skills of numerical models.

Challenges to reduce uncertainties

Besides understanding of key physical processes, their representation in ice-sheet models are also crucial. One way to assess the accuracy in the representation of physical processes in current ice sheet models is to organize large, international intercomparison projects. For example, the Marine Ice Sheet Model Intercomparison Project for planview models (MISMIP3d) greatly improved the representation grounding-line migration by conforming models to known analytical solutions (*38*). These numerical experiments demonstrated that in order to resolve grounding-line migration in marine ice-sheet models, a sufficiently high spatial resolution needs to be adopted, since membrane stresses need to be resolved across the grounding line to guarantee mechanical coupling, unless parameterizations are used (*14*) based on analytical solutions (*16*). Therefore, a series of ice-sheet models have implemented sub-element parameterizations or a spatial grid refinement, which also favours accurate data assimilation (*27*). In transient simulations the adaptive mesh approach enables the finest grid to follow the grounding-line migration (*27*). These higher spatial resolutions on the order of hundreds of meters in the vicinity of grounding lines also pose new challenges about data management for modelling purposes and demand precise bathymetry to resolve the grounding zone (*2*). Nevertheless, recent theoretical developments with respect to grounding-line stability in response to buttressing (*39*), basal drag (*40*) and external forcing (*41*) demonstrate that further efforts are required in the verification and validation of numerical ice sheet models.

Intercomparisons are also essential for improving coupled ocean/sub-ice-shelf cavity/ice-sheet models within a global system context (*42*). To better understand the influence of model initialization, an initial state intercomparison exercise (initMIP) has been developed (*43*). initMIP is the first set of experiments of the Ice Sheet Model Intercomparison Project for CMIP6 (ISMIP6), which is the primary Coupled Model Intercomparison Project Phase 6 (CMIP6) activity focusing on the Greenland and Antarctic ice sheets (*42*).

Besides multi-model ensembles, such as ISMIP6, uncertainty quantification (UQ) within the model parameter space is a powerful tool to characterize and investigate uncertainty in projections (*29, 30*), and to improve projections of future sea level rise. One of the advantages of UQ, is that it can quantify the uncertainty in the projections associated to different input parameters, related to either external forcing or to physical properties of the ice sheet (e.g., initial conditions, coefficients in parameterizations). It therefore makes it possible to show where progress should be made to reduce the uncertainty in projections of sea level rise most efficiently.

Model initialization remains another important factor, which relies on two distinct, but often combined approaches: spin-up versus data assimilation. The first approach spins up 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 present-day ice sheet geometry and flow speed, which may introduce significant biases on short term (i.e. decadal to centennial) projections. The alternative is the assimilation of data, such as satellite derived surface flow speeds, thinning/thickening rates, etc. 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 sea level rise (*43*). While data assimilation techniques cannot ensure consistent internal properties of the ice sheet, they are improving for centennial projections with the increasing access to high-resolution satellite products, which even allow for characterizing the subglacial conditions to a far better degree (*44*). They also enable to improve ice thickness and bedrock data sets at a high resolution for the Antarctic ice sheet (*2*). One of the challenges for the coming years is that the volume of data available is increasing exponentially, while ice sheet models are not equipped to ingest large amount of data from different sensors at different resolutions and acquired at different times. Some progress has been made by relying on tools such as automatic differentiation, but these methods have not yet been applied to large scale systems such as the entire Antarctic ice sheet.

Eventually, the full coupling between ice, ocean and atmosphere must be considered, which is currently the subject of ongoing research, but remains limited to decadal or multi-decadal timescales due to the high computational cost of coupled models. Full ice-ocean coupling on the Thwaites drainage basin revealed a continued mass loss over the coming decades at a sustained rate and show that uncoupled simulations significantly overestimate the rate of grounding-line retreat compared to the coupled model (*20*). Whole Antarctic semi-coupled simulations, on the other hand, show that meltwater from Antarctica will trap warm water below the sea surface, creating a positive feedback that increases Antarctic ice loss (*32*).

The increase in computational efficiency enabling high spatial resolution modelling, the availability of high-resolution datasets of bed topography, high-resolution satellite-based ice surface velocity and changes in ice velocity, longer time series on ice sheet changes, and the improved initialisation of ice sheet models are now allowing the ice sheet modelling community to produce increasingly more robust projections on the future behaviour of the Antarctic ice sheet. Closing knowledge gaps in drivers, forcing and processes and an improved understanding of feedbacks between the different systems will be necessary to more accurately comprehend when and how future tipping points of the ice sheet are reached, as they have a profound impact on global sea level rise around the planet.

References

1. F. Pattyn, *et al.*, *Nature Climate Change* **8**, 1053 (2018).

2. M. Morlighem, *et al.*, *Nature Geoscience* (2019).

3. E. Rignot, *et al.*, *Proceedings of the National Academy of Sciences* **116**, 1095 (2019).

4. A. Shepherd, *et al.*, *Nature* **558**, 219 (2018).

5. S. Schmidtko, K. J. Heywood, A. F. Thompson, S. Aoki, *Science* **346**, 1227 (2014).

6. P. Dutrieux, *et al.*, *Science (80-.).* **343** (2014).

7. D. W. J. Thompson, *et al.*, *Nature Geoscience* **4**, 741 (2011).

8. F. S. Paolo, H. A. Fricker, L. Padman, *Science (80-.).* **348**, 327 (2015).

9. K. Christianson, *et al.*, *Geophys. Res. Lett.* **43**, 10,817 (2016).

10. S. R. Rintoul, *et al.*, *Science Advances* **2** (2016).

11. B. Medley, E. R. Thomas, *Nature Climate Change* **9**, 34 (2019).

12. J. Turner, *et al.*, *Geophysical Research Letters* **46**, 3502 (2019).

13. J. Feldmann, A. Levermann, *Proceedings of the National Academy of Sciences* **112**, 14191 (2015).

14. R. M. DeConto, D. Pollard, *Nature* **531**, 591 (2016).

15. J. Weertman, *J. Glaciol.* **13**, 3 (1974).

16. C. Schoof, *Journal of Geophysical Research: Earth Surface* **112**, n/a (2007). F03S28.

17. L. Favier, *et al.*, *Nature Climate Change* **4**, 117 (2014).

18. I. Joughin, B. E. Smith, B. Medley, *Science* **344**, 735 (2014).

19. I. J. Nias, S. L. Cornford, A. J. Payne, *J. Glaciol.* **62**, 552 (2016).

20. H. Seroussi, *et al.*, *Geophys. Res. Lett.* pp. n/a–n/a (2017).

21. D. Pollard, R. M. DeConto, R. B. Alley, *Earth and Planetary Science Letters* **412**, 112 (2015).

22. G. R. Grant, *et al.*, *Nature* **574**, 237 (2019).

23. F. Clerc, B. M. Minchew, M. D. Behn, *Geophysical Research Letters* **46**, 12108 (2019).

24. J. N. Bassis, C. C. Walker, *Proceedings of the Royal Society A: Mathematical, Physical and Engineering Sciences* **468**, 913 (2012).

25. A. A. Robel, A. F. Banwell, *Geophysical Research Letters* **46**, 12092 (2019).

26. P. J. Bart, M. DeCesare, B. E. Rosenheim, W. Majewski, A. McGlannan, *Scientific Reports* **8**, 12392 (2018).

27. S. L. Cornford, *et al.*, *The Cryosphere* **9**, 1579 (2015).

28. N. R. Golledge, *et al.*, *Nature* **526**, 421 (2015).

29. C. Ritz, *et al.*, *Nature* **528**, 115 (2015).

30. K. Bulthuis, M. Arnst, S. Sun, F. Pattyn, *The Cryosphere* **13**, 1349 (2019).

31. T. L. Edwards, *et al.*, *Nature* **566**, 58 (2019).

32. N. R. Golledge, *et al.*, *Nature* **566**, 65 (2019).

33. V. R. Barletta, *et al.*, *Science* **1339**, 1335 (2018).

34. E. Larour, *et al.*, *Science* **7908** (2019).

35. L. Favier, *et al.*, *Geoscientific Model Development* **12**, 2255 (2019).

36. S. Sun, S. L. Cornford, J. C. Moore, R. Gladstone, L. Zhao, *The Cryosphere* **11**, 2543 (2017).

37. J. Brondex, F. Gillet-Chaulet, O. Gagliardini, *The Cryosphere* **13**, 177 (2019).

38. F. Pattyn, *et al.*, *Journal of Glaciology* **59**, 410 (2013).

39. M. Haseloff, O. V. Sergienko, *Journal of Glaciology* **64**, 417–431 (2018).

40. O. V. Sergienko, D. J. Wingham, *Journal of Glaciology* **65**, 833–849 (2019).

41. G. H. Gudmundsson, F. S. Paolo, S. Adusumilli, H. A. Fricker, *Geophysical Research Letters* **46**, 13903 (2019).

42. S. M. J. Nowicki, *et al.*, *Geoscientific Model Development* **9**, 4521 (2016).

43. H. Seroussi, *et al.*, *The Cryosphere* **13**, 1441 (2019).

44. F. Gillet-Chaulet, *et al.*, *Geophysical Research Letters* **43**, 10,311 (2016).

45. H. D. Pritchard, R. J. Arthern, D. G. Vaughan, L. A. Edwards, *Nature* **461**, 971 (2009).

46. E. Hanna, *et al.*, *Earth-Science Reviews* **201**, 102976 (2020).

47. A. Levermann, *et al.*, *Earth Syst. Dyn.* **5**, 271 (2014).

Acknowledgments

The authors would like to thank K. Bulthuis for drafting Fig. 4, Nick Golledge and two anonymous reviewers, as well as the editor, for their insightful comments.

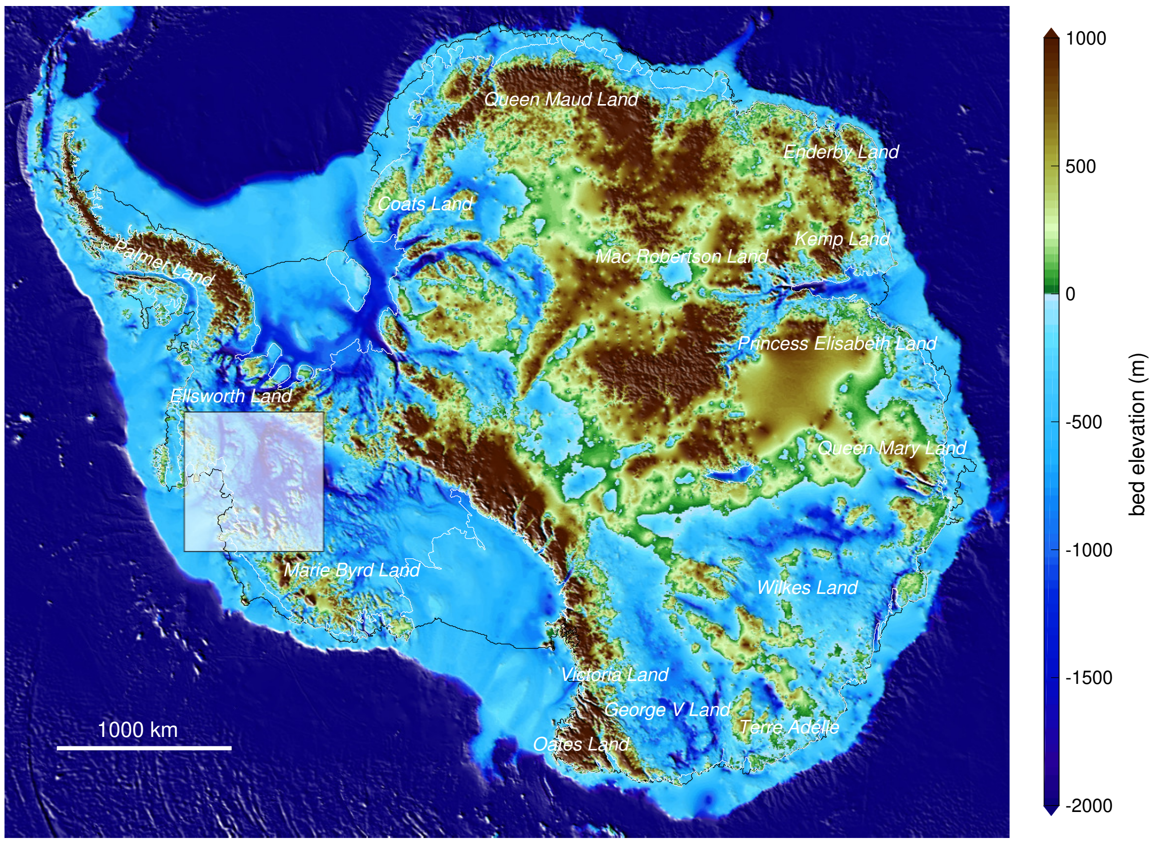

Figure 1: Bed topography (bathymetry) of Antarctica (*2*). Blue areas are marine based (below sea level). The ice-sheet grounding line is plotted in white and ice front in black. The greyed area indicates the Amundsen Sea Embayment, shown in Figure 2.

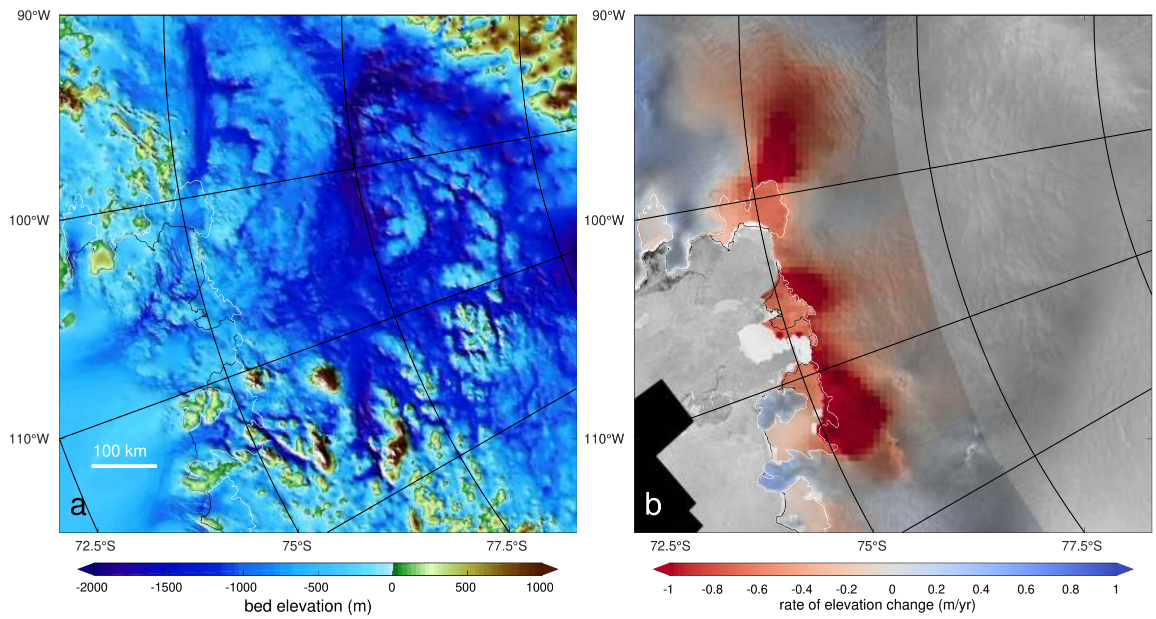

Figure 2: (a) Bed topography (bathymetry) of the Amundsen Sea Embayment (*2*) and (b) rate of ice-sheet elevation change (2003–2009) from ICESat GLAS laser altimetry (*45*).

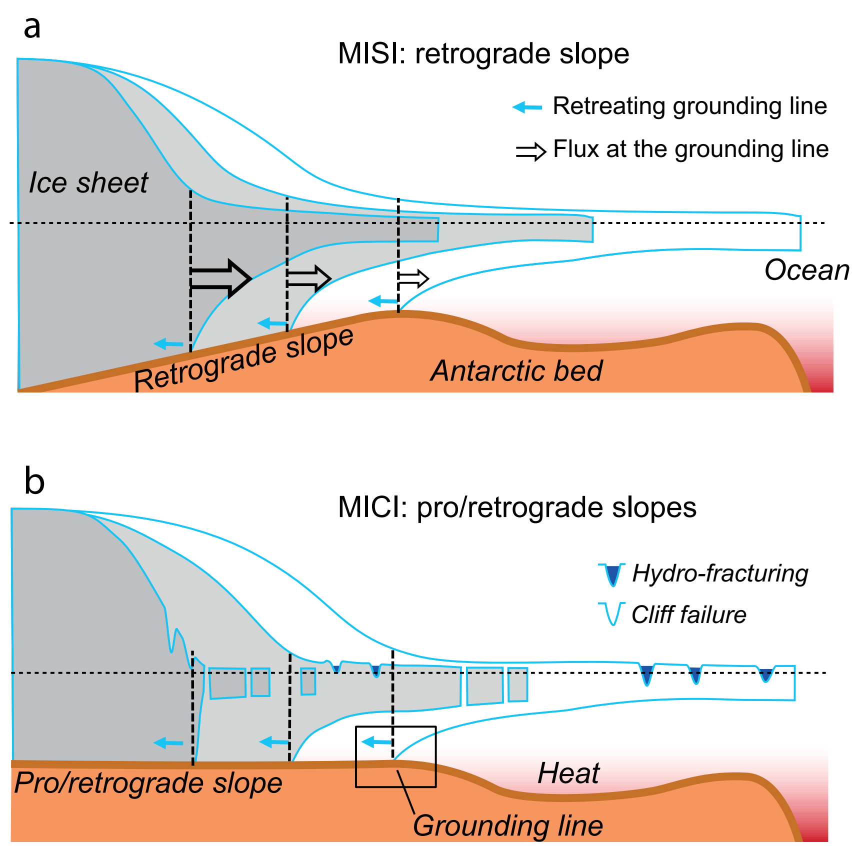

Figure 3: Schematic of the Marine Ice Sheet Instability (MISI) and the Marine Ice Cliff Instability (MICI). Redrawn after (*1*).

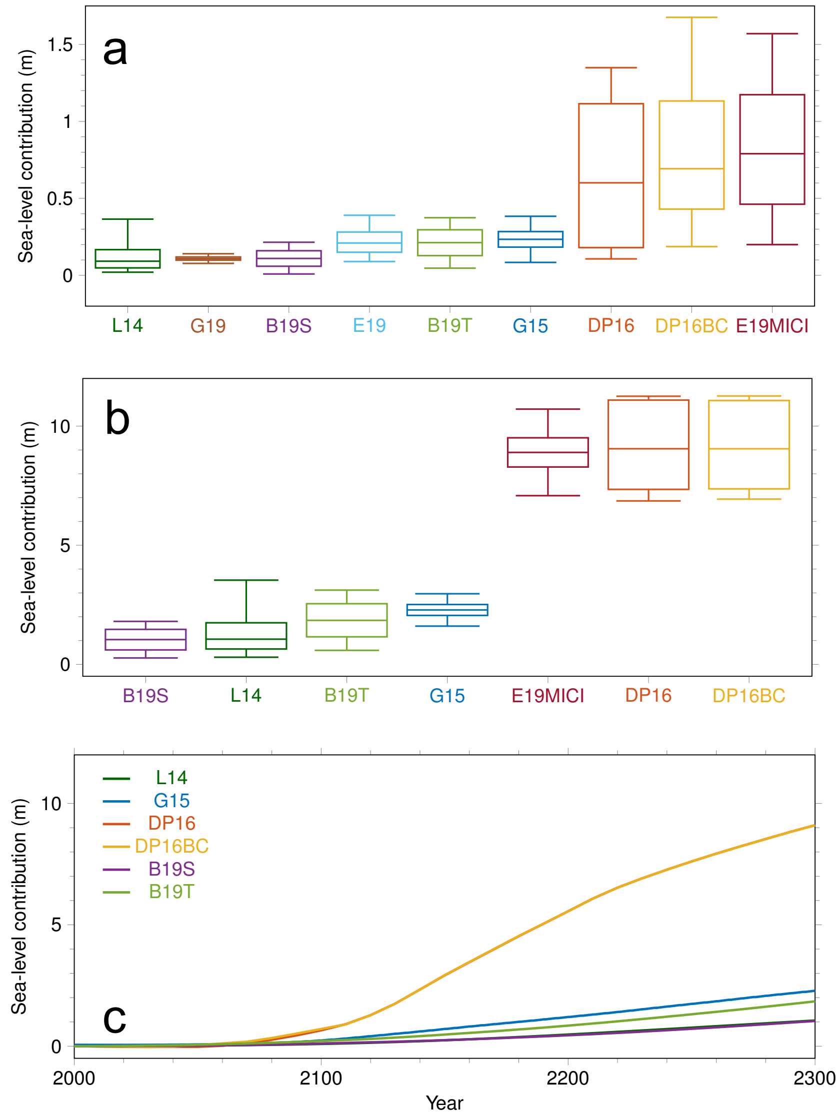

Figure 4: Projections of Antarctic sea-level contribution at (a) 2100 and (b) 2300 under RCP8.5 (*46*). Boxes and whiskers show the 5th, 25th, 50th, 75th and 95th percentiles. (c) Median projections of Antarctic sea-level contribution until 2300 (RCP8.5). Colour legend: L14 (*47*), G15 (*28*), DP16 (*14*), DP16BC: Bias-corrected simulations (*14*), B19S: Simulations with Schoof's parameterization (*30*), B19T: Simulations with Tsai's parameterization (*30*), E19: Simulations without MICI (*31*), E19MICI: Simulations with MICI (*31*), G19 (*32*).