

The uncertain future of the Antarctic ice sheet

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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 (*I*). 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

19 potential to raise sea level by 5.3 m (2), has its current base grounded well below sea level and
20 the bed deepens from the periphery of the ice sheet towards the interior (a so-called retrograde
21 bed slope). Marine basins are also present in certain areas of the East Antarctic ice sheet (Fig. 1),
22 which has a far greater sea level potential of 52.2 m (2). Marine ice sheets are in direct contact
23 with the ocean under floating ice shelves around the coast, and changes in ocean circulation
24 or heat content may lead to rapid ice loss on timescales of decades to centuries. The uncer-
25 tainty in the timing and extent of potential tipping points also stems from our poor knowledge
26 of both drivers of change and mechanisms that operate in the dynamics of marine ice sheets.
27 Despite these shortcomings, multi-model comparisons like ISMIP6 allow for a more standard-
28 ised approach that enable outliers to be more clearly identified. Hence, uncertainties in future
29 projections have since been reduced and more robust projections of sea level contributions from
30 the Antarctic ice sheet are to be expected.

31 **Observations and drivers of dynamical mass change**

32 Recent satellite observations indicate that the contribution of the Antarctic ice sheet to sea-
33 level rise has significantly increased in recent years (3). Antarctica has been contributing 0.15–
34 0.46 mm yr⁻¹ to sea level on average between 1992 and 2017, accelerating to 0.49–0.73 mm
35 yr⁻¹ between 2012 and 2017 (4). Most ice loss is concentrated in West Antarctica, where the
36 thinning of floating ice shelves is causing glacier flow to accelerate and grounding lines (the
37 contact between the grounded ice sheet and the ice shelf floating on the ocean) to retreat.

38 The acceleration and thinning of Pine Island Glacier, Thwaites Glacier, and nearby glaciers
39 draining into the Amundsen Sea (Fig. 2), which dominate the mass loss from the West Antarc-
40 tic Ice Sheet (WAIS), results from ice-shelf thinning and shrinkage, and associated grounding
41 line retreat. This is thought to be a response to a wind-driven increase in the circulation of
42 warm Circumpolar Deep Water (CDW) onto the continental shelf reaching ice shelf cavities

43 and grounding lines (5). The strengthening of the regional westerly winds that have forced
44 warmer waters to the grounding zones are attributed primarily to remote changes occurring in
45 the tropics (6). However, changes in larger-scale circulation owing to the recent stratospheric
46 cooling due to ozone depletion and increased concentration of greenhouse gases have also been
47 identified as potential drivers (7). Thwaites Glacier is today undergoing the largest changes of
48 any ice-ocean system in Antarctica (8). This ongoing mass loss will be modulated but likely not
49 reversed by variability in the ocean (9).

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

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

62 **Dynamics of the marine ice sheet**

63 The mass balance of the Antarctic ice sheet, and therefore its contribution to sea level, is deter-
64 mined by the balance between mass gain and mass loss. The ice sheet gains mass from snowfall
65 on its surface and loses mass primarily by ocean-induced melting beneath its floating ice shelves
66 along the coast, and by calving icebergs that drift away and melt in the ocean. While the surface

67 mass balance has been relatively stable over the past decades, ice flow in several sectors of the
68 ice sheet has accelerated, thereby increasing ice discharge. The dominant process triggering
69 these large, rapid changes is the loss of ice-shelf buttressing. This is initiated by changes in
70 ocean circulation and to a lesser extent atmospheric drivers that control summer surface-melt
71 rates (13, 14). In particular, the warmer waters of the CDW move toward the ice fronts and
72 ice-shelf grounding zones along troughs in the bathymetry, causing increased melting at the
73 ice-ocean interface. This process thins the ice shelves, reducing drag along their sides and at
74 local pinning points on sea-floor highs, which in turn reduces the buttressing i.e., the resistive
75 stress that the ice shelves exert on the grounded ice (8). Thinning ice shelves lead to faster
76 grounded-ice flow, which in turn leads to further thinning, causing previously grounded ice to
77 float as the grounding zone retreats farther inland. This process can be particularly fast and
78 unstable along retrograde slopes (i.e., the bed deepens inland), as more ice crossing the ground-
79 ing zone and a smaller accumulation area (15, 16) creates a positive feedback process known
80 as the marine ice sheet instability (MISI; Fig. 3). The process may halt when the bedrock rises
81 upward, i.e., when a prograde bed slope or pronounced ridge at the bed is encountered, or when
82 ice shelves exert enough buttressing to stop further grounding-line retreat.

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

90 More recently, the hypothesis of Marine Ice Cliff Instability has emerged (MICI) (14, 21),
91 postulating that ice cliffs become unstable and collapse if higher than ~ 90 m above sea level,

92 facilitating the rapid retreat of ice sheets. This process may have been significant in Antarctica
93 during past warm periods (14) by enhancing MISI (Fig. 3). During Pliocene warm periods,
94 sea level was 10 to 20 m higher than present (22), requiring extensive retreat or collapse of
95 the Greenland, West Antarctic and marine-based sectors of the East Antarctic ice sheets. The
96 MICI mechanism enables to increase the model sensitivity to reach such high sea-level stands
97 during that period (14). However, contrary to the MISI hypothesis, MICI is not supported by
98 a formal linear stability analysis (16), which hampers an adequate representation in marine
99 ice sheet models. Furthermore, MICI has not been observed at such a scale in Antarctica and
100 so it remains unclear how rapid an ice cliff would retreat as a function of its height (23). So
101 far, models including MICI parameterized the rate of retreat based on observed retreat rate of
102 Jakobshavn Isbræ in West Greenland, which reached 3 km yr^{-1} when its ice shelf collapsed in
103 the early 2000s.

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

117 **Projecting the future of the Antarctic ice sheet**

118 A major factor that limits reliable projections of the future Antarctic ice sheet response is how
119 global warming relates to ocean dynamics that bring CDW onto and across the continental shelf,
120 potentially increasing sub-shelf melt. Because of this uncertainty, several studies apply lin-
121 ear extrapolations of present-day observed melt rates or simple parameterizations of ice-ocean
122 melting rates, mostly focusing on unmitigated climate scenarios, such as RCP8.5. Numerous
123 large-scale modelling studies conducted in the last decade have simulated future collapse of
124 WAIS under various climate-warming scenarios (*13, 14, 27–30*). These studies find that future
125 grounding-zone retreat into the central WAIS region is expected on timescales of a few centuries
126 to a millennium, contributing several meters to global mean sea level rise. However, while the
127 time of onset of collapse is quite different across models and scenarios, all models produce
128 WAIS collapse under unmitigated emission scenarios on multi-centennial timescales.

129 Whole Antarctic simulations for unmitigated emission scenarios (RCP8.5) show a large
130 scatter on centennial and multi-centennial timescales (Fig. 4). However, the introduction of
131 MICI in one ice-sheet model (*14*) results in future sea-level rise estimates of almost one order
132 of magnitude larger compared to other studies (Fig. 4). While projected contributions of the
133 Antarctic ice sheets to sea-level rise by the end of this century for recent studies hover between
134 0 and 0.45 m (5%–95% probability range), the MICI model occupies a range of 0.2–1.7 m
135 (Fig. 4). The discrepancy is even more pronounced for 2300, where the MICI results and other
136 model estimates no longer agree within uncertainty bounds. Given the uncertainty range on
137 Pliocene sea-level stands, MICI is not necessarily required to lead to rapid multi-meter sea level
138 rise (*31*) and other mechanisms related to basal conditions may well be able to accelerate mass
139 loss on shorter timescales (*30, 32*).

140 Not all feedbacks in marine ice sheets enhance ice loss and collapse. Several mechanisms

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

154 **Sea-level commitment and tipping points**

155 On multacentennial to multimillennial timescales, feedbacks with the atmosphere and ocean
156 increase in importance. When subjected to perturbed climatic forcing over these timescales, ice
157 sheets manifest large changes in their volume and distribution. These changes typically occur
158 with a significant lag in response to the forcing applied, which leads to the concept of sea-level
159 commitment, i.e., ice mass losses that will occur in the long-term future are committed to that
160 loss at a much earlier stage. Ice sheets are subject to threshold behaviours in their stability,
161 as a change in boundary conditions such as climate forcing can cause the current ice-sheet
162 configuration to become unstable through, for instance, MISI. Crossing these tipping points
163 leads the system to equilibrate to a qualitatively different state (a complete collapse of WAIS,
164 for example). The existence of a tipping point implies that ice-sheet changes are potentially

165 irreversible. In other words, returning to a pre-industrial climate may not necessarily stabilize
166 the ice sheet once the tipping point has been crossed. Reversibility, however, may be possible
167 over large climate cycles, such as a glacial-interglacial cycle.

168 The projected long-term sea level rise contribution of the Antarctic ice sheet for warming
169 levels associated with the high-mitigation RCP2.6 scenario is limited to well below one metre,
170 although with a probability distribution that is not Gaussian, but skewed with a long tail towards
171 high values due to potential MICI (*1*). However, substantial future retreat in some basins (such
172 as Thwaites Glacier) cannot be ruled out, as grounding-line retreat may continue even with no
173 additional forcing (*18–20, 32*). The long-term sea level rise contribution of the Antarctic ice
174 sheet therefore crucially depends on the behaviour of individual ice shelves and outlet glacier
175 systems and whether they enter MISI for a given level of warming. Under sustained warming,
176 a threshold for the survival of Antarctic ice shelves, and thus the stability of the ice sheet,
177 seems to lie between 1.5 and 2°C mean annual air temperature above present (*28*). Crossing
178 these thresholds implies commitment to large ice-sheet changes and sea level rise that may take
179 thousands of years to be fully realized and be irreversible on longer timescales (*1*).

180 **Understanding key physical processes**

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

189 points of the Antarctic ice sheet.

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

203 Iceberg calving (ii) is responsible for the other part of the ice mass loss at the margins of
204 the Antarctic ice sheet. Calving occurs when ice chunks break off from the edge of floating
205 ice shelves in Antarctica. The rate at which icebergs detach from the ice shelf, or calving rate,
206 determines the dynamics of the ice front. When the ice front is stationery, the calving rate is
207 equal to the flow velocity of the ice. The calving rate therefore modulates buttressing induced
208 by ice shelves and hence indirectly controls upstream grounded ice speed and subsequent sea
209 level rise contribution. The large amount of ice lost through calving is common for Antarctica,
210 but its representation and quantification in models is hampered by the difficult access to field
211 sites, a high variability in time and space, and its inherent discontinuous nature, as opposed
212 to the continuum approach used in most models. Until recently, calving rates were essentially
213 either assumed to be equal to ice velocity (i.e. by keeping ice front fixed in space) or based

214 on empirical relationships that are not well constrained by observations. Recent studies apply
215 continuum damage mechanics to simulate crevasse formation. This approach represents initial
216 ice microfractures and their vertical development as crevasses, which in turn weakens the ice
217 through damage and decreases ice viscosity, and which can be advected with the ice flow (36).
218 Hydrofracturing, based on the surface meltwater widening and deepening crevasses is also ubiq-
219 uitously parameterized in ice sheet models, and forms the precursor for MICI (21, 24). Calving
220 remains one of the grand challenges of ice sheet modelling and no general calving law exists
221 yet, which profoundly limits our ability to model catastrophic calving events.

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

232 **Challenges to reduce uncertainties**

233 Besides understanding of key physical processes, their representation in ice-sheet models are
234 also crucial. One way to assess the accuracy in the representation of physical processes in
235 current ice sheet models is to organize large, international intercomparison projects. For ex-
236 ample, the Marine Ice Sheet Model Intercomparison Project for planview models (MISMIP3d)
237 greatly improved the representation grounding-line migration by conforming models to known

238 analytical solutions (38). These numerical experiments demonstrated that in order to resolve
239 grounding-line migration in marine ice-sheet models, a sufficiently high spatial resolution needs
240 to be adopted, since membrane stresses need to be resolved across the grounding line to guar-
241 antee mechanical coupling, unless parameterizations are used (14) based on analytical solu-
242 tions (16). Therefore, a series of ice-sheet models have implemented sub-element parameteriza-
243 tions or a spatial grid refinement, which also favours accurate data assimilation (27). In transient
244 simulations the adaptive mesh approach enables the finest grid to follow the grounding-line mi-
245 gration (27). These higher spatial resolutions on the order of hundreds of meters in the vicinity
246 of grounding lines also pose new challenges about data management for modelling purposes
247 and demand precise bathymetry to resolve the grounding zone (2). Nevertheless, recent the-
248 oretical developments with respect to grounding-line stability in response to buttressing (39),
249 basal drag (40) and external forcing (41) demonstrate that further efforts are required in the
250 verification and validation of numerical ice sheet models.

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

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

263 where progress should be made to reduce the uncertainty in projections of sea level rise most
264 efficiently.

265 Model initialization remains another important factor, which relies on two distinct, but of-
266 ten combined approaches: spin-up versus data assimilation. The first approach spins up the
267 model over glacial-interglacial periods, which ensures that the internal properties of the ice sheet
268 are consistent with each other but may provide an inaccurate representation of the present-day
269 ice sheet geometry and flow speed, which may introduce significant biases on short term (i.e.
270 decadal to centennial) projections. The alternative is the assimilation of data, such as satellite
271 derived surface flow speeds, thinning/thickening rates, etc. These two approaches lead to large
272 differences in the initial conditions from which projections are made and therefore create a sig-
273 nificant spread in projected contributions to future sea level rise (43). While data assimilation
274 techniques cannot ensure consistent internal properties of the ice sheet, they are improving for
275 centennial projections with the increasing access to high-resolution satellite products, which
276 even allow for characterizing the subglacial conditions to a far better degree (44). They also
277 enable to improve ice thickness and bedrock data sets at a high resolution for the Antarctic ice
278 sheet (2). One of the challenges for the coming years is that the volume of data available is in-
279 creasing exponentially, while ice sheet models are not equipped to ingest large amount of data
280 from different sensors at different resolutions and acquired at different times. Some progress
281 has been made by relying on tools such as automatic differentiation, but these methods have not
282 yet been applied to large scale systems such as the entire Antarctic ice sheet.

283 Eventually, the full coupling between ice, ocean and atmosphere must be considered, which
284 is currently the subject of ongoing research, but remains limited to decadal or multi-decadal
285 timescales due to the high computational cost of coupled models. Full ice-ocean coupling on the
286 Thwaites drainage basin revealed a continued mass loss over the coming decades at a sustained
287 rate and show that uncoupled simulations significantly overestimate the rate of grounding-line

288 retreat compared to the coupled model (20). Whole Antarctic semi-coupled simulations, on the
289 other hand, show that meltwater from Antarctica will trap warm water below the sea surface,
290 creating a positive feedback that increases Antarctic ice loss (32).

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

300 **References**

- 301 1. F. Pattyn, *et al.*, *Nature Climate Change* **8**, 1053 (2018).
- 302 2. M. Morlighem, *et al.*, *Nature Geoscience* (2019).
- 303 3. E. Rignot, *et al.*, *Proceedings of the National Academy of Sciences* **116**, 1095 (2019).
- 304 4. A. Shepherd, *et al.*, *Nature* **558**, 219 (2018).
- 305 5. S. Schmidtko, K. J. Heywood, A. F. Thompson, S. Aoki, *Science* **346**, 1227 (2014).
- 306 6. P. Dutrieux, *et al.*, *Science* (80-.). **343** (2014).
- 307 7. D. W. J. Thompson, *et al.*, *Nature Geoscience* **4**, 741 (2011).
- 308 8. F. S. Paolo, H. A. Fricker, L. Padman, *Science* (80-.). **348**, 327 (2015).

- 309 9. K. Christianson, *et al.*, *Geophys. Res. Lett.* **43**, 10,817 (2016).
- 310 10. S. R. Rintoul, *et al.*, *Science Advances* **2** (2016).
- 311 11. B. Medley, E. R. Thomas, *Nature Climate Change* **9**, 34 (2019).
- 312 12. J. Turner, *et al.*, *Geophysical Research Letters* **46**, 3502 (2019).
- 313 13. J. Feldmann, A. Levermann, *Proceedings of the National Academy of Sciences* **112**, 14191
314 (2015).
- 315 14. R. M. DeConto, D. Pollard, *Nature* **531**, 591 (2016).
- 316 15. J. Weertman, *J. Glaciol.* **13**, 3 (1974).
- 317 16. C. Schoof, *Journal of Geophysical Research: Earth Surface* **112**, n/a (2007). F03S28.
- 318 17. L. Favier, *et al.*, *Nature Climate Change* **4**, 117 (2014).
- 319 18. I. Joughin, B. E. Smith, B. Medley, *Science* **344**, 735 (2014).
- 320 19. I. J. Nias, S. L. Cornford, A. J. Payne, *J. Glaciol.* **62**, 552 (2016).
- 321 20. H. Seroussi, *et al.*, *Geophys. Res. Lett.* pp. n/a–n/a (2017).
- 322 21. D. Pollard, R. M. DeConto, R. B. Alley, *Earth and Planetary Science Letters* **412**, 112
323 (2015).
- 324 22. G. R. Grant, *et al.*, *Nature* **574**, 237 (2019).
- 325 23. F. Clerc, B. M. Minchew, M. D. Behn, *Geophysical Research Letters* **46**, 12108 (2019).
- 326 24. J. N. Bassis, C. C. Walker, *Proceedings of the Royal Society A: Mathematical, Physical and*
327 *Engineering Sciences* **468**, 913 (2012).

- 328 25. A. A. Robel, A. F. Banwell, *Geophysical Research Letters* **46**, 12092 (2019).
- 329 26. P. J. Bart, M. DeCesare, B. E. Rosenheim, W. Majewski, A. McGlannan, *Scientific Reports*
330 **8**, 12392 (2018).
- 331 27. S. L. Cornford, *et al.*, *The Cryosphere* **9**, 1579 (2015).
- 332 28. N. R. Golledge, *et al.*, *Nature* **526**, 421 (2015).
- 333 29. C. Ritz, *et al.*, *Nature* **528**, 115 (2015).
- 334 30. K. Bulthuis, M. Arnst, S. Sun, F. Pattyn, *The Cryosphere* **13**, 1349 (2019).
- 335 31. T. L. Edwards, *et al.*, *Nature* **566**, 58 (2019).
- 336 32. N. R. Golledge, *et al.*, *Nature* **566**, 65 (2019).
- 337 33. V. R. Barletta, *et al.*, *Science* **1339**, 1335 (2018).
- 338 34. E. Larour, *et al.*, *Science* **7908** (2019).
- 339 35. L. Favier, *et al.*, *Geoscientific Model Development* **12**, 2255 (2019).
- 340 36. S. Sun, S. L. Cornford, J. C. Moore, R. Gladstone, L. Zhao, *The Cryosphere* **11**, 2543
341 (2017).
- 342 37. J. Brondex, F. Gillet-Chaulet, O. Gagliardini, *The Cryosphere* **13**, 177 (2019).
- 343 38. F. Pattyn, *et al.*, *Journal of Glaciology* **59**, 410 (2013).
- 344 39. M. Haseloff, O. V. Sergienko, *Journal of Glaciology* **64**, 417–431 (2018).
- 345 40. O. V. Sergienko, D. J. Wingham, *Journal of Glaciology* **65**, 833–849 (2019).

- 346 41. G. H. Gudmundsson, F. S. Paolo, S. Adusumilli, H. A. Fricker, *Geophysical Research*
347 *Letters* **46**, 13903 (2019).
- 348 42. S. M. J. Nowicki, *et al.*, *Geoscientific Model Development* **9**, 4521 (2016).
- 349 43. H. Seroussi, *et al.*, *The Cryosphere* **13**, 1441 (2019).
- 350 44. F. Gillet-Chaulet, *et al.*, *Geophysical Research Letters* **43**, 10,311 (2016).
- 351 45. H. D. Pritchard, R. J. Arthern, D. G. Vaughan, L. A. Edwards, *Nature* **461**, 971 (2009).
- 352 46. E. Hanna, *et al.*, *Earth-Science Reviews* **201**, 102976 (2020).
- 353 47. A. Levermann, *et al.*, *Earth Syst. Dyn.* **5**, 271 (2014).

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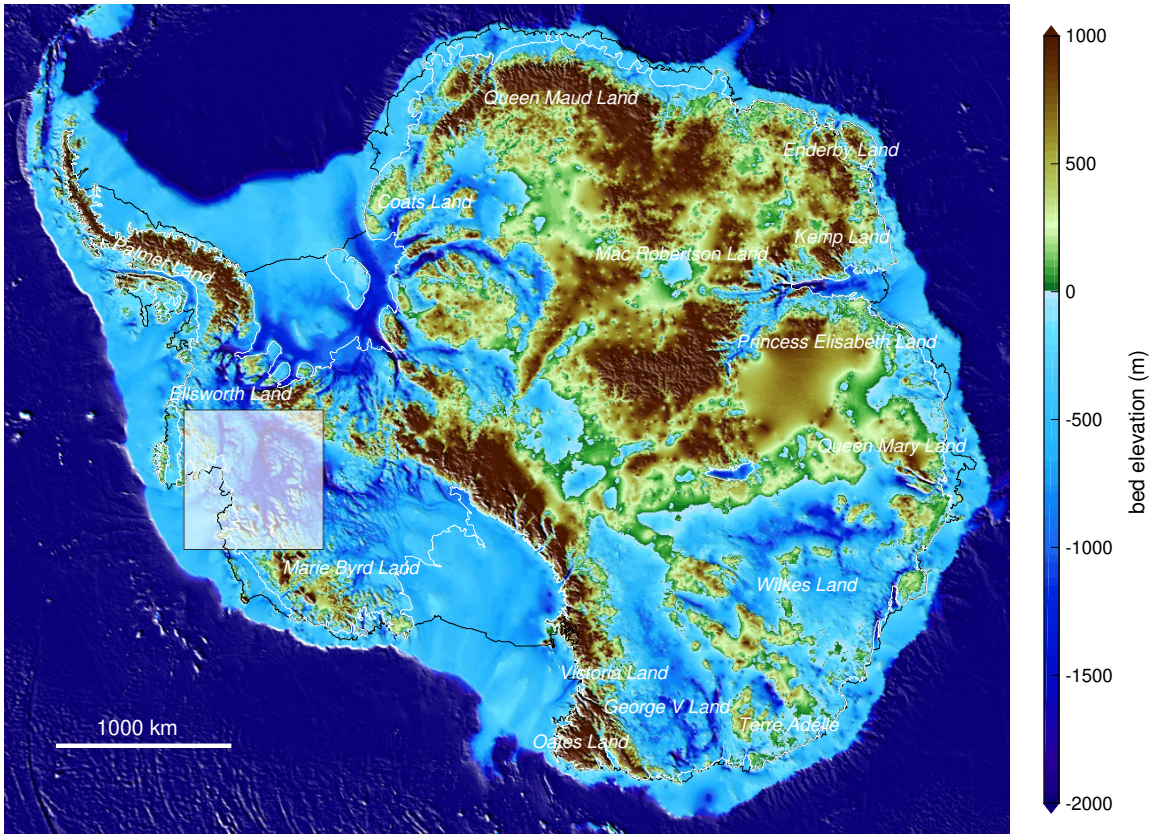


Figure 1: Bed topography (bathymetry) of Antarctica (2). Blue areas are marine based (below sea level). The ice-sheet grounding line is plotted in white and ice front in black. The greyed area indicates the Amundsen Sea Embayment, shown in Figure 2.

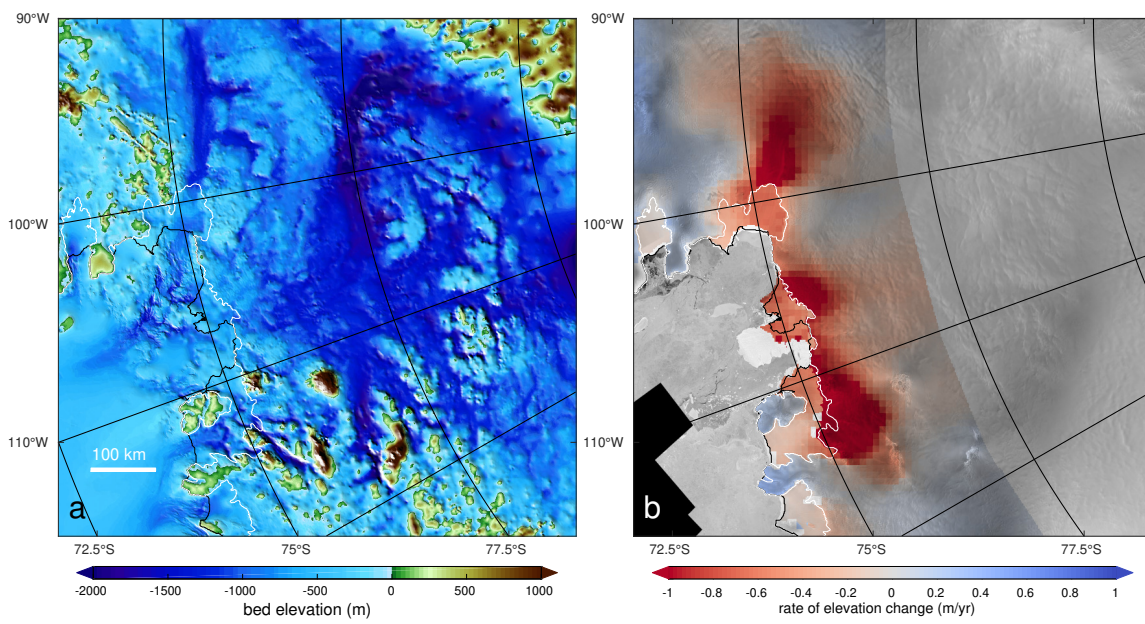


Figure 2: (a) Bed topography (bathymetry) of the Amundsen Sea Embayment (2) and (b) rate of ice-sheet elevation change (2003–2009) from ICESat GLAS laser altimetry (45).

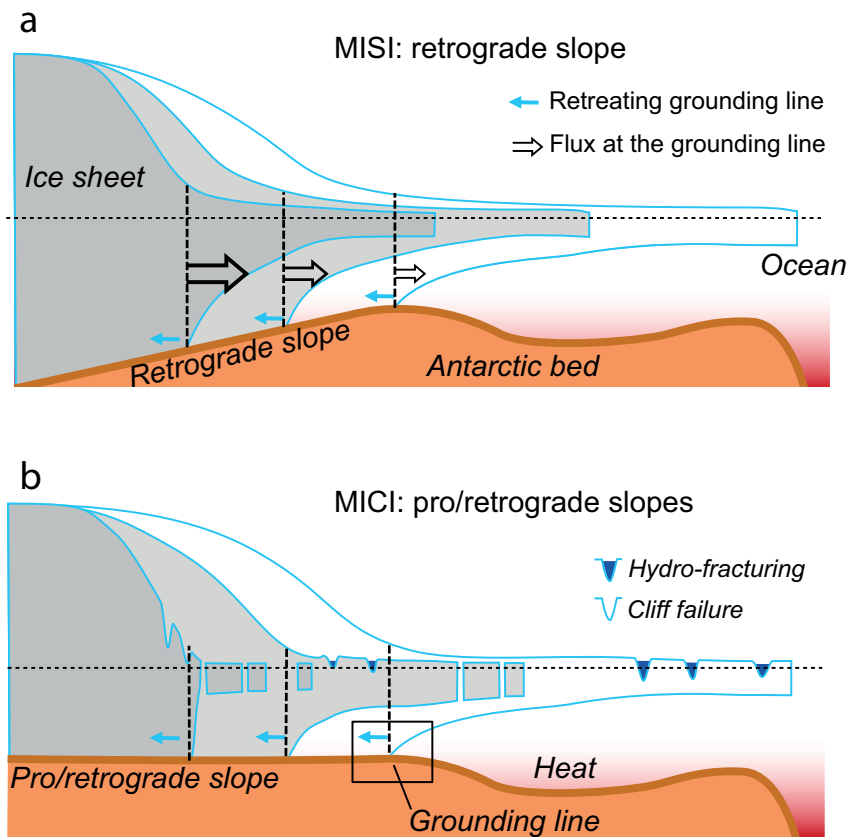


Figure 3: Schematic of the Marine Ice Sheet Instability (MISI) and the Marine Ice Cliff Instability (MICI). Redrawn after (1).

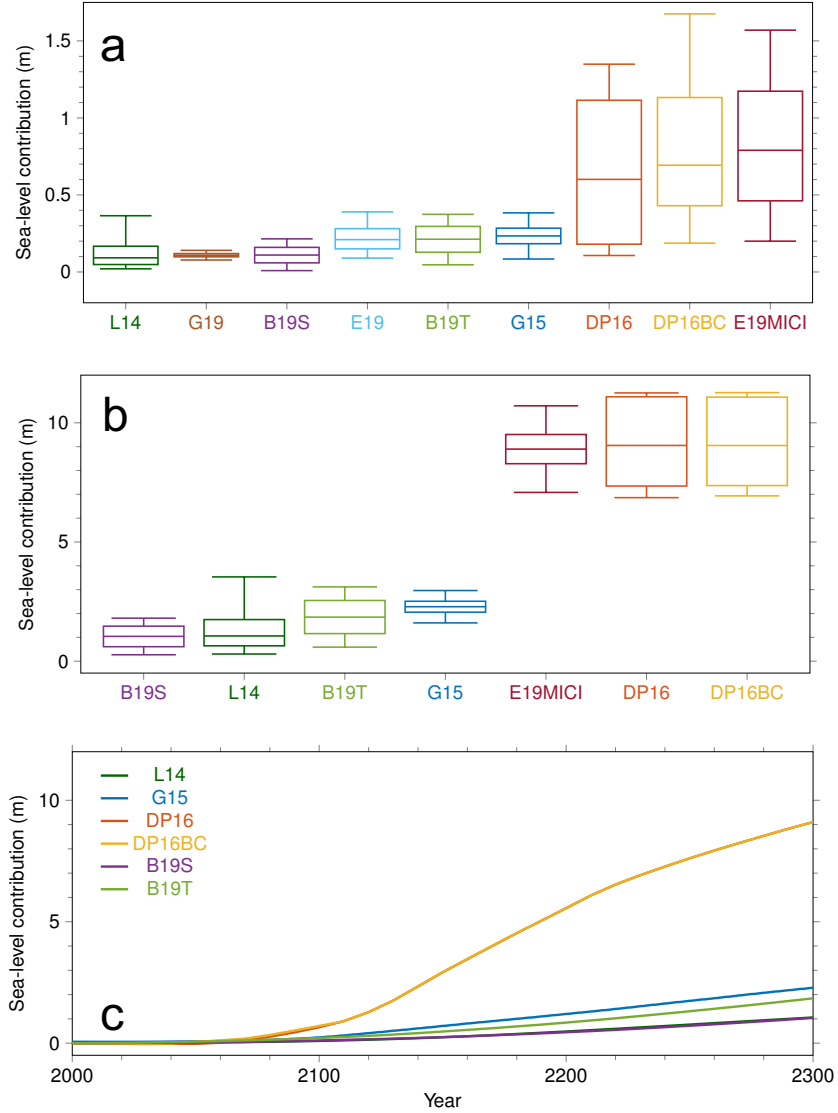


Figure 4: Projections of Antarctic sea-level contribution at (a) 2100 and (b) 2300 under RCP8.5 (46). Boxes and whiskers show the 5th, 25th, 50th, 75th and 95th percentiles. (c) Median projections of Antarctic sea-level contribution until 2300 (RCP8.5). Colour legend: L14 (47), G15 (28), DP16 (14), DP16BC: 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).