



POTSDAM-INSTITUT FÜR
KLIMAFOLGENFORSCHUNG

Originally published as:

[Garbe, J., Albrecht, T., Levermann, A., Donges, J. F., Winkelmann, R.](#) (2020): The hysteresis of the Antarctic Ice Sheet. - Nature, 585, 7826, 538-544.

DOI: [10.1038/s41586-020-2727-5](https://doi.org/10.1038/s41586-020-2727-5)

The hysteresis of the Antarctic Ice Sheet

Julius Garbe^{1,2}, Torsten Albrecht¹, Anders Levermann^{1,2,3},
Jonathan F. Donges^{1,4} & Ricarda Winkelmann^{1,2†}

¹Potsdam Institute for Climate Impact Research (PIK), Member of the Leibniz Association, Potsdam, Germany.

²Institute of Physics and Astronomy, University of Potsdam, Potsdam, Germany.

³Lamont-Doherty Earth Observatory, Columbia University, New York, NY, USA.

⁴Stockholm Resilience Centre, Stockholm University, Stockholm, Sweden.

†e-mail: ricarda.winkelmann@pik-potsdam.de

More than half of Earth's freshwater resources and thus by far the largest potential to raise global sea level under future warming is held by the Antarctic Ice Sheet¹. Its long-term stability determines the fate of our coastal cities and cultural heritage. Feedbacks between ice, atmosphere, ocean, and the solid Earth give rise to potential nonlinearities in its response to temperature changes. So far, we are missing a comprehensive stability analysis of the Antarctic Ice Sheet under different amounts of global warming. Here we show that the Antarctic Ice Sheet exhibits a multitude of temperature thresholds beyond which ice loss is irreversible. Consistent with palaeodata² we find, using the Parallel Ice Sheet Model^{3,4,5}, that at global warming levels around 2 °C above pre-industrial West Antarctica is committed to long-term partial collapse due to the marine ice sheet instability. The precise threshold temperature bears some uncertainty due to model resolution and parameters. Between 6 to 9 °C of warming, the loss of more than 70 % of the present-day ice volume is triggered, mainly due to the surface elevation feedback. Above 10 °C Antarctica is committed to becoming virtually ice free. The ice sheet's temperature sensitivity is 1.3 metres of sea-level equivalent per degree up to 2 °C of warming, almost doubling to 2.4 m °C⁻¹ between 2 and 6 °C and increasing to ~10 m °C⁻¹ between 6 and 9 °C. Each of these thresholds gives rise to hysteresis behaviour, i.e. the currently observed ice-sheet configuration is not regained even if temperatures are reversed to present-day levels. In particular, the West Antarctic Ice Sheet does not regrow to its modern extent until temperatures are at least -1 °C colder than pre-industrial levels. Our results show that if the Paris Climate Agreement is not met, Antarctica's long-term sea-level contribution will dramatically increase and exceed that of all other sources.

The Antarctic Ice Sheet comprises an ice mass equivalent to 58 m of global sea-level rise¹. Its future evolution and the associated sea-level change are therefore of profound importance to coastal populations,

ecosystems and economies. Over the past decades, the ice sheet has been losing mass at an accelerating rate^{6,7}. Although the current net mass loss from Antarctica is small compared to the other sea-level rise contributions, it is likely to increase with progressing global warming⁸. Snowfall can be expected to increase in a warming atmosphere⁹, but this additional accumulation is likely to be counteracted and eventually overcompensated by ice-dynamical effects¹⁰.

The long-term stability of the Antarctic Ice Sheet under a changing climate remains an issue of ongoing research. It will be determined by the interplay between a number of negative (dampening) and positive (amplifying) feedbacks⁸. The latter might eventually lead to the crossing of critical thresholds, with the ice sheet entering into an irreversible dynamic, committing it to a specific amount of sea-level contribution.

One such self-amplifying feedback is the so-called surface-melt–elevation feedback: While there is very little surface melt in Antarctica at present, this might increase with strong global warming. The resulting lowered surface elevation exposes the ice to warmer temperatures through the atmospheric lapse rate, in turn leading to more melting^{11,12,13,14}. Once a critical temperature threshold is crossed, this melt-induced lowering of the ice-sheet surface elevation can trigger accelerated ice loss. In addition, the ice flow is generally accelerated under warmer temperatures, as viscosity decreases and deformation rates for a given stress increase. Through enhanced strain heating, this effect facilitates basal sliding and sped-up ice stream flow and might eventually result in a runaway process known as “creep instability”¹⁵.

Regions of the ice sheet that rest on bedrock below sea level are prone to additional feedback mechanisms, which potentially can drive the collapse of Antarctic Ice Sheet basins. Such regions – termed marine ice sheet regions – can be found in most of West Antarctica as well as in substantial parts of East Antarctica (e.g. the Aurora and Wilkes subglacial basins). One self-reinforcing feedback mechanism associated with these regions is the marine ice sheet instability (MISI) which implies a potential for irreversible grounding-line retreat on retrograde sloping bedrock^{16,17,8}. However, under the presence of sufficiently large ice-shelf buttressing, grounding lines on retrograde bed slopes can be conditionally stable¹⁸.

A potential negative feedback is introduced due to solid-Earth rebound: Marine ice sheet mass loss leads to a drop in sea level at the grounding line as a response to post-glacial rebound of the unloaded crust, which can effectively reduce the ice outflow, thereby acting as a stabilizing mechanism¹⁹.

There are several further positive and negative feedback mechanisms due to the interaction of the Antarctic Ice Sheet with the surrounding ocean, sea ice, atmosphere, and the solid Earth²⁰. These, however, are not included in the analysis presented here since, as discussed below, they either still require a deeper understanding or they can be expected to have a less significant impact on the long-term stability of the ice sheet.

Here we investigate the long-term response of the Antarctic Ice Sheet to various levels of warming with the fully dynamic Parallel Ice Sheet Model (PISM^{3,4,5}).

Owing to the interplay between the dampening and amplifying feedbacks, the response of the Antarctic ice volume to temperature change is expected to be highly nonlinear, and critical threshold behaviour might occur^{21,22}, leading to committed sea-level contribution upon transgression. This implies the existence of a

hysteresis, i.e. a path-dependent difference between multiple stable ice-sheet states, similar to what has been shown for the Greenland Ice Sheet²³. As of yet, however, large uncertainties remain concerning the warming levels at which these thresholds are reached and which regions of Antarctica could undergo irreversible ice loss under future warming^{21,24–28}.

In an earlier study²⁴ focusing on the terrestrial parts of the Antarctic Ice Sheet, a step-warming approach was used to localise the critical thresholds leading to large-scale retreat of the West and East Antarctic ice sheets at regional surface air temperature anomalies of about 8 °C and 16 °C, respectively. However, the applied model did not account for any ice–ocean interactions, which means that ice-sheet retreat is purely driven by surface mass balance processes and therefore the warming required to destabilise the marine regions of the West Antarctic Ice Sheet is likely to be overestimated.

From palaeorecords and modelling studies, we know that rapid transitions on the order of one to several thousand years between glacial, intermediate and collapsed states for West Antarctica are possible^{29–31}, and hysteresis effects might have occurred in the Cenozoic^{32,33}. During the global climatic shift towards colder temperatures near the Eocene–Oligocene boundary ~34 million years ago which set off the sudden, widespread glaciation of Antarctica³⁴, globally averaged surface temperatures have cooled by about 4 to 5 °C (ref.³⁵). Human action is currently heading at reversing this major transition, underpinning the urgency to adhere to the Paris Agreement’s target of limiting global warming to well below 2 °C in order to avoid the crossing of critical thresholds, committing us to long-term and possibly irreversible sea-level rise.

Long-term stability simulations

In order to study the future long-term response of the Antarctic Ice Sheet to changing global temperatures, we trace the ice sheet’s hysteresis with a technique that is routinely used to analyse, for instance, the stability of the Atlantic meridional overturning circulation and other climate components, as described in detail in ref.³⁶, section 2b: The global mean temperature (GMT), which we define here as the globally averaged surface air temperatures over land and ocean, is converted into regional changes of surface air and ocean temperature and ramped up incrementally until complete deglaciation is reached (for details, see Methods). The applied rate of GMT change is slower than the typical response timescale of the ice sheet and is chosen in a way that the system can follow the change while remaining as close as possible to equilibrium at all times. Advantages of this method compared to step-like temperature changes are that it avoids any abrupt overshoot effects and allows a continuous disentanglement of the relevant mass balance processes over the entire temperature range. Ideally, the branches of the hysteresis would be determined by an infinitely slow temperature change. Since available computational resources pose a constraint on the lower limit of this rate, we approximate this ideal case by applying the slowest computationally feasible rate of GMT change of 0.0001 °C yr⁻¹ in a quasi-static reference simulation (hereafter referred to as “quasi-static” simulations) and investigate the influence of faster rates on the ice sheet’s response by varying the applied rate between 0.0001 and 0.001 °C yr⁻¹. The quasi-static simulations are further extended at fixed temperature levels (at every full degree, as well as every half degree between 6 and 9 °C during retreat) until the ice sheet fully reaches a steady state, i.e. volume changes become negligible (hereafter referred to as “equilibrium” simulations). To investigate the ice sheet’s ability to regrow after complete disintegration, we reverse the temperature anomaly at the same rate (i.e. -0.0001 °C yr⁻¹ in the quasi-static reference case).

Our simulations are initialized from a reference equilibrium state which resembles the pre-industrial Antarctic Ice Sheet geometry as closely as possible while adequately approximating observed ice velocities at the same time (see Methods and Extended Data Figs. 1 and 2). We specifically aim for a good representation of the ice-sheet geometry in steady state, as starting from steady state is required for a systematic realization of the hysteresis methodology as described above. In a different study³⁷ we have shown that our model is also capable of adequately reproducing Antarctica's evolution over the last two glacial cycles.

Surface speeds range over more than five orders of magnitude from merely centimetres per year in the ice sheet's interior to several kilometres per year in the ice streams and ice shelves (Fig. 1).

Most of the ice sheet is surrounded by ocean waters with temperatures below zero. However, ice shelves in the Amundsen and Bellingshausen Sea are in contact with relatively warm ocean waters of up to 1.8 °C at the depth of the continental shelf region³⁸. Recent observations show that the glaciers, especially in the Amundsen region, have accelerated significantly over the past decades^{39,40} and a marine ice sheet collapse might already be underway in this sector^{41,42}.

Ice-sheet hysteresis

The equilibrium response of the Antarctic Ice Sheet reveals a strong hysteresis behaviour over the entire range of global warming up to around 10 °C above pre-industrial temperatures (Fig. 2, blue shaded area and triangle markers). Above these warming levels, Antarctica would eventually become virtually ice-free. Palaeoevidence suggests this has not been the case since ~34 million years ago³⁴, when global mean surface temperatures have been around 6 to 8 °C warmer compared to the pre-industrial average³⁵.

The rates of ice-sheet retreat and regrowth, i.e. the ice sheet's sensitivity to the external forcing, generally depend on the magnitude of the applied rate of GMT change (Fig. 2, grey shadings), with smaller forcing rates resulting in stronger retreat and regrowth. The quasi-static simulations (Fig. 2, blue curves) thereby approximate the equilibrium volumes (the 'true' hysteresis) most closely, as must be expected.

The decline of the ice sheet occurs in several disjunct stages: Initially, below 1 °C of warming, the ice volume in the quasi-static simulation in fact increases slightly due to the effect of additional snowfall⁹, especially in East Antarctica. The influence of enhanced snowfall is, however, minor compared to the overall mass losses of Antarctica in response to warmer temperatures. At warming levels between 1 to 2.5 °C, grounding lines in West Antarctica start strongly retreating (Supplementary Video 1), resulting in mass losses equivalent to more than 2 m of sea-level rise in equilibrium and even exceeding that value in the quasi-static simulation (that is, at a warming rate of 0.0001 °C yr⁻¹). The quasi-static curve up to 4 °C as obtained here is consistent with Antarctic palaeodata of the last five million years⁴³ as well as corresponding simulations with a different ice-sheet model²⁹ as shown in ref.⁴⁴.

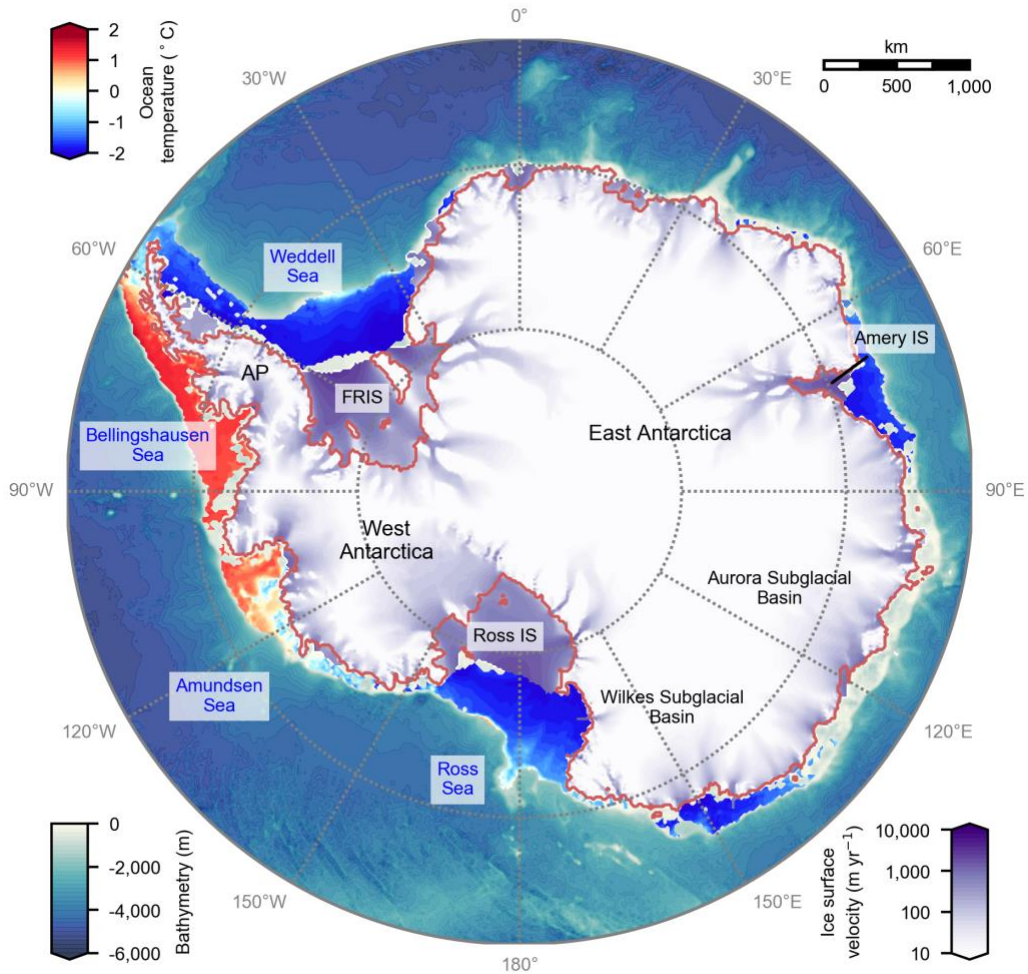


Fig. 1: Antarctic ice velocities and surrounding ocean temperatures.

Simulated ice surface velocities (in metres per year) of the reference ice-sheet state revealing the fast-flowing ice streams (purple shadings). Ocean temperatures at continental shelf depth (blue–red shading) are from ref.³⁸. The simulated grounding-line locations are shown in red. AP = Antarctic Peninsula, IS = ice shelf, FRIS = Filchner–Ronne Ice Shelf.

Between 6 to 9 °C above pre-industrial temperatures a second major threshold is found, manifesting in a loss of more than 70 % of the total ice mass upon transgression, corresponding to long-term global sea-level rise of more than 40 m SLE (metres sea-level equivalent; only accounting for ice volume above flotation).

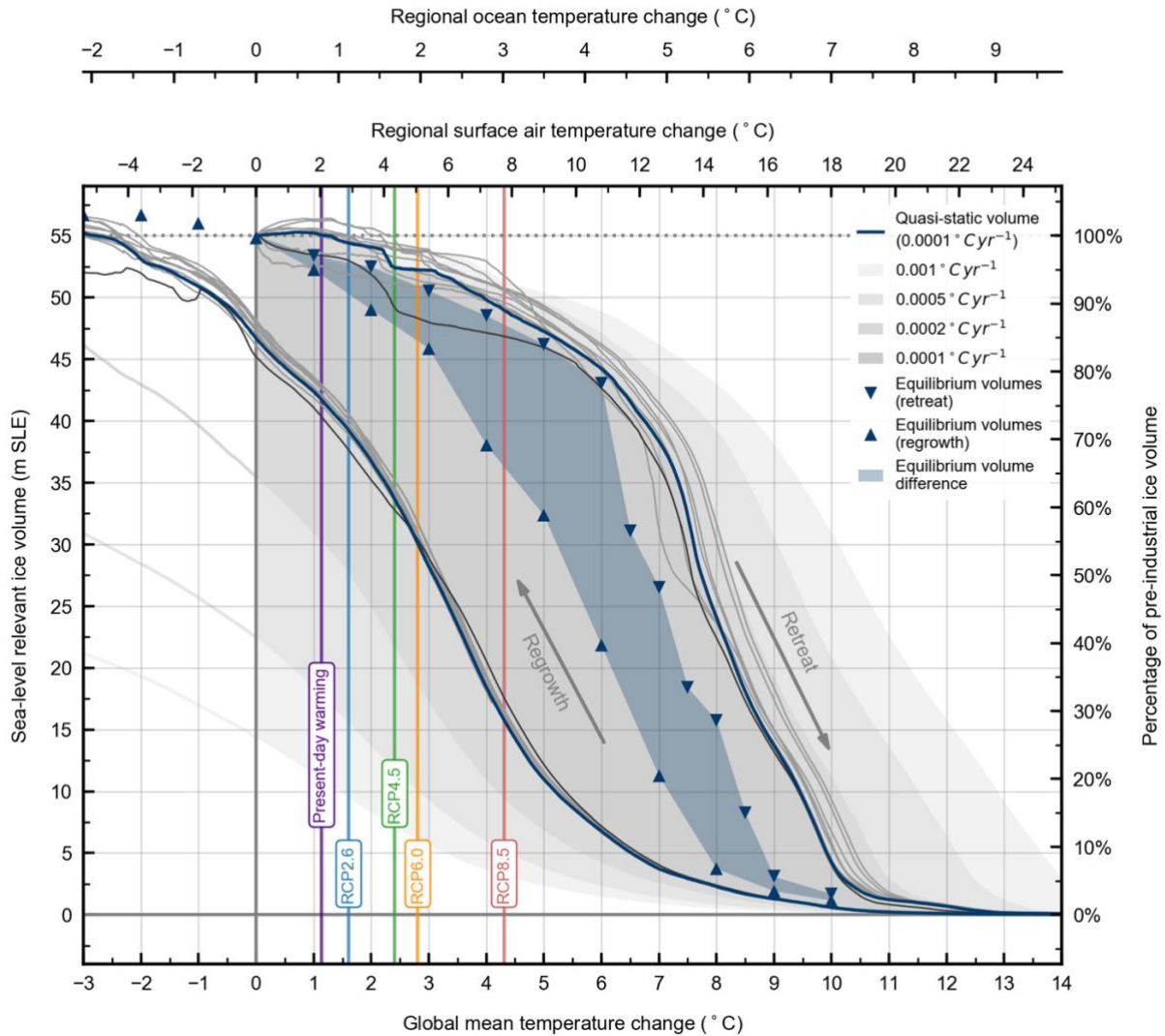


Fig. 2: Hysteresis of the Antarctic Ice Sheet.

Sea-level relevant ice volume (in metres sea-level equivalent, m SLE) for the quasi-static reference simulations (blue curve) as well as the corresponding equilibrium states at discrete temperature levels (blue triangles). The blue filled area marks the hysteresis gap, i.e. the equilibrium volume difference between the upper and lower hysteresis branches. Grey shadings correspond to different rates of applied GMT change. The quasi-static simulations based on an ensemble of perturbed model parameters are given as thin light grey lines and are shown in more detail in Extended Data Figure 4. The dark grey line denotes the quasi-static simulation using a model grid resolution of 8 km. Vertical coloured bars mark the GMT levels of present-day observed warming as well as expected end-of-century warming levels for the four different RCP scenarios⁸.

For some warming levels, equilibrium and quasi-static sea-level relevant ice volumes differ significantly, suggesting that the crossing of a threshold caused the system to diverge from its stable evolutionary path. Hence, it can be assumed that the critical temperature at which a large-scale decline of ice-sheet sectors is induced is therefore effectively lower than it appears in the quasi-static simulations. In particular, while ice-sheet sectors in West Antarctica seem to still be largely intact at 1 to 2 °C in the quasi-static simulations, the equilibrium response shows that even below this temperature level, major portions of the marine regions of the West Antarctic Ice Sheet are already committed to long-term collapse (compare Supplementary Video 1 and Fig. 4, upper panels), which is consistent with earlier results²⁷. Similar discrepancies in quasi-static and equilibrium ice loss can be seen at temperatures around 6 to 7 °C, associated with the collapse of major basins in East Antarctica.

The ice-sheet volume follows two substantially different paths during retreat and regrowth (Fig. 2, lower branch, and Supplementary Video 2). At pre-industrial temperatures, the ice sheet nearly reobtains its original sea-level relevant ice volume in equilibrium. However, the spatial ice-thickness distribution varies strongly from the starting configuration (Extended Data Fig. 3). In particular, West Antarctic grounding lines do not re-advance to their present-day locations until temperatures are at least -1 °C colder than pre-industrial levels. In the quasi-static simulations, the sea-level relevant ice volume deviates from the initial value by more than 8 m SLE at pre-industrial temperatures. The original volume is only regained at around -3 °C. Around -4 °C, the grounding lines in most regions re-advance close to their original position. This could suggest that the present-day configuration of the Antarctic Ice Sheet might indeed be a legacy of the last ice age, when temperatures were about 3 to 4 °C colder than today.

The sizable gap between the upper and lower branch of the hysteresis diagram shows that Antarctic ice-sheet retreat is to a great extent irreversible. The difference between the retreat and regrowth ice volumes is largest around 6 °C of warming, reaching more than 20 m SLE in equilibrium and even more than 35 m SLE in the quasi-static simulations (Fig. 3).

The discrepancy between quasi-static and equilibrium volume responses is even larger for the regrowth path, mainly due to the longer timescales associated with the build-up of the ice-sheet.

The ice-sheet response and general behaviour is robust with respect to changes in model resolution, ice-shelf calving, mantle response to ice loading, ice-stream sliding, and ice-flow enhancement factors. This applies in particular to the overall results for warming levels exceeding ~8 °C (Extended Data Fig. 4). Our sensitivity tests reveal an average uncertainty spread of around 5.5 m SLE for the tested range of parameter values, mainly originating from the retreat branch and the steepness of the hysteresis curve between 6 and 9 °C. The uncertainty at temperatures below ~5 °C is dominated by model resolution and the ice-flow enhancement factor, whereas at temperatures between 6 to 8 °C it is dominated by the sliding factor and mantle response, peaking at about 12 m SLE around 7 to 8 °C upon transgression of the second major threshold. The exact temperature threshold of marine ice sheet instabilities in West Antarctica and the Wilkes subglacial basin are highly sensitive to model parameters and resolution, pointing at the particularly low resilience of the system in this temperature range, and the respective thresholds need to be further constrained in a per-basin approach.

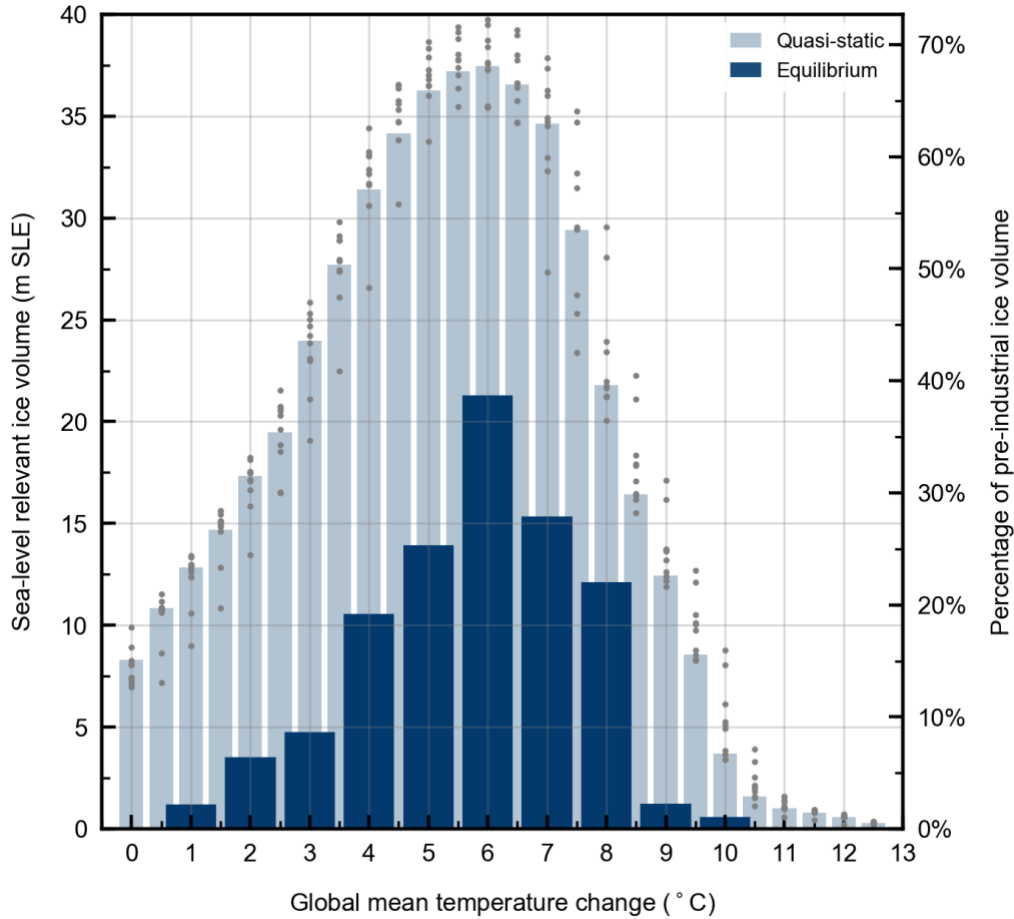


Fig. 3: Ice-sheet volume differences between retreat and regrowth.

The bar charts show the ice volume differences (in metres sea-level equivalent, m SLE) between the retreat and regrowth branches of the hysteresis curve for discrete warming levels. Light blue bars are based on the quasi-static simulations (blue lines in Fig. 2), dark blue bars on the equilibrium states (blue triangle markers in Fig. 2). Grey dots denote the uncertainty range for the quasi-static simulations (full ensemble spread of all tested model parameters, as detailed in Extended Data Fig. 4).

Figure 4 compares the ice-sheet equilibrium configurations during retreat and regrowth at different levels of warming. At 1 °C of warming, the Filchner-Ronne Ice Shelf is still largely intact following the retreat path, while it fails to regrow at the same temperature level on the regrowth path. At a warming level of 2 °C, various parts of the East Antarctic Ice Sheet that are still present at this temperature during retreat do not fully re-advance during regrowth, as, for instance, in the Wilkes subglacial basin. This is consistent with earlier studies which have found that Wilkes basin is currently protected by a small ice plug, but large-scale, irreversible retreat can occur once this ice plug is removed^{45,46}. This difference is even more pronounced at 4 °C, where additionally major parts of the Aurora subglacial basin are still ice-free during

regrowth. For temperatures between 6 and 8 °C, the difference is most prominent, with ice margins almost nowhere reaching their corresponding extents during the regrowth (see also Extended Data Fig. 5). At even higher warming levels, the differences between retreat and regrowth lessen again, when smaller ice caps can only exist at some high-altitude locations.

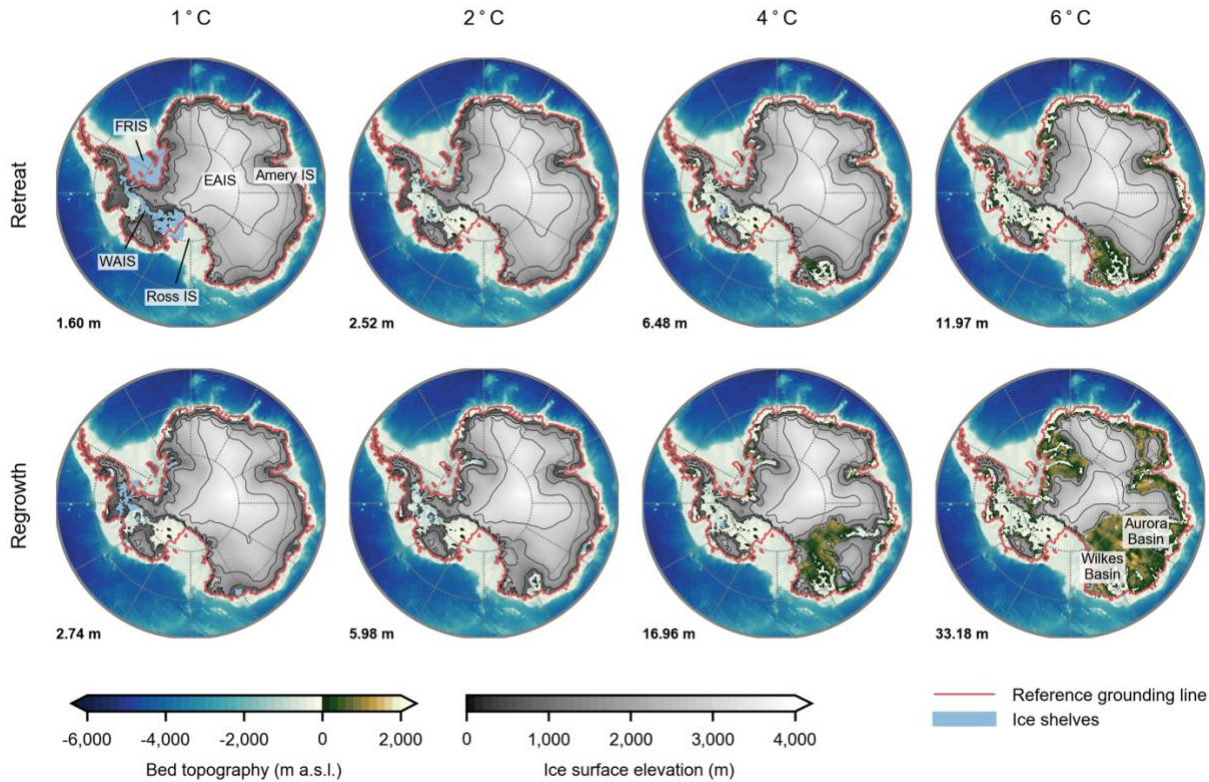


Fig. 4: Long-term ice loss for different warming levels.

Shown is the equilibrium ice-sheet surface elevation in metres for different warming levels (1, 2, 4, and 6 °C GMT anomaly above pre-industrial level), comparing the retreat (upper panels) and regrowth (lower panels) branch of the hysteresis curve. Ice surface-height contours are delineated at 1,000 m intervals. Grounding-line locations of the reference state are shown in red; ice shelves are marked in light blue. Absolute sea-level relevant ice-volume anomaly compared to the reference state (in metres sea-level equivalent, m SLE), i.e. the committed sea-level rise, is given for each panel. Blue shadings illustrate the bedrock depth below, brown shadings above present-day sea level (m a.s.l. = metres above sea level). EAIS = East Antarctic Ice Sheet, WAIS = West Antarctic Ice Sheet, IS = ice shelf, FRIS = Filchner–Ronne Ice Shelf.

Ocean- vs. atmosphere-induced changes

Disentangling the different relevant mass balance processes for the upper hysteresis branch reveals the distinct natures of the two major thresholds (Fig. 5): The collapse of the West Antarctic Ice Sheet around

2 °C of warming is mainly driven by increased sub-shelf melting and dynamic discharge (seen as major peak in the basal mass balance flux). This process is also associated with substantial grounding-line retreat and the subsequent formation of extensive shelf areas, hinting at an active MISI (Supplementary Video 1). At around 7 °C of warming, the overall surface mass balance turns negative. At these temperatures, the ice sheet has only very little contact with the ocean left. Consequently, in contrast to the first threshold, the strong decline of the East Antarctic Ice Sheet initiated around 6 to 7 °C of warming is dominated by the surface elevation feedback, resulting in a steep lowering of the ice sheet's surface altitude and reduction of its area (sharp drop in surface mass balance flux; see also Supplementary Video 1).

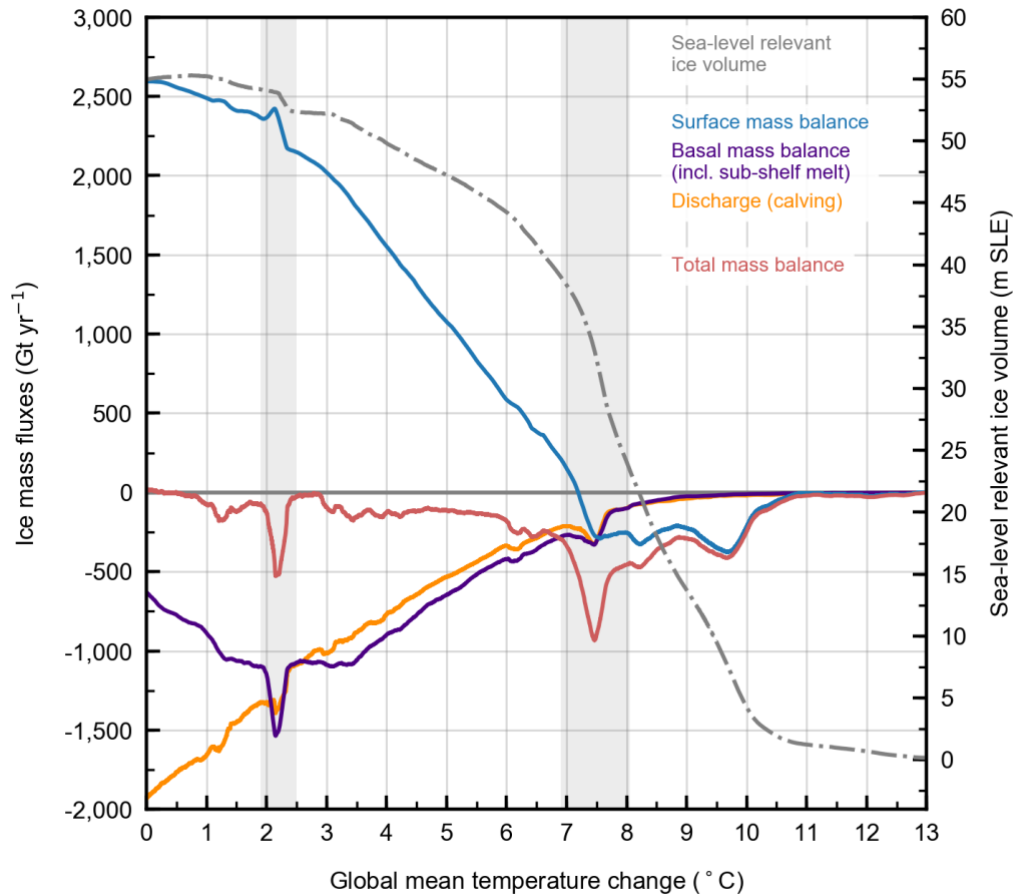


Fig. 5: Ocean-driven vs. atmosphere-driven ice loss.

Antarctic ice mass fluxes (in gigatonnes per year) showing the contributions of different atmospheric and oceanic processes to the total ice mass changes over the entire range of GMT anomalies along the upper hysteresis branch derived from the quasi-static reference simulation. Positive flux values denote mass gains, negative values mass losses. The vertical grey bars mark the locations of the two major temperature thresholds described in the text, which are associated with sharp drops in the total mass balance. The sea-level relevant ice-sheet volume (in metres sea-level equivalent, m SLE) is indicated by the dashed grey line with respect to the right-hand axis.

A similar split-up of the lower hysteresis branch confirms that certain processes are dominant for given temperature regimes (Extended Data Fig. 6): The regrowth of the ice sheet is generally dominated by surface mass balance processes. Mass loss from basal melt is highest around +1 °C, while in the retreat case the maximum is around +2 °C. Interestingly, the surface mass balance flux reaches a similar value at 0 °C GMT anomaly during both the retreat and regrowth phase. In both cases, mass losses through basal melt and discharge (iceberg calving) practically vanish for temperatures above 8 °C, when the ice sheet has no contact to the ocean (anymore) and basal melt from grounded ice is negligible. The majority of net mass loss during retreat occurs between 6 to 10 °C – mainly caused by the surface elevation feedback in East Antarctica – while most of the mass gain during regrowth happens at lower temperatures at 2 to 5 °C (resulting in a ‘flatter’ distribution in the phase space of total mass flux vs. warming). This is expected because, in order for ice to form, temperatures must first cool considerably, as the snowline first has to descend down to the ice-free bedrock surface.

Discussion and conclusion

Our analysis reveals a strong, multi-step hysteresis behaviour of the Antarctic Ice Sheet, confirming that in certain regions ice loss is in fact irreversible even if temperatures were to return to colder levels. While we here focus solely on the ice-internal feedbacks, it should be noted that additional feedback mechanisms like the ice–albedo feedback, as well as, for instance, sea-level changes due to the effect of self-gravitation or due to a large amount of meltwater added to the ocean, could further delay or accelerate the ice sheet’s response to sustained warming. One such positive feedback in the ice-ocean interaction, which might lead to warm waters being trapped below the sea surface, received renewed attention just recently⁴⁷.

Another hypothesized positive feedback process, termed the marine ice cliff instability (MICI), has only lately been brought to wider attention²⁵. Underpinned by observations of rapid ice-shelf and glacier retreat it has been suggested that hydrofracturing might lead to high coastal ice cliffs which become mechanically unstable at heights exceeding 90 m above the waterline, eventually causing the disintegration of marine ice sheets⁴⁸. However, this effect is still disputed^{49,50} and is not considered in the simulations presented here.

While our approach encompasses the dominant processes and feedbacks relevant for studying the long-term response of the Antarctic Ice Sheet, fully interactive ice–atmosphere–ocean simulations are required to project the Antarctic ice evolution under future emission trajectories.

In reality, temperatures have been and will likely continue to change at much faster rates than considered here. Our results should thus not be confused with sea-level projections – we deliberately choose small forcing rates here because this allows us to identify critical temperature thresholds decisive for the overall stability of the Antarctic Ice Sheet.

Our results imply that, if the Paris Agreement to limit global warming to well below 2 °C is not met, one or more critical thresholds might be subsequently crossed in Antarctica, committing us to long-term, possibly irreversible, sea-level rise of several up to dozens of metres.^{51,52}

1. Fretwell, P. et al. Bedmap2: improved ice bed, surface and thickness datasets for Antarctica. *The Cryosphere* 7, 375–393 (2013).
2. Turney, C. S. M. et al. Early Last Interglacial ocean warming drove substantial ice mass loss from Antarctica. *PNAS* 117, 3996–4006 (2020).
3. Bueler, E. & Brown, J. Shallow shelf approximation as a ‘sliding law’ in a thermomechanically coupled ice sheet model. *J. Geophys. Res.* 114, F03008 (2009).
4. Winkelmann, R. et al. The Potsdam Parallel Ice Sheet Model (PISM-PIK) – Part 1: Model description. *The Cryosphere* 5, 715–726 (2011).
5. The PISM Authors. PISM, a Parallel Ice Sheet Model: User's Manual. (<http://pism-docs.org/>, 2017).
6. The IMBIE team. Mass balance of the Antarctic Ice Sheet from 1992 to 2017. *Nature* 558, 219–222 (2018).
7. Rignot, E. et al. Four decades of Antarctic Ice Sheet mass balance from 1979–2017. *PNAS* 116, 1095–1103 (2019).
8. Intergovernmental Panel on Climate Change (IPCC). *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change.* Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA (2013).
9. Frieler, K. et al. Consistent evidence of increasing Antarctic accumulation with warming. *Nature Climate Change* 5, 348–352 (2015).
10. Winkelmann, R., Levermann, A., Martin, M. A. & Frieler, K. Increased future ice discharge from Antarctica owing to higher snowfall. *Nature* 492, 239–242 (2012).
11. Weertman, J. Stability of Ice-Age Ice Sheets. *J. Geophys. Res.* 66, 3783–3792 (1961).
12. Oerlemans, J. Some Basic Experiments with a Vertically-Integrated Ice-Sheet Model. *Tellus* 33, 1–11 (1981).
13. Huybrechts, P. et al. Response of the Greenland and Antarctic Ice Sheets to Multi-Millennial Greenhouse Warming in the Earth System Model of Intermediate Complexity LOVECLIM. *Surv Geophys* 32, 397–416 (2011).
14. Levermann, A. & Winkelmann, R. A simple equation for the melt elevation feedback of ice sheets. *The Cryosphere* 10, 1799–1807 (2016).
15. Clarke, G. K. C., Nitsan, U. & Paterson, W. S. B. Strain heating and creep instability in glaciers and ice sheets. *Reviews of Geophysics* 15, 235–247 (1977).

16. Weertman, J. Stability of the junction of an ice sheet and an ice shelf. *Journal of Glaciology* 13, 3–11 (1974).
17. Mercer, J. H. West Antarctic ice sheet and CO₂ greenhouse effect: a threat of disaster. *Nature* 271, 321–325 (1978).
18. Gudmundsson, G. H., Krug, J., Durand, G., Favier, L. & Gagliardini, O. The stability of grounding lines on retrograde slopes. *The Cryosphere* 6, 1497–1505 (2012).
19. Gomez, N., Pollard, D. & Holland, D. Sea-level feedback lowers projections of future Antarctic Ice-Sheet mass loss. *Nature Communications* 6, 8798 (2015).
20. Fyke, J., Sergienko, O., Löfverström, M., Price, S. F. & Lenaerts, J. T. M. An Overview of Interactions and Feedbacks Between Ice Sheets and the Earth System. *Reviews of Geophysics* 56, 361–408 (2018).
21. Golledge, N. R. et al. The multi-millennial Antarctic commitment to future sea-level rise. *Nature* 526, 421–425 (2015).
22. Winkelmann, R., Levermann, A., Ridgwell, A. & Caldeira, K. Combustion of available fossil fuel resources sufficient to eliminate the Antarctic Ice Sheet. *Science Advances* 1, e1500589 (2015).
23. Robinson, A., Calov, R. & Ganopolski, A. Multistability and critical thresholds of the Greenland ice sheet. *Nature Climate Change* 2, 429–432 (2012).
24. Huybrechts, P. Formation and disintegration of the Antarctic ice sheet. *Annals of Glaciology* 20, 336–340 (1994).
25. DeConto, R. M. & Pollard, D. Contribution of Antarctica to past and future sea-level rise. *Nature* 531, 591–597 (2016).
26. Sutter, J., Gierz, P., Grosfeld, K., Thoma, M. & Lohmann, G. Ocean temperature thresholds for Last Interglacial West Antarctic Ice Sheet collapse. *Geophys. Res. Lett.* 43, 2675–2682 (2016).
27. Golledge, N. R., Levy, R. H., McKay, R. M. & Naish, T. R. East Antarctic ice sheet most vulnerable to Weddell Sea warming. *Geophys. Res. Lett.* 44, 2343–2351 (2017).
28. Pattyn, F. et al. The Greenland and Antarctic ice sheets under 1.5 °C global warming. *Nature Climate Change* 8, 1053–1061 (2018).
29. Pollard, D. & DeConto, R. M. Modelling West Antarctic ice sheet growth and collapse through the past five million years. *Nature* 458, 329–332 (2009).
30. Alley, R. B. et al. Oceanic Forcing of Ice-Sheet Retreat: West Antarctica and More. *Annu. Rev. Earth Planet. Sci.* 43, 207–231 (2015).
31. Dutton, A. et al. Sea-level rise due to polar ice-sheet mass loss during past warm periods. *Science* 349, aaa4019 (2015).

32. Pollard, D. & DeConto, R. M. Hysteresis in Cenozoic Antarctic ice-sheet variations. *Global and Planetary Change* 45, 9–21 (2005).
33. Gasson, E. G. W., DeConto, R. M., Pollard, D. & Levy, R. Dynamic Antarctic ice sheet during the early to mid-Miocene. *PNAS* 113, 3459–3464 (2016).
34. Liu, Z. et al. Global Cooling During the Eocene-Oligocene Climate Transition. *Science* 323, 1187–1190 (2009).
35. Hansen, J., Sato, M., Russell, G. & Kharecha, P. Climate sensitivity, sea level and atmospheric carbon dioxide. *Phil. Trans. R. Soc. A* 371, 20120294 (2013).
36. Rahmstorf, S. & England, M. H. Influence of Southern Hemisphere winds on North Atlantic Deep Water flow. *Journal of Physical Oceanography* 27, 2040–2054 (1997).
37. Albrecht, T., Winkelmann, R. & Levermann, A. Glacial-cycle simulations of the Antarctic Ice Sheet with the Parallel Ice Sheet Model (PISM) – Part 1: Boundary conditions and climatic forcing. *The Cryosphere* 14, 599–632 (2020).
38. Schmidtko, S., Heywood, K. J., Thompson, A. F. & Aoki, S. Multidecadal warming of Antarctic waters. *Science* 346, 1227–1231 (2014).
39. Mouginot, J., Rignot, E. & Scheuchl, B. Sustained increase in ice discharge from the Amundsen Sea Embayment, West Antarctica, from 1973 to 2013. *Geophys. Res. Lett.* 41, 1576–1584 (2014).
40. Rignot, E., Mouginot, J., Morlighem, M., Seroussi, H. & Scheuchl, B. Widespread, rapid grounding line retreat of Pine Island, Thwaites, Smith, and Kohler glaciers, West Antarctica, from 1992 to 2011. *Geophys. Res. Lett.* 41, 3502–3509 (2014).
41. Favier, L. et al. Retreat of Pine Island Glacier controlled by marine ice-sheet instability. *Nature Climate Change* 4, 117–121 (2014).
42. Joughin, I., Smith, B. E. & Medley, B. Marine ice sheet collapse potentially under way for the Thwaites Glacier Basin, West Antarctica. *Science* 344, 735–738 (2014).
43. Naish, T. R. et al. Obliquity-paced Pliocene West Antarctic ice sheet oscillations. *Nature* 458, 322–328 (2009).
44. Levermann, A. et al. The multimillennial sea-level commitment of global warming. *PNAS* 110, 13745–13750 (2013).
45. Mengel, M. & Levermann, A. Ice plug prevents irreversible discharge from East Antarctica. *Nature Climate Change* 4, 451–455 (2014).
46. Golledge, N. R. et al. Antarctic climate and ice-sheet configuration during the early Pliocene interglacial at 4.23 Ma. *Clim. Past* 13, 959–975 (2017).

47. Golledge, N. R. et al. Global environmental consequences of twenty-first-century ice-sheet melt. *Nature* 566, 65–72 (2019).
48. Bassis, J. N. & Walker, C. C. Upper and lower limits on the stability of calving glaciers from the yield strength envelope of ice. *Proc. R. Soc. A* 468, 913–931 (2012).
49. Edwards, T. L. et al. Revisiting Antarctic ice loss due to marine ice-cliff instability. *Nature* 566, 58–64 (2019).
50. Meredith, M. et al. Polar Regions. In: IPCC Special Report on the Ocean and Cryosphere in a Changing Climate [H.-O. Pörtner, D.C. Roberts, V. Masson-Delmotte, P. Zhai, M. Tignor, E. Poloczanska, K. Mintenbeck, A. Alegría, M. Nicolai, A. Okem, J. Petzold, B. Rama, N.M. Weyer (eds.)]. In press (2019).
51. Schellnhuber, H. J., Rahmstorf, S. & Winkelmann, R. Why the right climate target was agreed in Paris. *Nat. Clim. Chang.* 6, 649–653 (2016).
52. Lenton, T. M. et al. Climate tipping points—too risky to bet against. *Nature* 575, 592–595 (2019).

Methods

Ice-sheet model

We use a modified version of the Parallel Ice Sheet Model (PISM) release v1.0, an open-source, three-dimensional, thermo-mechanically coupled ice-sheet/ice-shelf model^{3,4,5}. Running in hybrid mode, PISM combines equations of the shallow-ice approximation (SIA) and shallow-shelf (or shelfy-stream) approximation (SSA) of the Stokes flow over the entire ice-sheet/ice-shelf domain. Superimposing the SIA and SSA velocity solutions enables a consistent transition of stress regimes across the grounded-ice to floating-ice boundary⁴. The ice rheology is based on the Glen-Paterson-Budd-Lliboutry-Duval flow law⁵³ with Glen exponent $n = 3$.

Basal shear stress in the vicinity of the grounding line is computed at subgrid scale using a linear interpolation of the height above buoyancy between adjacent grounded and floating cells⁵⁴. Melt rates at the shelf base are not interpolated across the grounding line.

Basal sliding and subglacial hydrology

In PISM, basal sliding is parameterized based on a generalized power-law formulation⁵⁵:

$$\boldsymbol{\tau}_b = -\tau_c \frac{\mathbf{u}_b}{u_{thres}^q |\mathbf{u}_b|^{1-q}},$$

where $\boldsymbol{\tau}_b$ is the basal shear stress, \mathbf{u}_b is the basal sliding velocity, τ_c is the yield stress (see below) and u_{thres} a threshold velocity, chosen here to be 100 m yr^{-1} . Depending on the sliding exponent $0 \leq q \leq 1$, this “pseudo-plastic” approach spans a wide range from purely plastic Coulomb sliding ($q = 0$) to sliding in which basal velocity and basal shear stress are linearly related ($q = 1$). In our simulations we use a sliding exponent of $q = 0.75$.

The yield stress τ_c is determined as a function of till material properties and of the effective till pressure N_{till} , using the Mohr-Coulomb criterion⁵⁶:

$$\tau_c = \tan(\phi) N_{till}.$$

ϕ is called the till friction angle, a heuristic shear strength parameter for the till material property, which is optimized iteratively in the grounded-ice region such that the mismatch of equilibrium and modern surface elevation is minimized³⁷.

To determine the effective pressure on the till, the modelled amount of water in the till layer W_{till} is used. Meltwater generated in the subglacial layer is stored locally in the till up to a thickness of saturated substrate of $W_{till}^{max} = 2 \text{ m}$. The hydrology parameterization is non-conserving in the sense that additional water above this maximum thickness is lost permanently.

With the effective water thickness of the till layer $s = W_{till}/W_{till}^{max}$ and the ice overburden pressure P_0 , N_{till} is parameterized as in ref.⁵⁷:

$$N_{till} = \min \left\{ P_0, N_0 \left(\frac{\delta P_0}{N_0} \right)^s 10^{(e_0/C_c)(1-s)} \right\}$$

For the constant parameters N_0 , e_0 , and C_c , we use the values adopted from ref.⁵⁷. For saturated till ($s = 1$) we find $N_{till}^{min} = \delta P_0$, where δ is a certain fraction of the overburden (here set to 4 %), at which the till reaches maximum capacity and excess water will be drained. Note, that the effective till pressure cannot exceed the overburden pressure, i.e. $N_{till}^{max} = P_0$.

Glacial isostatic adjustment

Changes in the elevation of the bed topography due to changes in ice load are modelled using a sophisticated Earth-deformation model based on refs.^{58,59}. It consists of a purely elastic plate lithosphere overlying a linearly viscous half-space (of viscosity $\eta = 1 \times 10^{21}$ Pa s and density $\rho = 3300$ kg m⁻³) that represents the upper mantle (see Fig. 3 in ref.⁵⁸). The half-space model utilizes a time-dependent partial differential equation, which generalizes and improves upon the widely used Elastic Lithosphere Relaxing Asthenosphere (ELRA) model⁵⁹. This allows for computationally inexpensive calculations based on the Fast Fourier transform. A further advantage of this formulation is that the relaxation time of the bed elevation is not considered a constant, instead, relaxation times are mode-dependent, thus closely approximating the approach used within more complex GIA models. As in our simulations typical ice-dynamical timescales are much longer than those of the elastic part of the Earth model (~10 years), we only applied the viscous part.

Surface mass balance

Mean annual and mean summer surface air temperatures are parameterized based on multiple regression analysis of ERA-Interim data⁶⁰, as a function of latitude and surface elevation, using an atmospheric lapse rate Γ of -8.2 °C km⁻¹ (ref.³⁷). For surface accumulation we use the 1986–2005 mean precipitation from the output of the Regional Atmospheric Climate MOdel (RACMOv2.3p2; ref.⁶¹). As a modification from PISM release v1.0, we introduce a climatic correction for precipitation P based on the difference between ice-model surface elevation and the interpolated elevation in the climate model or observational data set. The base precipitation pattern P_0 is now scaled exponentially with model ice surface elevation change Δh , using an exponential factor of 41 % per km of elevation change, similar to the approach followed by ref.²⁵:

$$P = P_0 \times \exp(\gamma \cdot \Delta h), \text{ where } \gamma = 5 \% \text{ } ^\circ\text{C}^{-1} \times \Gamma$$

Thereby, the factor γ accounts for a 5 % change in accumulation per degree of atmospheric temperature change⁹, assuming a linear relationship between these two quantities. For consistency reasons, the exponential correction of precipitation is also used for the applied temperature anomaly forcing using the same γ .

Ice surface melt and runoff are computed using a positive degree-day (PDD) scheme, with melt coefficients of 3 mm PDD⁻¹ for snow and 9 mm PDD⁻¹ for ice, respectively, and also accounting for natural variability using a 5 °C standard deviation.

Sub-shelf melting

Sub-shelf melt rates are computed using the Potsdam Ice-shelf Cavity mOdel (PICO; ref.⁶²). PICO extends the ocean box model approach by ref.⁶³, simulating the vertical overturning circulation in ice-shelf cavities, thus enabling the computation of sub-shelf melt rates consistent with this circulation. It mimics the

exchange with ocean water masses in front of the ice shelves, which is physically based on the “ice-pump” mechanism, using parameters for overturning strength and turbulent heat exchange of $0.5 \times 10^6 \text{ m}^6 \text{ s}^{-1} \text{ kg}^{-1}$ and $1 \times 10^{-5} \text{ m s}^{-1}$, respectively.

Calving

Iceberg calving at the ice-shelf margin is calculated from principal strain rates⁶⁴. In this “eigen-calving” approach, the average calving rate is proportional to the product of principal components of the horizontal strain rates derived from SSA-velocities at the shelf front. The proportionality factor is set to be $1 \times 10^{17} \text{ m s}$. Additionally, we apply a minimum thickness criterion of 50 m (PISM default) at the calving front, based on the observation that ice-shelf calving fronts commonly have a minimum thickness and appear to be unstable below that⁶⁵. Ice shelves are also removed in all areas which were “open ocean” during the initial timestep. The two latter calving conditions are mainly imposed due to numerical reasons, but show almost no influence on the ice volume, in particular during the regrowth phase (Extended Data Fig. 4b).

Ice-sheet reference state

To define appropriate ice-sheet initial conditions for our experiments, we generate the required equilibrium state under climatic boundary conditions from the available data from the second half of the 20th century. Since the Antarctic Ice Sheet has experienced notable changes during the past decades due to anthropogenic climate change, much larger than the natural variability experienced over most of the Holocene⁶⁶, we interpret the pre-industrial ice-sheet configuration as the closest analogue to this equilibrium state. Temperature anomalies are therefore taken with respect to pre-industrial levels in all experiments presented here.

The equilibrium state is a slightly modified version of an equilibrium state generated as part of the initial state intercomparison exercise (initMIP-Antarctica; ref.⁶⁷), the latest 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. The initMIP equilibrium simulation was initialized from Bedmap2 geometry¹, with surface mass balance from RACMOv2.3p2 1986–2005 mean⁶¹ and observed values of ocean temperature and salinity at the sea floor on the continental shelf, averaged over the period 1975–2012 (ref.³⁸), to drive PICO. (For more details, see appendix B12 of ref.⁶⁷.)

Using the initMIP equilibrium as input, we extend the simulation for another 150 kyr with our version of PISM. Main model modifications relate to the bedrock now being allowed to deform under changing ice-sheet geometry as well as the parameterization of the surface air temperature in the atmosphere module, now being based on multiple regression analysis of ERA Interim data, from which the surface melt is calculated via the above-described PDD scheme.

All simulations were performed on a regular rectangular grid with 16 km horizontal resolution. The vertical grid is quadratically spaced, ranging between 20 m at the ice base and 100 m at the top of the thickest ice domes. The simulations shown in Extended Data Figure 4g use a horizontal grid resolution of 8 km and a vertical grid spacing ranging from 13 m at the ice base to 87 m at the upper ice surface.

Climate and ocean inputs and forcings

To study the long-term response of the Antarctic Ice Sheet to changing global temperatures, we trace the ice sheet's hysteresis with respect to temperature changes with an approach based on section 2b of ref.³⁶. In this approach, a spatially uniform temperature anomaly that is gradually changing over time is applied to the boundary climate in the model. The rate of warming is slower than the typical response timescale of the ice sheet. This ensures that the system can follow the change, remaining as close as possible to equilibrium at all times, while taking into account computational constraints. Our simulation is initialized using the ice-sheet reference state described above and then the incrementally increasing temperature anomaly is applied until complete deglaciation is achieved. To trace the lower branch of the hysteresis, the temperature anomaly is then incrementally decreased again, starting from the bare bedrock, until the ice sheet is regrown to its initial extent. Different warming rates have been tested (0.001, 0.0005, 0.0002, and 0.0001 °C yr⁻¹). The rate of change of GMT anomaly used here to derive the quasi-static reference hysteresis diagram (blue curve in Fig. 2) is 0.0001 °C yr⁻¹. At every full degree, as well as every half degree between 6 and 9 °C on the upper branch, we further extend the simulations at fixed temperatures until the ice sheet fully reaches a steady state, i.e. volume changes become negligible.

In order to translate the global mean surface air temperature anomaly into regional atmospheric and oceanic temperature changes, the GMT anomalies are scaled using constant scaling factors. In a long-term simulation with the coupled climate model ECHAM5/MPIOM it has been shown that the ratios between the GMT and the near-surface air temperature and ocean temperature at about 400 m depth in the region south of 66 °S are almost constant on long timescales⁶⁸. In these simulations the scaling factors approach values of 1.8 and 0.7 with respect to GMT for the regional near-surface air temperature and the ocean temperature, respectively (section 4.3 of ref.³⁷).

Sensitivity ensemble

An ensemble of model sensitivity simulations was carried out to verify the robustness of the presented findings with respect to changes in various critical model parameters. These include the horizontal grid resolution, the maximum allowed extents for ice-shelf calving, the viscous response of the upper mantle to ice loading, ice-stream sliding, and the SSA ice-flow enhancement factor⁴. The results of the sensitivity ensemble are shown in Extended Data Figure 4.

Data availability

All data used for this study are publicly available. Antarctic surface mass balance data from RACMO2.3p2 were downloaded from https://www.projects.science.uu.nl/iceclimate/publications/data/2018/index.php#vwessem2018_tc. Antarctic bedrock topography and ice thickness data are from the Bedmap2 compilation, available at <https://secure.antarctica.ac.uk/data/bedmap2/>. The Schmidtke ocean temperature and salinity dataset can be retrieved at <https://www.geomar.de/en/staff/fb1/po/sschmidtke/southern-ocean/>. The datasets generated and analysed during this study are available from the corresponding author upon reasonable request.

Code availability

PISM is freely available as open-source code from <https://github.com/pism/pism>. The code version used in this study is available at <https://doi.org/10.5281/zenodo.3956431>. PISM input data are pre-processed using <https://github.com/pism/pism-ais> with original data citations.

53. Lliboutry, L. A. & Duval, P. Various isotropic and anisotropic ices found in glaciers and polar ice caps and their corresponding rheologies. *Annales Geophys.* 3, 207–224 (1985).
54. Feldmann, J., Albrecht, T., Khroulev, C., Pattyn, F. & Levermann, A. Resolution-dependent performance of grounding line motion in a shallow model compared with a full-Stokes model according to the MISMIP3d intercomparison. *Journal of Glaciology* 60, 353–360 (2014).
55. Schoof, C. & Hindmarsh, R. C. A. Thin-Film Flows with Wall Slip: An Asymptotic Analysis of Higher Order Glacier Flow Models. *The Quarterly Journal of Mechanics and Applied Mathematics* 63, 73–114 (2010).
56. Cuffey, K. M. & Paterson, W. S. B. *The Physics of Glaciers*. Elsevier, 4th edition. Academic Press, Amsterdam, etc. (2010).
57. Bueler, E. & van Pelt, W. Mass-conserving subglacial hydrology in the Parallel Ice Sheet Model version 0.6. *Geosci. Model Dev.* 8, 1613–1635 (2015).
58. Lingle, C. S. & Clark, J. A. A Numerical Model of Interactions Between a Marine Ice-Sheet and the Solid Earth: Application to a West Antarctic Ice Stream. *Journal of Geophysical Research-Oceans* 90, 1100–1114 (1985).
59. Bueler, E., Lingle, C. S. & Brown, J. Fast computation of a viscoelastic deformable Earth model for ice-sheet simulations. *Annals of Glaciology* 46, 97–105 (2007).
60. Dee, D. P. et al. The ERA-Interim reanalysis: configuration and performance of the data assimilation system. *Q.J.R. Meteorol. Soc.* 137, 553–597 (2011).
61. van Wessem, J. M. et al. Modelling the climate and surface mass balance of polar ice sheets using RACMO2 – Part 2: Antarctica (1979–2016). *The Cryosphere* 12, 1479–1498 (2018).
62. Reese, R., Albrecht, T., Mengel, M., Asay-Davis, X. & Winkelmann, R. Antarctic sub-shelf melt rates via PICO. *The Cryosphere* 12, 1969–1985 (2018).
63. Olbers, D. & Hellmer, H. A box model of circulation and melting in ice shelf caverns. *Ocean Dynamics* 60, 141–153 (2010).
64. Levermann, A. et al. Kinematic first-order calving law implies potential for abrupt ice-shelf retreat. *The Cryosphere* 6, 273–286 (2012).

65. Albrecht, T., Martin, M. A., Haseloff, M., Winkelmann, R. & Levermann, A. Parameterization for subgrid-scale motion of ice-shelf calving fronts. *The Cryosphere* 5, 35–44 (2011).
66. Cuffey, K. M. et al. Deglacial temperature history of West Antarctica. *PNAS* 113, 14249–14254 (2016).
67. Seroussi, H. et al. initMIP-Antarctica: an ice sheet model initialization experiment of ISMIP6. *The Cryosphere* 13, 1441–1471 (2019).
68. Li, C., von Storch, J.-S. & Marotzke, J. Deep-ocean heat uptake and equilibrium climate response. *Clim Dyn* 40, 1071–1086 (2012).
69. Rignot, E., Mouginot, J. & Scheuchl, B. Ice Flow of the Antarctic Ice Sheet. *Science* 333, 1427–1430 (2011).

Acknowledgements J.G., J.F.D., and R.W. were supported by the Leibniz Association project DominoES. T.A. and R.W. are supported by the Deutsche Forschungsgemeinschaft (DFG) in the framework of the priority program “Antarctic Research with comparative investigations in Arctic ice areas” by grants WI4556/2-1 and WI4556/4-1 and in the framework of the PalMod project (FKZ: 01LP1925D), supported by the German Federal Ministry of Education and Research (BMBF) as a Research for Sustainability initiative (FONA). J.F.D. is grateful for financial support by the Stordalen Foundation via the Planetary Boundary Research Network (PB.net), the Earth League’s EarthDoc program, and the European Research Council Advanced Grant project ERA (Earth Resilience in the Anthropocene; grant ERC-2016-ADG-743080). This research was supported by the European Union’s Horizon 2020 research and innovation programme under grant agreement no. 820575 (TiPACCs). Development of PISM is supported by NASA grant NNX17AG65G and NSF grants PLR-1603799 and PLR-1644277. The authors further acknowledge the European Regional Development Fund (ERDF), the German Federal Ministry of Education and Research (BMBF) and the Land Brandenburg for supporting this project by providing resources on the high-performance computer system at the Potsdam Institute for Climate Impact Research. We thank Matthias Mengel for providing the Antarctic equilibrium used as a basis for the simulations.

Author contributions R.W. conceived the study. All authors designed the research and contributed to the analysis. J.G. carried out the model simulations. J.G. and R.W. wrote the manuscript with contributions from all co-authors.

Competing interests The authors declare no competing interests.

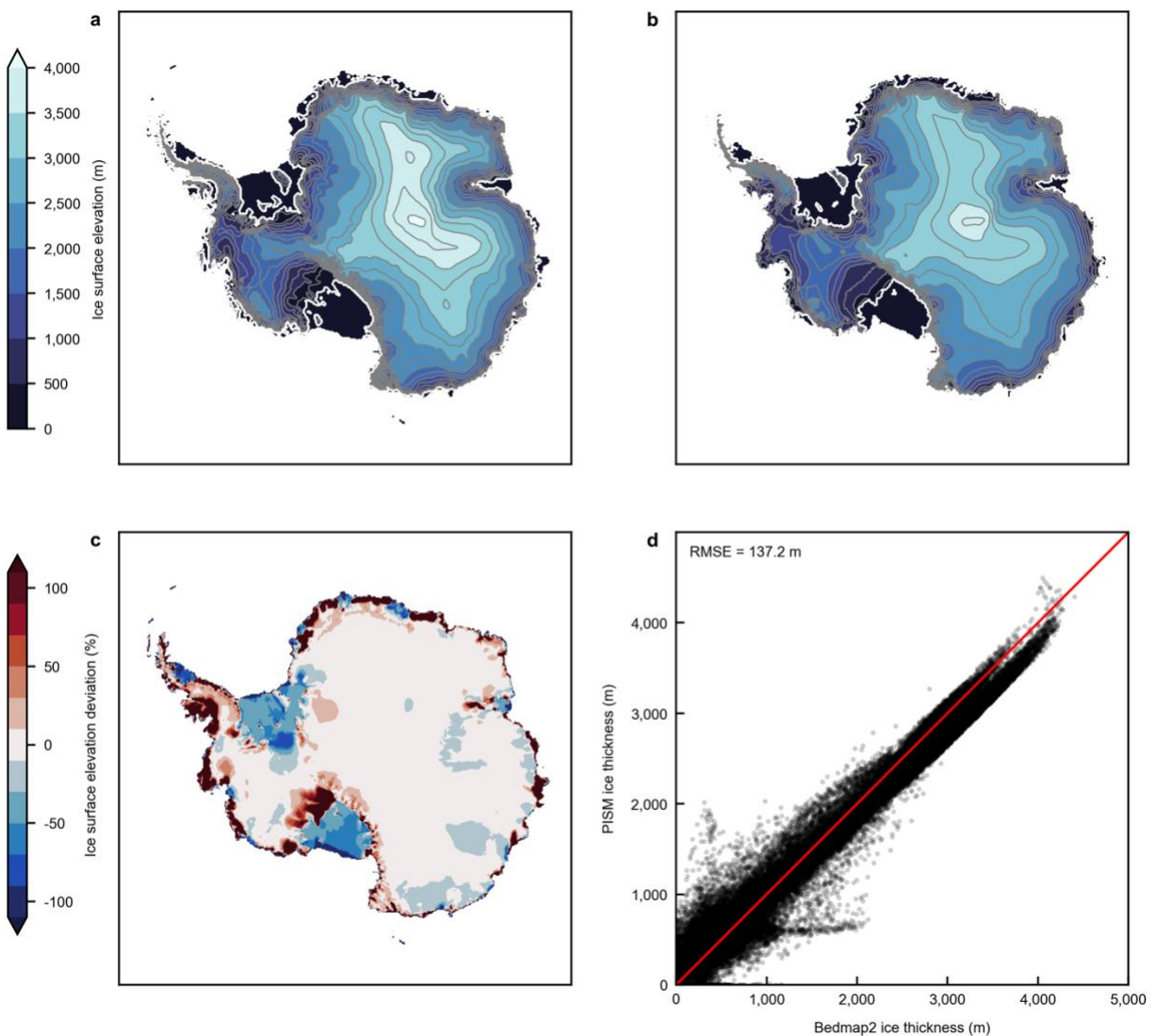
Additional Information

Supplementary Information is available for this paper.

Correspondence and requests for materials should be addressed to R.W.

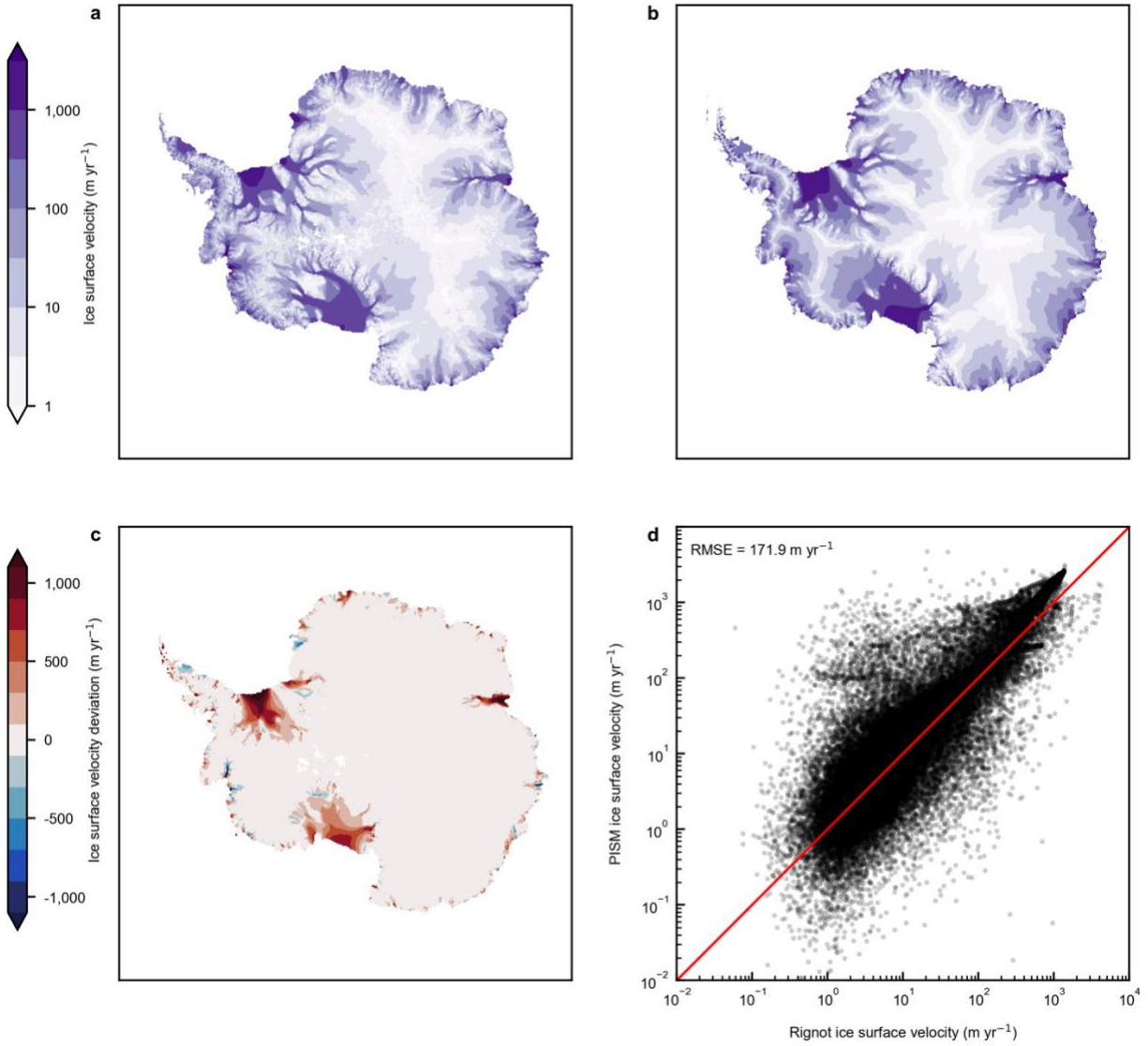
Reprints and permissions information is available at www.nature.com/reprints.

Extended Data



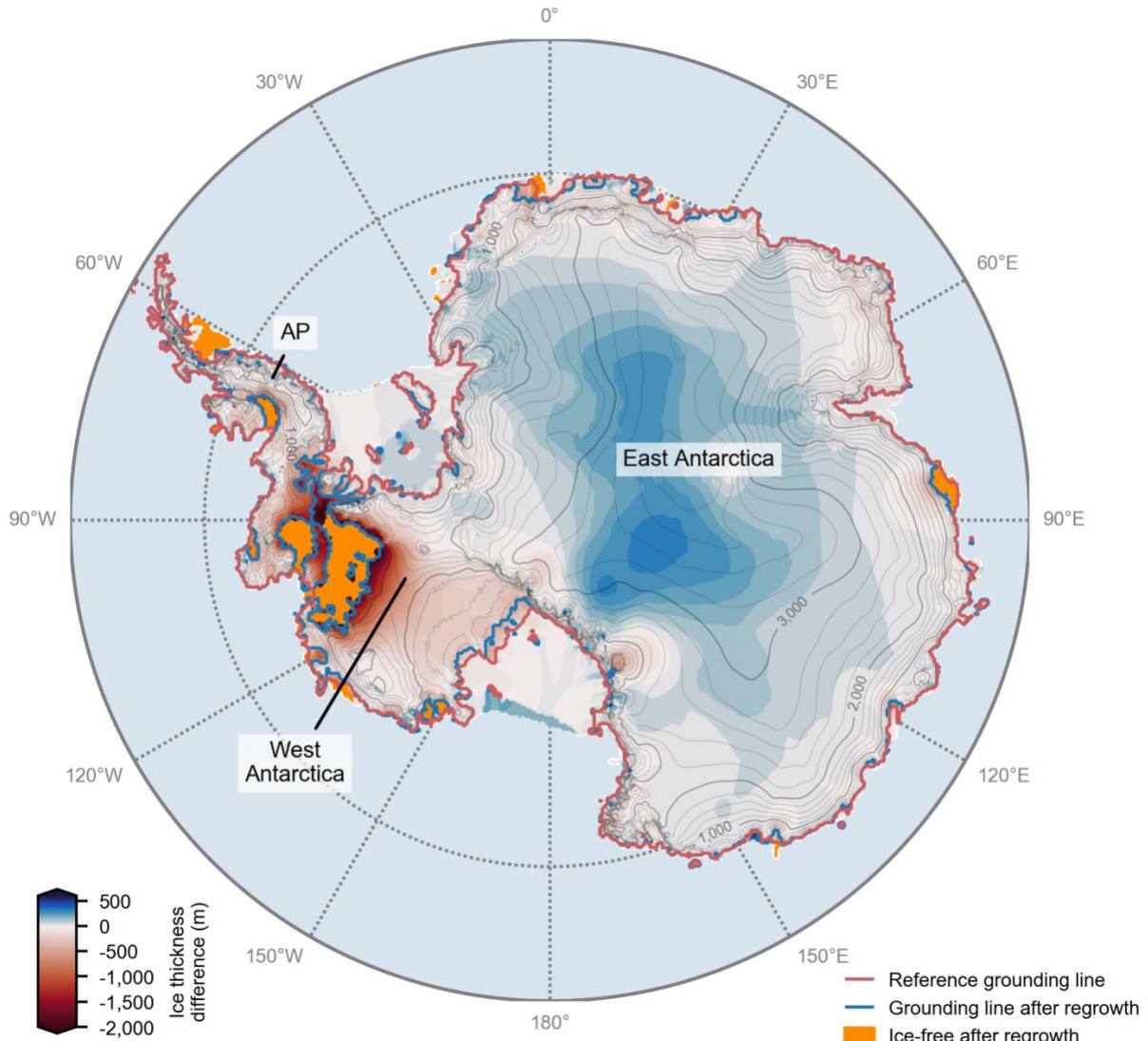
Extended Data Fig. 1: Comparison of modelled and observed ice geometry.

(a) Observed ice surface elevation (in metres; Bedmap2, ref.¹), regridded to 16 km. Grounding lines are shown in white; surface-height contours are given every 250 m. (b) Modelled ice surface elevation (in metres) of the reference state serving as initial configuration for the hysteresis experiments. Grounding lines are shown in white; surface-height contours are given every 250 m. (c) Modelled minus observed (Bedmap2) ice surface elevation (in per cent). (d) Scatter plot for comparison of modelled and observed ice thickness (in metres) for each grid cell. Red identity line illustrates where modelled ice thickness would perfectly match the observations. RMSE = root-mean-square error.



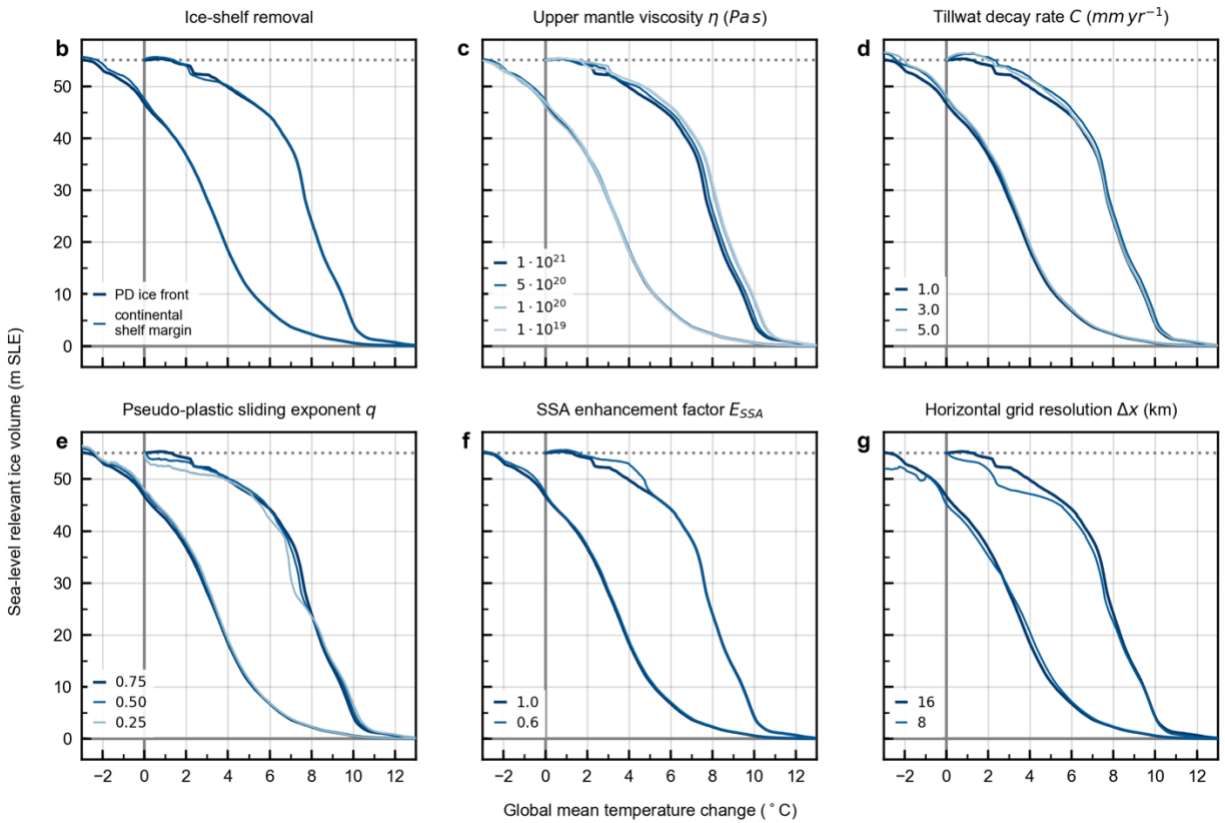
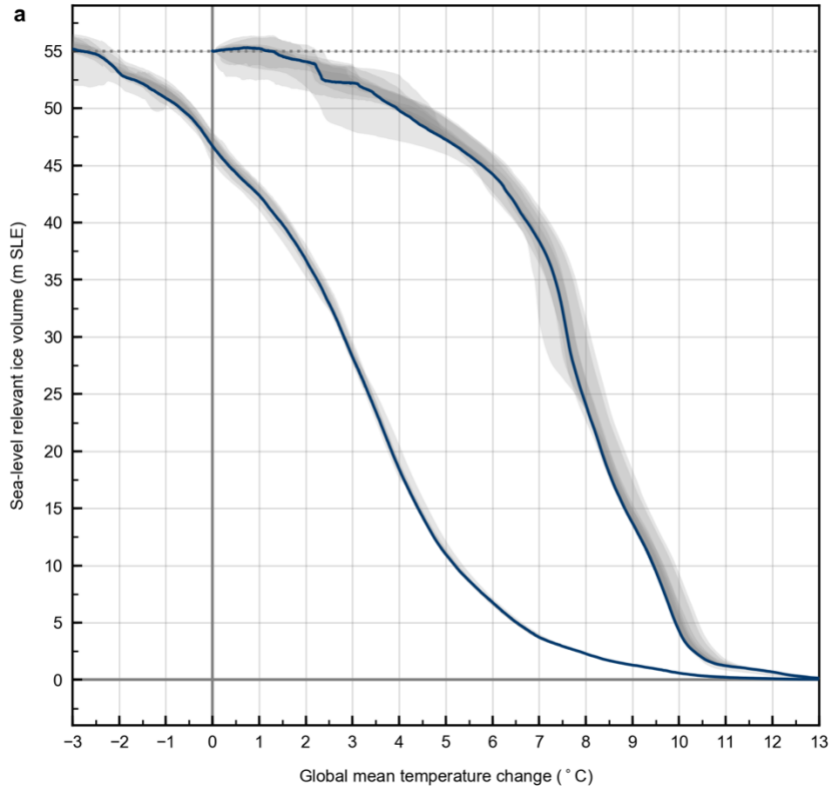
Extended Data Fig. 2: Comparison of modelled and observed ice velocities.

(a) Observed ice surface velocities⁶⁷ (in metres per year), regridded to 16 km. (b) Modelled ice surface velocities (in metres per year) of reference state. (c) Model minus observed ice surface velocity (in metres per year). (d) Scatter plot of model versus observed ice surface velocity (in metres per year). The red line is the identity line. RMSE = root-mean-square error.



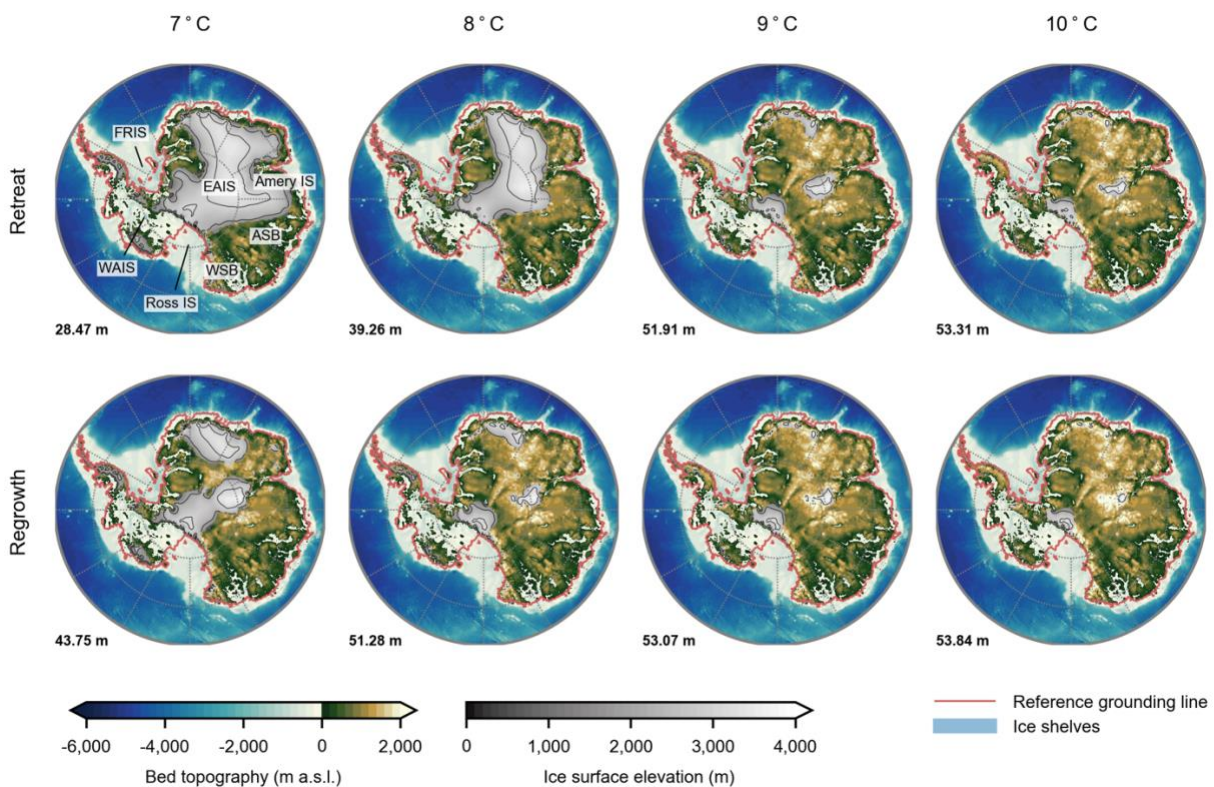
Extended Data Fig. 3: Regrown Antarctica.

Equilibrium ice thickness difference (in metres) of the regrown ice-sheet configuration (at pre-industrial temperatures, i.e. 0 °C GMT anomaly) minus the reference ice-sheet thickness. Red lines denote the reference grounding-line locations, blue lines show their respective locations after regrowth. Areas of the ice sheet which do not regrow to their original extents are highlighted in orange. Surface-height contour lines of the regrown ice sheet are given at 200 m intervals. AP = Antarctic Peninsula.



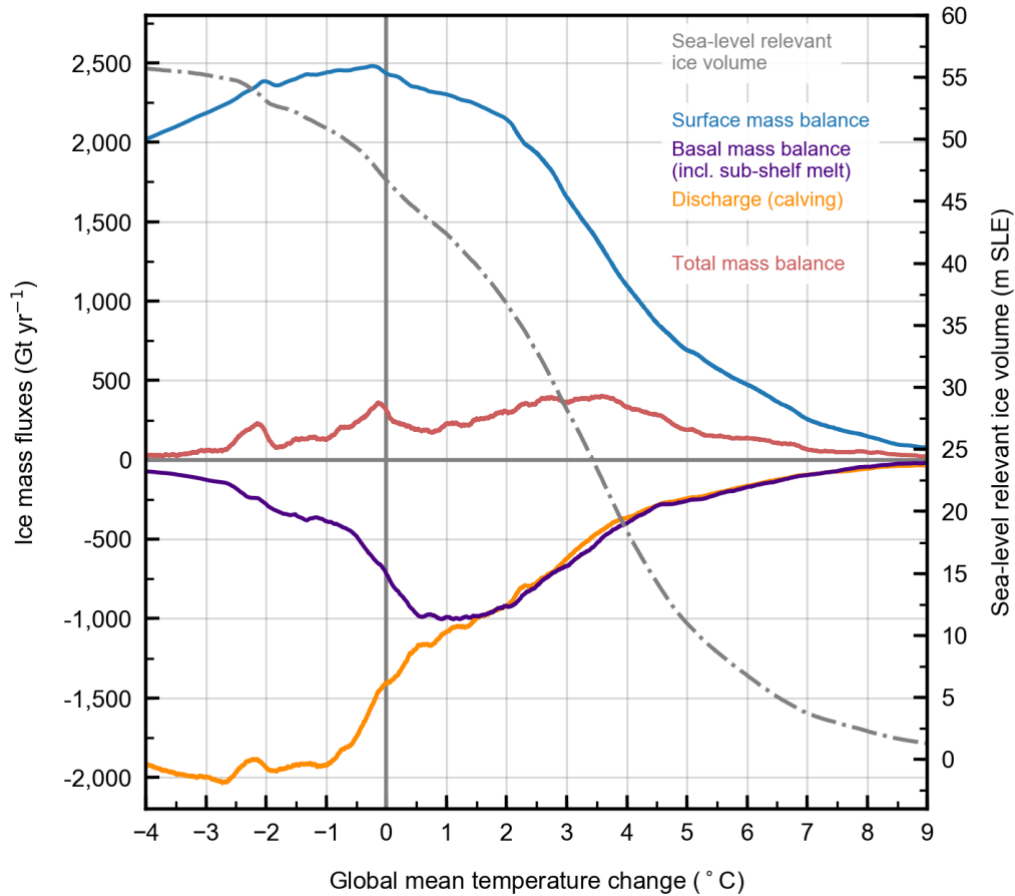
Extended Data Fig. 4: Hysteresis sensitivity to model parameter variations.

(a) Sea-level relevant ice volume (in metres sea-level equivalent, m SLE) for a global warming rate of $1\text{ }^{\circ}\text{C}$ per 10,000 years above pre-industrial conditions. The blue curve is the same as the blue curve in Figure 2; the grey shadings show the model sensitivity by encompassing the total range of individual model responses with respect to the variation of critical model parameters, as detailed below. (b-g) Same as (a), but showing the respective simulations for the tested model parameters (thin blue lines) in comparison to the reference simulation (thick blue line): (b) ice-shelf removal mechanism ($PD = \text{present-day}$), (c) viscosity of the upper mantle in the Earth-deformation model (in Pascal-seconds), (d) decay rate of the subglacial meltwater in the till layer (in millimetres per year), (e) exponent in the “pseudo-plastic” sliding law (unitless), (f) flow enhancement factor for the SSA velocities⁴ (unitless), (g) horizontal model grid resolution (in kilometres).



Extended Data Fig. 5: Long-term ice loss for different warming levels (cont.).

Shown is the equilibrium ice-sheet surface elevation in metres for different warming levels (7, 8, 9, and $10\text{ }^{\circ}\text{C}$ GMT anomaly above pre-industrial level), comparing the retreat (upper panels) and regrowth (lower panels) branch of the hysteresis curve. Ice surface-height contours are delineated at 1,000 m intervals. Grounding-line locations of the reference state are shown in red; ice shelves are marked in light blue. Absolute sea-level relevant ice-volume anomaly compared to the reference state (in metres sea-level equivalent, m SLE), i.e. the committed sea-level rise, is given for each panel. Blue shadings illustrate the bedrock depth below, brown shadings above present-day sea level (m a.s.l. = metres above sea level). EAIS = East Antarctic Ice Sheet, WAIS = West Antarctic Ice Sheet, IS = ice shelf, FRIS = Filchner–Ronne Ice Shelf, ASB = Aurora subglacial basin, WSB = Wilkes subglacial basin.



Extended Data Fig. 6: Ocean-driven vs. atmosphere-driven ice loss (regrowth branch).

Antarctic ice mass fluxes (in gigatonnes per year) showing the contributions of different atmospheric and oceanic processes to total ice mass changes over the entire range of GMT anomalies along the lower hysteresis branch derived from the quasi-static reference simulation. Positive flux values denote mass gains, negative values mass losses. The sea-level relevant ice-sheet volume (in metres sea-level equivalent, m SLE) is indicated by the dashed grey line with respect to the right-hand axis.

Supplementary Information

Supplementary Video 1: Ice-sheet retreat along the upper branch of the hysteresis.

External link: https://static-content.springer.com/esm/art%3A10.1038%2Fs41586-020-2727-5/MediaObjects/41586_2020_2727_MOESM1_ESM.mp4

Shown is the quasi-static evolution of the ice-sheet surface elevation in metres for a global warming rate of 1 °C per 10,000 years above pre-industrial conditions (grey shading; contour lines at 500 m intervals), sub-shelf melt rates in metres per year (purple–orange shading) as well as bedrock topography below (blue shading) and above (brown shading) present-day sea level (m a.s.l. = metres above sea level). EAIS = East Antarctic Ice Sheet, WAIS = West Antarctic Ice Sheet, IS = ice shelf, FRIS = Filchner–Ronne Ice Shelf.

Lower panel shows the ice volume change (blue curve, in metres sea-level equivalent, m SLE) as well as total Antarctic ice mass balance fluxes (purple curve, in gigatonnes per year).

Supplementary Video 2: Ice-sheet regrowth along the lower branch of the hysteresis.

External link: https://static-content.springer.com/esm/art%3A10.1038%2Fs41586-020-2727-5/MediaObjects/41586_2020_2727_MOESM2_ESM.mp4

Shown is the quasi-static evolution of the ice-sheet surface elevation in metres for a global cooling rate of -1 °C per 10,000 years starting from ice-free conditions (grey shading; contour lines at 500 m intervals), sub-shelf melt rates in metres per year (purple–orange shading) as well as bedrock topography below (blue shading) and above (brown shading) present-day sea level (m a.s.l. = metres above sea level). EAIS = East Antarctic Ice Sheet, WAIS = West Antarctic Ice Sheet, IS = ice shelf, FRIS = Filchner–Ronne Ice Shelf.

Lower panel shows the ice volume change (blue curve, in metres sea-level equivalent, m SLE) as well as total Antarctic ice mass balance fluxes (purple curve, in gigatonnes per year).