

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

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25 **Even if anthropogenic warming were constrained to less than 2 °C above**
26 **pre-industrial, the Greenland and Antarctic ice sheets will continue to lose mass**
27 **this century, with rates similar to those observed over the last decade. However,**
28 **nonlinear responses cannot be excluded, which may lead to larger rates of mass**
29 **loss. Furthermore, large uncertainties in future projections still remain, pertaining**
30 **to knowledge gaps in atmospheric (Greenland) and oceanic (Antarctica) forcing.**
31 **On millennial time scales, both ice sheets have tipping points at or slightly above the**
32 **1.5-2.0 °C threshold; for Greenland, this may lead to irreversible mass loss due to**
33 **the surface mass balance-elevation feedback, while for Antarctica, this could result**
34 **in a collapse of major drainage basins due to ice-shelf weakening.**

35 Projecting future sea-level rise (SLR, Box 1) is primarily hampered by our
36 incomplete knowledge of the contributions of the Greenland and the Antarctic Ice Sheets
37 (GrIS and AIS, respectively), Earth's largest ice masses. In this paper we review the
38 potential contribution of both ice sheets under a strongly mitigated climate change
39 scenario that limits the rise in global near-surface temperature to less than 2 °C above
40 pre-industrial (targeting 1.5 °C), as agreed at the 21st UNFCCC climate conference in
41 Paris. We base the review on both present-day observed/modelled changes and future
42 forcings according to the RCP2.6 scenario. We use RCP2.6, the most conservative of the

43 four Representative Concentration Pathways of greenhouse gas concentration trajectories
44 adopted by the IPCC for its Fifth Assessment Report, because it is the scenario in the
45 published literature that best approximates to the above warming range. Ice-sheet mass
46 balance is defined as the net result of all mass gains and losses, and surface mass balance
47 (SMB) as the net mass balance at the ice-sheet surface (where a negative mass balance
48 means mass loss), including the firn layer. Hence, SMB does not include dynamical
49 mass loss associated with ice flow at the ice-sheet margin or melting at the ice-ocean
50 interface. Increased ice flow accounts for about one third of the recent GrIS mass loss¹.
51 For Antarctica, where mass lost through ice discharge past the grounding line (the limit
52 between the grounded ice sheet and floating ice shelf) is roughly evenly shared between
53 oceanic basal melt before reaching the ice front and iceberg calving, increased ice flow
54 accounts for all of the recent mass loss^{2,3}.

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

61 **Greenland forcing and mass-balance changes**

62 Greenland has warmed by ~ 5 °C in winter and ~ 2 °C in summer since the mid-1990s⁴,
63 which is more than double the global mean warming rate in that period. The GrIS has

64 also been losing mass at an increasing rate since the 1990s⁵ with a 0.65-0.73 mm a⁻¹
65 mean sea-level rise equivalent (sle) for 2012-2016⁶. Since 2000, both SMB decrease and
66 ice discharge increase contributed to mass loss, but the relative contribution of SMB
67 decrease to the total mass loss went up from 42% to 68% between 2000 and 2012¹. The
68 current observed SMB decrease is mainly driven by increased melt and subsequent
69 runoff⁷ and is in part attributed to anthropogenic global warming and concurrent Arctic
70 Amplification (exacerbated Arctic warming due to regional feedbacks of global
71 warming), but also to recent atmospheric circulation changes in summer observed since
72 the 2000's⁸. The occurrence of a negative North Atlantic Oscillation (NAO) and a
73 concurrent positive phase of the East Atlantic Pattern since 2000 can be interpreted as a
74 weakening and southward displacement of the jet stream^{9,10}, allowing for anomalous
75 high pressure⁸ and enhanced atmospheric blocking¹¹ over the GrIS. These circulation
76 changes in summer have favoured the advection of warm southerly air masses¹² and
77 increased incoming solar radiation¹³, leading to more melt, which is further enhanced by
78 the melt-albedo feedback. The relative contribution of global warming and natural
79 climate variability to the recent atmospheric circulation changes of Greenland remains
80 an open question¹⁴. However, the CMIP5 (Coupled Model Intercomparison Project
81 Phase 5) models do not exhibit such circulation changes, either in future warming
82 scenarios or in present-day simulations¹². This explains why the recent observed SMB is
83 lower and runoff is higher than predicted by these models (Fig. 1a,b).

84 That climate models have limited skill in representing future changes in the North
85 Atlantic jet stream⁹ also affects how well clouds and precipitation over Greenland are
86 simulated in future scenarios. The general relation between precipitation and temperature

87 (+5% K⁻¹) derived using CMIP5 future projections¹² is subject to modification by
88 structural changes in the North Atlantic atmospheric polar jet-stream. Moreover, model
89 (mis-)representation of clouds has a major effect on projected melt and runoff¹⁵. In one
90 CMIP5-forced regional climate model, runoff depends linearly on temperature for
91 low-warming scenarios (Fig. 1b). In this model, runoff from the GrIS at the end of the
92 21st century is estimated at around 1 mm a⁻¹ sle (360 Gt a⁻¹) for the +1.5 °C scenario.
93 These end-of-century temperature and runoff values are close to what is currently
94 observed, which may be attributed to the recent circulation changes mentioned above.

95 A decrease in SMB lowers the ice sheet surface, which in turn lowers SMB because
96 at lower elevations, near-surface air temperature is generally higher^{16,17}. Additional SMB
97 changes due to the SMB-surface-elevation feedback are small for limited warming: in a
98 coupled SMB-ice-dynamical simulation, the feedback contributes 11% to the GrIS
99 runoff rate in an RCP2.6 scenario, or ~3 mm of additional sea-level rise by 2100¹⁷.

100 Apart from SMB, changes in the discharge of ice from iceberg calving and melt from
101 the fronts of marine-terminating outlet glaciers have the potential to increase the rate at
102 which the GrIS contributes to future SLR and many of these processes are starting to be
103 included in state-of-the-art Greenland ice-sheet models¹⁸. Calving and frontal melt has
104 already led to ice front retreat along most of the GrIS and acceleration of
105 marine-terminating glaciers since about 2000¹⁹. GrIS discharge increased from 1960 to
106 2005 but stabilised thereafter, although with large interannual fluctuations^{1,20}. These
107 recent changes in discharge are thought to be linked in part to fluctuations in the North
108 Atlantic ocean circulation^{21,22}. There is evidence that the 1970s to early 2000s increase
109 in ice discharge, as measured through changes in iceberg numbers, is also closely related

110 to increasing runoff²⁰, for example through increased melting of ice fronts by upwelling
111 freshwater plumes and the filling and hydro-fracturing of crevasses²³.

112 Increased runoff, percolation of meltwater to the base of the ice sheet and subsequent
113 basal lubrication has also been proposed as a mechanism for general ice flow
114 acceleration in the ablation zone (the Zwally effect)²⁴, but has since been shown to result
115 in only moderate speedup at the beginning of the melt season, which can be counteracted
116 by the development of an efficient drainage system²⁵. Modelling studies indicate that on
117 decadal to centennial timescales, the Zwally effect has a very limited contribution to
118 global SLR^{26,27}.

119 Future SMB and discharge components of the mass budget cannot be separated
120 entirely because of the SMB-elevation feedback and, more importantly, due to
121 interaction between the two components as more negative SMB removes ice before it
122 can reach the marine margins^{27,28}. However, both these effects become more important
123 with stronger climate forcing and therefore remain limited for the low-emission scenario
124 considered here. Modelling studies indicate that the partitioning between mass losses
125 from SMB and ice discharge and their spatial distribution are likely to remain similar to
126 today^{17,27}, although these studies do not account for the full range of uncertainty
127 associated with outlet-glacier changes. However, given that recent SMB changes
128 dominate the recent GrIS mass loss¹⁴, the largest source of uncertainty in future SLR is
129 likely to be linked to SMB.

130 **Expected Greenland response**

131 Modelling studies of the GrIS, according to RCP2.6, report a large spread in ice-sheet
132 volume change of 14-78 mm sle by 2100^{17,27}, with uncertainty arising mainly from
133 differences between climate models. The largest discrepancies between different climate
134 projections and ice-sheet models occur over the fast-flowing outlet glaciers²⁹. Recent
135 advances in high-resolution model simulations³⁰ highlight the importance of bed
136 topography in controlling ice-front retreat for a given amount of ocean warming.
137 However, capturing the dynamics of outlet glaciers remains difficult for several reasons:
138 (i) outlet glacier flux is not always well determined due to the limited knowledge of the
139 subglacial topography³¹ despite the significant progress made through mass-conservation
140 algorithms³²; (ii) the impact of ocean temperature on ice discharge at the margin is
141 poorly constrained; (iii) understanding of iceberg calving remains limited³³, while such
142 mechanisms drive most of the dynamic changes of marine-terminating glaciers³⁴.

143 On longer timescales (Box 2), a tipping point (when the ice sheet enters a state of
144 irreversible mass loss and complete melting is initiated) exists as part of the coupled ice
145 sheet-atmospheric system. This consists of two inter-related feedback mechanisms: the
146 SMB-elevation feedback, as described above, and the melt-albedo feedback³⁵⁻³⁷. The
147 latter acts on the surface energy balance, by allowing more absorption of solar radiation
148 from a melting and darkening snow surface, or removal of all snow leading to a darker
149 ice surface. This feedback may be enhanced by ice-based biological processes, such as
150 the growth of algae³⁸. Thus, the activation of these feedbacks can lead to self-sustained
151 melting of the entire ice sheet, even if the anomalous climatic forcing is removed.

152 It is clear that if the tipping point is crossed, a complete disappearance of the GrIS

153 would occur on a multi-millennial time scale³⁹⁻⁴¹. However, further work is urgently
154 needed to diagnose how close the GrIS is to this tipping point. Fig. 2 shows results from
155 an ensemble of simulations using one model varying key parameters related to
156 precipitation changes and melt rates⁴⁰. Simulations were performed with slowly
157 increasing climatic forcing, allowing the ice sheet to maintain a state of
158 quasi-equilibrium. Each simulation in the ensemble reached a tipping point, when the ice
159 sheet could no longer sustain itself. Fig. 2a compares this equilibrium threshold with the
160 diagnosed SMB of the GrIS given its present-day distribution, which can roughly be
161 used as a proxy for stability. SMB is spatially inhomogeneous, however, with high
162 accumulation and melt rates in the south, and cold, desert-like conditions in the north.
163 These simulations show that the Northwest sector of the ice sheet is particularly sensitive
164 to small changes in SMB, given the relatively low accumulation rates and associated
165 slower flow of ice from inland as compared to the South. Thus, in this model, a negative
166 SMB in the Northwest sector is a good predictor for the estimated threshold for complete
167 melting of the ice sheet.

168 The 95% confidence interval for the regional summer temperature threshold leading
169 to GrIS decline ranges from 1.1-2.3 °C above pre-industrial, with a best estimate of 1.8
170 °C⁴⁰. This level of warming is well within the range of expected regional temperature
171 changes given global warming limited to 1.5 °C, as CMIP5 models predict that
172 Greenland near-surface air temperatures increase more than the global average and
173 current levels of summer warming already reach this limit. This means that the threshold
174 will likely be exceeded, even for aggressive anthropogenic carbon emissions reductions.
175 However, in some peak-and-decline scenarios of CO₂ levels, full retreat can probably be

176 avoided despite the threshold having been temporally crossed.

177 The committed SLR after 1000, 5000 or 15,000 years, i.e., how much the ice sheet
178 will melt for a given climatic perturbation today (assumed constant in time), increases
179 non-linearly for higher levels of warming (Fig. 2b). The lag in response implies that such
180 a retreat would be set in motion much sooner, on timescales of the order of decades to
181 centuries (see Box 2). Thus, crossing the limit of 1.5 °C global warming this century
182 may impose a commitment to much larger and possibly irreversible changes in the far
183 future^{40,41}.

184 **Antarctic forcing and mass-balance changes**

185 The AIS has been losing mass since the mid-1990s, contributing 0.15-0.46 mm a⁻¹ sle
186 on average between 1992 and 2017, accelerating to 0.49-0.73 mm a⁻¹ between 2012 and
187 2017⁴². Observations over the last five years show that mass loss mainly occurs in the
188 Antarctic Peninsula and West Antarctica (0.42-0.65 mm a⁻¹ sle), with no significant
189 contribution from East Antarctica (-0.01-0.16 mm a⁻¹ sle)⁴². The mass loss from the
190 West Antarctic Ice Sheet (WAIS) is primarily caused by the acceleration of outlet
191 glaciers in the Amundsen Sea Embayment (ASE), where the ice discharge of large outlet
192 glaciers like Pine Island and Thwaites Glaciers (PIG and TG, respectively) increased
193 threefold since the early 1990s⁴². However, this ASE mass loss is not a recent
194 phenomenon, as ocean sediment records indicate that PIG experienced grounding-line
195 retreat since approximately the 1940s⁴³.

196 Antarctic SMB is projected to increase under atmospheric warming, governed by

197 increased snowfall due to increased atmospheric saturation water vapour pressure, the
198 availability of more open coastal water, and changing cloud properties⁴⁴. Ice cores
199 suggest that on centennial time scales SMB has increased especially in the Antarctic
200 Peninsula, representing a net reduction in sea level of ~ 0.04 mm per decade since 1900
201 CE⁴⁵. According to CMIP5 model means for RCP2.6, increased snowfall mitigates SLR
202 by 19 mm by 2100 and by 22 mm if only those CMIP5-models are used that best capture
203 CloudSat-observed Antarctic snowfall rates⁴⁶. Under atmospheric warming, Antarctic
204 surface melt, estimated at ~ 0.3 mm a⁻¹ sle⁴⁷, is projected to increase approximately
205 twofold by 2050, independent of the RCP forcing scenario⁴⁸. Recent studies show that
206 meltwater in Antarctica can be displaced laterally in flow networks⁴⁹, and sometimes
207 even enters the ocean⁵⁰. However, further research is needed to assess whether these
208 processes can challenge the present view that almost all surface meltwater refreezes in
209 the cold firn⁴⁷.

210 Major ice loss from the Antarctic ice sheet stems from an increased discharge of
211 grounded ice into the ocean, with ice shelves (the floating extensions of the grounded ice
212 sheet) playing a crucial role. The buttressing provided by ice shelves can affect inland
213 ice hundreds of kilometres away⁵¹, and hence controls grounding-line retreat and
214 associated ice flow acceleration. Ice shelves are directly affected by oceanic and
215 atmospheric conditions, and any change in these conditions may alter their buttressing
216 effect and impact the glaciers feeding them. For instance, increased sub-shelf melting
217 causes ice shelves to thin, increasing their sensitivity to mechanical weakening and
218 fracturing. This causes changes in ice shelf rheology and reduces buttressing of the
219 inland ice, leading to increased ice discharge⁵². Warming of the atmosphere promotes

220 rainfall and surface melt on the ice shelves and cause hydrofracturing as water present at
221 the ice sheet surface propagates into crevasses^{53,54} or by tensile stresses induced by lake
222 drainage⁵⁵. Anomalously low sea ice cover and the associated increase in ocean swell
223 has also been identified as an important precursor of Antarctic Peninsula ice shelf
224 collapse⁵⁶. These mechanisms were likely involved in the rapid breakup of Larsen B ice
225 shelf in 2002⁵⁵. While ice cores show that surface melting in the Antarctic Peninsula is
226 currently larger than ever recorded in recent history⁵⁷, for low emission scenarios, the
227 presence of significant rainfall and surface runoff is unlikely to spread far south of the
228 Antarctic Peninsula by 2100^{48,54}. Assessment of future surface melt-induced ice-shelf
229 collapse is therefore highly uncertain for mitigated scenarios, with largely diverging
230 estimates in recent literature. Parts of Larsen C, George VI, and Abbot ice shelves may
231 become susceptible to hydrofracturing by 2100 under a mitigated climate scenario⁵⁴, but
232 most studies identify significant potential ice-shelf collapse by 2100 only under
233 unmitigated scenarios^{48,58}.

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

240 However, the link between CDW upwelling and global climate change is not yet
241 clearly demonstrated, and decadal variability, such as El Niño/Southern Oscillation
242 (ENSO), may dominate ice-shelf mass variability in this sector⁶³. This variability may

243 increase as interannual atmospheric variability increases in a warming climate⁶³. The
244 CMIP5 ensemble also shows a modest mean warming of Antarctic Shelf Bottom Water
245 (ASBW), the ocean water masses occupying the sea floor on the Antarctic continental
246 shelf that provide the heat for basal melting of Antarctic ice shelves, of 0.25 ± 0.5 °C by
247 2100 under RCP2.6⁶⁴. Given that present-day biases in ASBW in CMIP5 models are of
248 the same order or larger than this warming and that the main limitation is the ability of
249 these models to resolve significant features in both bedrock topography and the ocean
250 flow⁶⁵, RCP2.6 projections of future sub-ice shelf melt remain poorly constrained⁶⁴.
251 Moreover, the link between increased presence of warm deep water on the continental
252 shelf and higher basal melt rates is not always clear; simulations of strengthened
253 westerly winds near the western Antarctic Peninsula showed an increase in warm deep
254 water on the continental shelf but a coincident decrease in ice-shelf basal melt⁶⁶.

255 Increasing the wind forcing over the Antarctic Circumpolar Current (ACC) has been
256 shown to have little effect on ice shelf basal melting⁶⁷. Ocean-sea ice projections that
257 include ice-shelf cavities have indicated the possibility that significant amounts of warm
258 deep water could gain access to the Filchner-Ronne ice-shelf cavities in the coming
259 century, increasing melt rates by as much as two orders of magnitude^{68,69}. This process
260 was seen with forcing from only one of two CMIP3 models and was more dependent on
261 the model that produced the forcing than on the emissions scenario⁶⁹, suggesting that
262 this scenario has a low probability.

263 Reduction of buttressing of ice shelves via the processes described above may
264 eventually lead to the so-called Marine Ice Sheet Instability (MISI; Fig. 3). For WAIS,
265 where the bedrock lies below sea level and slopes down towards the interior of the ice

266 sheet, MISI may lead to a (partial) collapse of this marine ice sheet. This process, first
267 hypothesized in the 1970's, was recently theoretically confirmed⁷⁰ and demonstrated in
268 numerical models⁷¹. It arises from thinning and eventually flotation of the ice near the
269 grounding line, which moves the latter into deeper water where the ice is thicker.
270 Thicker ice results in increased ice flux, which further thins (and eventually floats) the
271 ice, which results in further retreat into deeper water (and thicker ice), and so on. The
272 possibility that some glaciers, such as PIG and TG, are already undergoing MISI has
273 been suggested by numerical simulations using state-of-the-art ice sheet models^{72,73}. The
274 past retreat (up to 2010) of PIG has been attributed to MISI^{72,74} triggered by oceanic
275 forcing, although its recent slowdown may be due to a combination of abated forcing⁷⁵
276 and concomitant increase in glacier buttressing. TG is currently in a less buttressed state,
277 and several simulations using state-of-the-art ice sheet models indicate a continued mass
278 loss and possibly MISI even under present climatic conditions^{73,76,77}.

279 Additionally, evidence from the observed Larsen B collapse and rapid front retreat of
280 Jakobshavn Isbrae in Greenland, suggests that hydrofracturing could lead to rapid
281 collapse of ice shelves and potentially produce high ice cliffs with vertical exposure
282 above 90 m rendering the cliffs mechanically unsustainable, possibly resulting in what
283 has been termed Marine Ice Cliff Instability (MICI; Fig. 3)⁷⁸. This effect, if triggered by
284 a rapid disintegration of ice shelves due to hydrofracturing could lead to an acceleration
285 of ice discharge in Antarctica but is unlikely in a low emission scenario^{58,79}. However,
286 this process has not yet been observed in Antarctica, and may be prevented or delayed by
287 refreezing of meltwater in firn⁵⁴ or if efficient surface drainage exists⁵⁰.

288 **Expected Antarctic response**

289 A major limiting factor in projecting future Antarctic ice sheet response is how global
290 warming relates to ocean dynamics that bring CDW onto and across the continental
291 shelf, potentially increasing sub-shelf melt. Because of this uncertainty, several studies
292 apply linear extrapolations of present-day observed melt rates, while focusing on
293 unmitigated scenarios (RCP8.5). Mass loss according to mitigated scenarios are
294 essentially limited to dynamic losses in the Amundsen Sea Embayment of up to 0.05 m
295 sle by 2100. This is not much different than a linear extrapolation of the present-day
296 mass losses^{76,77,80} and in contrast with the observed acceleration of mass loss over the
297 last decade⁴². For the whole AIS, a mass loss between 0.01 and 0.1 m by 2100 is
298 projected according to RCP2.6⁸¹, which is not dissimilar (-0.11 to 0.15 m by 2100) from
299 model simulations based on Pliocene sea-level (5-15 m higher than today) tuning⁵⁸,
300 associated with a different melt parametrization at the grounding line (Fig. 4). Since the
301 value of sea level at the Pliocene is still debated⁸², tuning the model with a higher
302 Pliocene sea-level target (10-20 m) increases the model sensitivity, with an upper bound
303 of 0.22 m by 2100 according to the same scenario⁵⁸.

304 Because ocean heat supply is the crucial forcing for sub-shelf melting, oceanic
305 forcing has the potential to modulate the retreat rate. Significant regional differences
306 exist between Antarctic drainage basins in terms of oceanic heat fluxes and the
307 topographic configuration of the ice sheet bed⁸³. Consequently, the ice sheet response to
308 ocean thermal forcing, even for small temperature anomalies, may be governed by bed
309 geometry as much as by environmental conditions^{83,84}. Observations and modelling
310 show that surface melt occurs on some smaller ice shelves^{44,47,48}, but also that this may

311 not be a recent phenomenon⁴⁹. According to global and regional atmospheric modelling,
312 under intermediate emissions scenarios, Antarctic ice shelf surface melt will likely
313 increase gradually and linearly⁴⁸. It should be noted, however, that while surface melt is
314 not the major present-day forcing component, the high-end SLR contributions reached
315 for RCP8.5 scenarios⁵⁸ stem from increased surface melting rather than oceanic forcing.

316 The projected long-term SLR contribution (500 years) of AIS for warming levels
317 associated with the RCP2.6 scenario are limited to well below a metre, although with a
318 probability distribution that is not Gaussian and presents a long tail toward high values
319 due to potential MICI⁵⁸, with the caveats listed above. Importantly, substantial future
320 retreat in some basins (e.g. TG) cannot be ruled out and grounding-line retreat may
321 continue even with no additional forcing^{73,77,85,86}. The long-term SLR contribution of
322 AIS therefore crucially depends on the behaviour of individual ice shelves and outlet
323 glacier systems and whether they enter into MISI for the given level of warming. Under
324 sustained warming, a key threshold for survival of Antarctic ice shelves, and thus
325 stability of the ice sheet, appears to lie between 1.5 and 2 °C mean annual air
326 temperature above present (Figs. 1d and 4)⁸¹. Activation of several larger systems such
327 as the Ross and Ronne-Filchner drainage basins and onset of much larger SLR
328 contributions is estimated to be triggered by global warming between 2 and 2.7 °C⁸¹.
329 This implies that substantial Antarctic ice loss can be prevented only by limiting
330 greenhouse gas emissions to RCP2.6 levels or lower^{58,81}. Crossing these thresholds
331 implies commitment to large ice sheet changes and SLR that may take thousands of
332 years to be fully realised and are irreversible on longer timescales.

333 **Need for improvement**

334 While considerable progress has been made over the last decade with respect to
335 understanding processes at the interface between ice sheets, atmosphere and ocean,
336 significant uncertainties in both forcing and response of the ice sheets remain^{18,87}. For
337 the AIS, for instance, the majority of present-day mass loss (essentially the ASE) is
338 driven by changes in ocean circulation. The ability to simulate those changes into the
339 future is so far limited, leading to large remaining uncertainties for any projection of AIS
340 mass balance. Similar challenges remain in modelling changes in regional atmospheric
341 circulation that affect GrIS mass loss. Therefore, it is not clear to what degree global
342 warming must be limited to reduce future ice sheet-related SLR contributions.

343 Other challenges in climate and ice sheet modelling concern model resolution,
344 initialization and coupling. Model resolution is a key issue, as climate and ocean models
345 tend to be too diffusive. Higher model resolutions increase eddy activity and advective
346 heat transfer more readily than at lower resolution⁸⁸. Recent work⁸⁹ uses high-resolution,
347 non-hydrostatic atmospheric and detailed SMB models to better represent surface
348 physical processes at <10 km scales. Likewise, in order to resolve grounding-line
349 dynamics, ice sheet models need high spatial resolution across the grounding line⁹⁰ and
350 new numerical techniques, such as adaptive meshing, have been developed in recent
351 years to achieve this⁹¹. Model initialization relies on two distinct, but often combined
352 approaches (spin-up versus data assimilation; Box 1), the latter technique improving for
353 centennial projections with the increasing access to high-resolution satellite products.

354 Further developments include the need for two-way coupling of ice sheets with
355 coupled atmosphere-ocean models, meaning that climate models not only force ice-sheet

356 models but the reverse is also true. This calls for closer collaborations across disciplines,
357 which is exemplified by ice-sheet model intercomparisons (such as ISMIP6⁹²) within the
358 Coupled Model Intercomparison Project CMIP6. A similar intercomparison exercise for
359 SMB and ocean models is urgently needed, given remaining uncertainties in absolute
360 SMB values and sub-shelf melting, with the former especially relevant for
361 Greenland^{7,14,93} and the latter for Antarctica. For instance, if a possible link is found
362 between global warming and the current circulation changes observed in summer over
363 Greenland, this could significantly amplify the melt acceleration projected for the future
364 via a newly recognized positive feedback. Therefore, to achieve this, it will be critical to
365 further understand and improve the representation of changes in atmosphere and ocean
366 global circulation in global and regional climate model simulations.

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378 **Author contributions**

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

383 **Competing interests**

384 The authors declare no competing interests.

385 **References**

- 386 1. Enderlin, E. M. *et al.* An improved mass budget for the Greenland ice sheet.
387 *Geophys. Res. Lett.* **41**, 866–872. ISSN: 19448007 (2014).
- 388 2. Rignot, E., Jacobs, S, Mouginot, J & Scheuchl, B. Ice-shelf melting around
389 Antarctica. *Science* **341**, 266–70. ISSN: 1095-9203 (2013).
- 390 3. Depoorter, M. A. *et al.* Calving fluxes and basal melt rates of Antarctic ice
391 shelves. *Nature* **502**, 89–92. ISSN: 1476-4687 (2013).
- 392 4. Hanna, E., Mernild, S. H., Cappelen, J. & Steffen, K. Recent warming in
393 Greenland in a long-term instrumental (1881-2012) climatic context: I. Evaluation
394 of surface air temperature records. *Environmental Research Letters* **7**, 045404
395 (2012).
- 396 5. Hanna, E. *et al.* Ice-sheet mass balance and climate change. *Nature* **498**, 51–59.
397 ISSN: 0028-0836 (2013).

- 398 6. Bamber, J. L., Westaway, R. M., Marzeion, B. & Wouters, B. The land ice
399 contribution to sea level during the satellite era. *Environmental Research Letters*
400 **13**, 063008 (2018).
- 401 7. Wilton, D. *et al.* High resolution (1 km) positive degree-day modelling of
402 Greenland ice sheet surface mass balance, 1870–2012 using reanalysis data.
403 *Journal of Glaciology* (2016).
- 404 8. Fettweis, X. *et al.* Brief communication: Important role of the mid-tropospheric
405 atmospheric circulation in the recent surface melt increase over the Greenland ice
406 sheet. *Cryosphere* **7**, 241–248. ISSN: 19940416 (2013).
- 407 9. Hall, R., Erdélyi, R., Hanna, E., Jones, J. M. & Scaife, A. A. Drivers of North
408 Atlantic Polar Front jet stream variability. *International Journal of Climatology*
409 **35**, 1697–1720. ISSN: 1097-0088 (2015).
- 410 10. Lim, Y.-K. *et al.* Atmospheric summer teleconnections and Greenland Ice Sheet
411 surface mass variations: insights from MERRA-2 RECEIVED. *Environ. Res. Lett.*
412 **11** (2016).
- 413 11. Hanna, E., Cropper, T. E., Hall, R. J. & Cappelen, J. Greenland Blocking Index
414 1851–2015: a regional climate change signal. *Int. J. Climatol.* **36**, 4847–4861.
415 ISSN: 10970088 (2016).
- 416 12. Fettweis, X. *et al.* Estimating the Greenland ice sheet surface mass balance
417 contribution to future sea level rise using the regional atmospheric climate model
418 MAR. *The Cryosphere* **7**, 469–489 (2013).
- 419 13. Hofer, S., Tedstone, A. J., Fettweis, X. & Bamber, J. L. Decreasing cloud cover
420 drives the recent mass loss on the Greenland Ice Sheet. *Science Advances* **3**,
421 n/a–n/a (2017).
- 422 14. Van den Broeke, M. *et al.* Greenland Ice Sheet Surface Mass Loss: Recent
423 Developments in Observation and Modeling. *Current Climate Change Reports* **3**,
424 345–356. ISSN: 2198-6061 (2017).
- 425 15. Van Tricht, K. *et al.* Clouds enhance Greenland ice sheet meltwater runoff. *Nature*
426 *Communications* **7**. Article, 10266 EP – (2016).
- 427 16. Edwards, T. L. *et al.* Effect of uncertainty in surface mass balance-elevation
428 feedback on projections of the future sea level contribution of the Greenland ice
429 sheet. *Cryosphere* **8**, 195–208. ISSN: 19940416 (2014).
- 430 17. Vizcaino, M. *et al.* Coupled simulations of Greenland Ice Sheet and climate
431 change up to A.D. 2300. *Geophys. Res. Lett.* **42** (2015).

- 432 18. Goelzer, H., Robinson, A., Seroussi, H. & van de Wal, R. Recent Progress in
433 Greenland Ice Sheet Modelling. *Current Climate Change Reports* **3**, 291–302
434 (2017).
- 435 19. Moon, T., Joughin, I., Smith, B. & Howat, I. 21st-Century Evolution of Greenland
436 Outlet Glacier Velocities. *Science* **336**, 576–578. ISSN: 0036-8075 (2012).
- 437 20. Bigg, G. R. *et al.* A century of variation in the dependence of Greenland iceberg
438 calving on ice sheet surface mass balance and regional climate change.
439 *Proceedings of the Royal Society of London A: Mathematical, Physical and*
440 *Engineering Sciences* **470**, n/a–n/a. ISSN: 1364-5021 (2014).
- 441 21. Holland, D. M., Thomas, R., deYoung, B., Ribergaard, M. & Lyberth, B.
442 Acceleration of Jakobshavn Isbrae Triggered by Warm Subsurface Ocean Waters.
443 *Nature Geosci.* **1**, 659–664 (2008).
- 444 22. Khan, S. A. *et al.* Sustained mass loss of the northeast Greenland ice sheet
445 triggered by regional warming. *Nature Climate Change* (2014).
- 446 23. Nick, F. M. *et al.* Future sea-level rise from Greenland’s main outlet glaciers in a
447 warming climate. *Nature* **497**, 235–238. ISSN: 1476-4687 (2013).
- 448 24. Zwally, H. J. *et al.* Surface Melt-Induced Acceleration of Greenland Ice-Sheet
449 Flow. *Science* **297**, 218–222. ISSN: 0036-8075 (2002).
- 450 25. Sundal, A. *et al.* Melt-induced speed-up of Greenland ice sheet offset by efficient
451 subglacial drainage. *Nature* **469**, 521–524. ISSN: 0028-0836 (Jan. 2011).
- 452 26. Shannon, S. R. *et al.* Enhanced basal lubrication and the contribution of the
453 Greenland ice sheet to future sea-level rise. *Proc. Natl. Acad. Sci.* **110**,
454 14156–14161. ISSN: 0027-8424 (2013).
- 455 27. Fürst, J. J., Goelzer, H. & Huybrechts, P. Ice-dynamic projections of the
456 Greenland ice sheet in response to atmospheric and oceanic warming. *The*
457 *Cryosphere* **9**, 1039–1062 (2015).
- 458 28. Goelzer, H. *et al.* Sensitivity of Greenland ice sheet projections to model
459 formulations. *J. Glaciol.* **59**, 733–749. ISSN: 00221430 (2013).
- 460 29. Nowicki, S. *et al.* Insights into spatial sensitivities of ice mass response to
461 environmental change from the SeaRISE ice sheet modeling project II:
462 Greenland. *Journal of Geophysical Research: Earth Surface* **118**, 1025–1044.
463 ISSN: 2169-9011 (2013).
- 464 30. Morlighem, M. *et al.* Modeling of Store Gletscher’s calving dynamics, West
465 Greenland, in response to ocean thermal forcing. *Geophys. Res. Lett.* **43**,
466 2659–2666. ISSN: 19448007 (2016).

- 467 31. Aschwanden, A., Fahnestock, M. A. & Truffer, M. Complex Greenland outlet
468 glacier flow captured. *Nature Communications* **7**, 10524. ISSN: 2041-1723 (2016).
- 469 32. Morlighem, M., Rignot, E., Mouginot, J., Seroussi, H. & Larour, E. Deeply
470 incised submarine glacial valleys beneath the Greenland ice sheet. *Nature*
471 *Geoscience* **7**, 418–422 (June 2014).
- 472 33. Benn, D. I., Warren, C. R. & Mottram, R. H. Calving processes and the dynamics
473 of calving glaciers. *Earth-Science Reviews* **82**, 143–179. ISSN: 00128252 (2007).
- 474 34. Bondzio, J. H. *et al.* The mechanisms behind Jakobshavn Isbrae’s acceleration
475 and mass loss: A 3-D thermomechanical model study. *Geophysical Research*
476 *Letters* **44**. 2017GL073309, 6252–6260. ISSN: 1944-8007 (2017).
- 477 35. Robinson, A. & Goelzer, H. The importance of insolation changes for paleo ice
478 sheet modeling. *The Cryosphere* **8**, 1419–1428 (2014).
- 479 36. Tedesco, M. *et al.* The darkening of the Greenland ice sheet: trends, drivers, and
480 projections (1981–2100). *The Cryosphere* **10**, 477–496 (2016).
- 481 37. Tedstone, A. J. *et al.* Dark ice dynamics of the south-west Greenland Ice Sheet.
482 *The Cryosphere* **11**, 2491–2506 (2017).
- 483 38. Ryan, J. C. *et al.* Dark zone of the Greenland Ice Sheet controlled by distributed
484 biologically-active impurities. *Nature Communications* **9**, 1065. ISSN: 2041-1723
485 (2018).
- 486 39. Ridley, J., Gregory, J. M., Huybrechts, P. & Lowe, J. Thresholds for irreversible
487 decline of the Greenland ice sheet. *Clim. Dyn.* **35**, 1065–1073. ISSN: 09307575
488 (2010).
- 489 40. Robinson, A., Calov, R. & Ganopolski, A. Multistability and critical thresholds of
490 the Greenland ice sheet. *Nat. Clim. Chang.* **2**, 429–432. ISSN: 1758-678X (2012).
- 491 41. Levermann, A. *et al.* The multimillennial sea-level commitment of global
492 warming. *Proc. Natl. Acad. Sci.* **110**, 13745–13750. ISSN: 0027-8424 (2013).
- 493 42. Shepherd, A. *et al.* Mass balance of the Antarctic Ice Sheet from 1992 to 2017.
494 *Nature* **558**, 219–222. ISSN: 1476-4687 (2018).
- 495 43. Smith, A. M., Bentley, C. R., Bingham, R. G. & Jordan, T. A. Rapid subglacial
496 erosion beneath Pine Island Glacier, West Antarctica. *Geophysical Research*
497 *Letters* **39**. L12501, n/a–n/a. ISSN: 1944-8007 (2012).
- 498 44. Lenaerts, J. T. M., Vizcaino, M., Fyke, J., van Kampenhout, L. &
499 van den Broeke, M. R. Present-day and future Antarctic ice sheet climate and
500 surface mass balance in the Community Earth System Model. *Clim. Dyn.* **47**,
501 1367–1381. ISSN: 14320894 (2016).

- 502 45. Thomas, E. R. *et al.* Regional Antarctic snow accumulation over the past 1000
503 years. *Climate of the Past* **13**, 1491–1513 (2017).
- 504 46. Palerme, C. *et al.* Evaluation of current and projected Antarctic precipitation in
505 CMIP5 models. *Clim. Dyn.* **48**, 225–239. ISSN: 14320894 (2017).
- 506 47. Kuipers Munneke, P., Picard, G., Van Den Broeke, M. R., Lenaerts, J. T. M. &
507 Van Meijgaard, E. Insignificant change in Antarctic snowmelt volume since 1979.
508 *Geophys. Res. Lett.* **39**, 6–10. ISSN: 00948276 (2012).
- 509 48. Trusel, L. D. *et al.* Divergent trajectories of Antarctic surface melt under two
510 twenty-first-century climate scenarios. *Nat. Geosci.* **8**, 927–932. ISSN: 1752-0894
511 (2015).
- 512 49. Kingslake, J., Ely, J. C., Das, I. & Bell, R. E. Widespread movement of meltwater
513 onto and across Antarctic ice shelves. *Nature* **544**, 349 EP – (2017).
- 514 50. Bell, R. E. *et al.* Antarctic ice shelf potentially stabilized by export of meltwater
515 in surface river. *Nature* **544**. Letter, 344–348. ISSN: 0028-0836 (2017).
- 516 51. Reese, R., Gudmundsson, G. H., Levermann, A. & Winkelmann, R. The far reach
517 of ice-shelf thinning in Antarctica. *Nature Climate Change* **8**, 53–57. ISSN:
518 1758-6798 (2018).
- 519 52. Borstad, C. *et al.* A constitutive framework for predicting weakening and reduced
520 buttressing of ice shelves based on observations of the progressive deterioration of
521 the remnant Larsen B Ice Shelf. *Geophys. Res. Lett.* **43**, 2027–2035. ISSN:
522 19448007 (2016).
- 523 53. Scambos, T. A., Hulbe, C., Fahnestock, M. & Bohlander, J. The link between
524 climate warming and break-up of ice shelves in the Antarctic Peninsula. *J.*
525 *Glaciol.* **46**, 516–530. ISSN: 00221430 (2000).
- 526 54. Munneke, P. K., Ligtenberg, S. R. M., Van Den Broeke, M. R. & Vaughan, D. G.
527 Firn air depletion as a precursor of Antarctic ice-shelf collapse. *J. Glaciol.* **60**,
528 205–214. ISSN: 00221430 (2014).
- 529 55. Banwell, A. F., MacAyeal, D. R. & Sergienko, O. V. Breakup of the Larsen B Ice
530 Shelf triggered by chain reaction drainage of supraglacial lakes. *Geophys. Res.*
531 *Lett.* **40**, 5872–5876. ISSN: 00948276 (2013).
- 532 56. Massom, R. A. *et al.* Antarctic ice shelf disintegration triggered by sea ice loss
533 and ocean swell. *Nature* **558**, 383–389. ISSN: 1476-4687 (2018).
- 534 57. Abram, N. J. *et al.* Acceleration of snow melt in an Antarctic Peninsula ice core
535 during the twentieth century. *Nature Geoscience* **6**, 404–411 (2013).
- 536 58. DeConto, R. M. & Pollard, D. Contribution of Antarctica to past and future
537 sea-level rise. *Nature* **531**, 591–597. ISSN: 0028-0836 (2016).

- 538 59. Jacobs, S. S., Jenkins, A., Giulivi, C. F. & Dutrieux, P. Stronger ocean circulation
539 and increased melting under Pine Island Glacier ice shelf. *Nature Geoscience* **4**,
540 519 EP – (2011).
- 541 60. Pritchard, H. D. *et al.* Antarctic ice-sheet loss driven by basal melting of ice
542 shelves. *Nature* **484**, 502 EP – (2012).
- 543 61. Greenbaum, J. S. *et al.* Ocean access to a cavity beneath Totten Glacier in East
544 Antarctica. *Nature Geoscience* **8**, 294–298 (Apr. 2015).
- 545 62. Wouters, B. *et al.* Dynamic thinning of glaciers on the Southern Antarctic
546 Peninsula. *Science* **348**, 899–903. ISSN: 0036-8075 (2015).
- 547 63. Paolo, F. S. *et al.* Response of Pacific-sector Antarctic ice shelves to the El
548 Niño/Southern Oscillation. *Nature Geoscience*, n/a–n/a. ISSN: 1752-0908 (2018).
- 549 64. Little, C. M. & Urban, N. M. CMIP5 temperature biases and 21st century
550 warming around the Antarctic coast. *Ann. Glaciol.* **57**, 69–78. ISSN: 02603055
551 (2016).
- 552 65. Asay-Davis, X. S., Jourdain, N. C. & Nakayama, Y. Developments in Simulating
553 and Parameterizing Interactions Between the Southern Ocean and the Antarctic
554 Ice Sheet. *Current Climate Change Reports* **3**, 316–329 (2017).
- 555 66. Dinniman, M. S., Klinck, J. M. & Hofmann, E. E. Sensitivity of circumpolar deep
556 water transport and ice shelf basal melt along the west antarctic peninsula to
557 changes in the winds. *J. Clim.* **25**, 4799–4816. ISSN: 08948755 (2012).
- 558 67. Kusahara, K. & Hasumi, H. Pathways of basal meltwater from Antarctic ice
559 shelves: A model study. *Journal of Geophysical Research: Oceans* **119**,
560 5690–5704. ISSN: 2169-9291 (2014).
- 561 68. Hellmer, H. H., Kauker, F., Timmermann, R., Determann, J. & Rae, J.
562 Twenty-first-century warming of a large Antarctic ice-shelf cavity by a redirected
563 coastal current. *Nature* **485**, 225–228. ISSN: 0028-0836 (2012).
- 564 69. Timmermann, R. & Hellmer, H. H. Southern Ocean warming and increased ice
565 shelf basal melting in the twenty-first and twenty-second centuries based on
566 coupled ice-ocean finite-element modelling. *Ocean Dyn.* **63**, 1011–1026. ISSN:
567 16167341 (2013).
- 568 70. Schoof, C. Ice sheet grounding line dynamics: Steady states, stability, and
569 hysteresis. *Journal of Geophysical Research: Earth Surface* **112**. F03S28,
570 n/a–n/a. ISSN: 2156-2202 (2007).
- 571 71. Pattyn, F. *et al.* Results of the Marine Ice Sheet Model Intercomparison Project,
572 MISMIP. *The Cryosphere* **6**, 573–588 (2012).

- 573 72. Favier, L. *et al.* Retreat of Pine Island Glacier controlled by marine ice-sheet
574 instability. *Nat. Clim. Chang.* **4**, 117–121. ISSN: 1758-678X (2014).
- 575 73. Joughin, I., Smith, B. E. & Medley, B. Marine Ice Sheet Collapse Potentially
576 Under Way for the Thwaites Glacier Basin, West Antarctica. *Science* **344**,
577 735–738. ISSN: 0036-8075 (2014).
- 578 74. Mouginot, J., Rignot, E. & Scheuchl, B. Sustained increase in ice discharge from
579 the Amundsen Sea Embayment, West Antarctica, from 1973 to 2013.
580 *Geophysical Research Letters* **41**, 1576–1584. ISSN: 1944-8007 (2014).
- 581 75. Dutrieux, P. *et al.* Strong Sensitivity of Pine Island Ice-Shelf Melting to Climatic
582 Variability. *Science (80-.)*. **343** (2014).
- 583 76. Nias, I. J., Cornford, S. L. & Payne, A. J. Contrasting the Modelled sensitivity of
584 the Amundsen Sea Embayment ice streams. *J. Glaciol.* **62**, 552–562. ISSN:
585 00221430 (2016).
- 586 77. Seroussi, H. *et al.* Continued retreat of Thwaites Glacier, West Antarctica,
587 controlled by bed topography and ocean circulation. *Geophys. Res. Lett.* n/a–n/a.
588 ISSN: 00948276 (2017).
- 589 78. Bassis, J. N. & Walker, C. C. Upper and lower limits on the stability of calving
590 glaciers from the yield strength envelope of ice. *Proc. R. Soc. A Math. Phys. Eng.*
591 *Sci.* **468**, 913–931. ISSN: 1364-5021 (2012).
- 592 79. Pollard, D., DeConto, R. M. & Alley, R. B. Potential Antarctic Ice Sheet retreat
593 driven by hydrofracturing and ice cliff failure. *Earth and Planetary Science*
594 *Letters* **412**, 112–121 (2015).
- 595 80. Cornford, S. L. *et al.* Century-scale simulations of the response of the West
596 Antarctic Ice Sheet to a warming climate. *Cryosphere* **9**, 1579–1600. ISSN:
597 19940424 (2015).
- 598 81. Gollledge, N. R. *et al.* The multi-millennial Antarctic commitment to future
599 sea-level rise. *Nature* **526**, 421–425. ISSN: 0028-0836 (2015).
- 600 82. Gasson, E., DeConto, R. M., Pollard, D. & Levy, R. H. Dynamic Antarctic ice
601 sheet during the early to mid-Miocene. *Proceedings of the National Academy of*
602 *Sciences* **113**, 3459–3464. ISSN: 0027-8424 (2016).
- 603 83. Gollledge, N. R., Levy, R. H., McKay, R. M. & Naish, T. R. East Antarctic ice
604 sheet most vulnerable to Weddell Sea warming. *Geophysical Research Letters* **44**.
605 2016GL072422, 2343–2351. ISSN: 1944-8007 (2017).
- 606 84. Mengel, M. *et al.* Future sea level rise constrained by observations and long-term
607 commitment. *Proc. Natl. Acad. Sci.* **113**, 2597–2602. ISSN: 0027-8424 (2016).

- 608 85. Arthern, R. J. & Williams, C. R. The sensitivity of West Antarctica to the
609 submarine melting feedback. *Geophys. Res. Lett.* **44**, 2352–2359. ISSN: 19448007
610 (2017).
- 611 86. Waibel, M. S., Hulbe, C. L., Jackson, C. S. & Martin, D. F. Rate of Mass Loss
612 Across the Instability Threshold for Thwaites Glacier Determines Rate of Mass
613 Loss for Entire Basin. *Geophysical Research Letters*. 2017GL076470, n/a–n/a.
614 ISSN: 1944-8007 (2018).
- 615 87. Pattyn, F., Favier, L. & Sun, S. Progress in Numerical Modeling of Antarctic
616 Ice-Sheet Dynamics. *Current Climate Change Reports* **3**, 174–184 (2017).
- 617 88. Stewart, A. L., Klocker, A. & Menemenlis, D. Circum-Antarctic Shoreward Heat
618 Transport Derived From an Eddy- and Tide-Resolving Simulation. *Geophysical
619 Research Letters* **45**. 2017GL075677, 834–845. ISSN: 1944-8007 (2018).
- 620 89. Niwano, M. *et al.* NHM–SMAP: spatially and temporally high-resolution
621 nonhydrostatic atmospheric model coupled with detailed snow process model for
622 Greenland Ice Sheet. *The Cryosphere* **12**, 635–655 (2018).
- 623 90. Durand, G. & Pattyn, F. Reducing uncertainties in projections of Antarctic ice
624 mass loss. *The Cryosphere* **9**, 2043–2055 (2015).
- 625 91. Cornford, S. L. *et al.* Adaptive mesh, finite volume modeling of marine ice sheets.
626 *J. Comput. Phys.* **232**, 529–549. ISSN: 00219991 (2013).
- 627 92. Nowicki, S. M. J. *et al.* Ice Sheet Model Intercomparison Project (ISMIP6)
628 contribution to CMIP6. *Geoscientific Model Development* **9**, 4521–4545 (2016).
- 629 93. Vernon, C. L. *et al.* Surface mass balance model intercomparison for the
630 Greenland ice sheet. *The Cryosphere* **7**, 599–614 (2013).
- 631 94. Nowicki, S. *et al.* Insights into spatial sensitivities of ice mass response to
632 environmental change from the SeaRISE ice sheet modeling project I: Antarctica.
633 *Journal of Geophysical Research: Earth Surface* **118**, 1025–1044. ISSN:
634 21699011 (2013).
- 635 95. Goelzer, H. *et al.* Design and results of the ice sheet model initialisation
636 experiments initMIP-Greenland: an ISMIP6 intercomparison. *The Cryosphere* **12**,
637 1433–1460 (2018).
- 638 96. Zickfeld, K., Solomon, S. & Gilford, D. M. Centuries of thermal sea-level rise
639 due to anthropogenic emissions of short-lived greenhouse gases. *Proceedings of
640 the National Academy of Sciences* **114**, 657–662 (2017).
- 641 97. Clark, P. U. *et al.* Consequences of twenty-first-century policy for
642 multi-millennial climate and sea-level change. *Nat. Clim. Chang.* **6**, 360–369.
643 ISSN: 1758-678X (2016).

- 644 98. Thomas, Z. A. Using natural archives to detect climate and environmental tipping
645 points in the Earth System. *Quat. Sci. Rev.* **152**, 60–71. ISSN: 02773791 (2016).
- 646 99. Robinson, A., Calov, R. & Ganopolski, A. An efficient regional energy-moisture
647 balance model for simulation of the Greenland Ice Sheet response to climate
648 change. *The Cryosphere* **4**, 129–144 (2010).
- 649 100. Ganopolski, A., Winkelmann, R. & Schellnhuber, H. J. Critical insolation-CO₂
650 relation for diagnosing past and future glacial inception. *Nature* **534**, 1–2. ISSN:
651 0028-0836 (2016).

652 **Highlighted references**

653 **Bamber et al.**⁶ A systematic, detailed and insightful review of Greenland Ice Sheet (and
654 other land ice) mass balance changes between 1992 and 2016, that provides a very
655 useful post-AR5 synthesis.

656 **Hofer et al.**¹³ This study highlights the importance of Greenland cloud cover changes
657 on surface energy and mass balance.

658 **van den Broeke et al.**¹⁴ A state-of-the-science critical review of outstanding research
659 questions in Greenland Ice Sheet surface mass balance work.

660 **Fuerst et al.**²⁷ Authoritative study on Greenland Ice Sheet future change and resulting
661 sea-level rise to 2300, indicating that volume loss is mainly caused by increased
662 surface melting and that the largest modelled uncertainties relate to surface mass
663 balance and the underpinning climate projections rather than ice-sheet dynamics.

664 **Tedesco et al.**³⁶ An excellent and detailed review highlighting the importance of
665 Greenland albedo changes.

666 **Shepherd et al.**⁴² Most recent and up-to-date mass balance estimate of the Antarctic ice
667 sheet showing significant increased contributions the ice sheet to SLR over the last
668 decade.

669 **DeConto and Pollard**⁵⁸ High-end projections of Antarctic ice sheet contribution to
670 SLR based on ice shelf hydrofracturing and subsequent ice cliff collapse.

671 **Golledge et al.**⁸¹ Long-term (multi-millennial) projections of the Antarctic ice sheets
672 and potential tipping points.

673 **Pattyn et al.**⁸⁷ Review on recent advances in modelling of the Antarctic ice sheet.

674 Highlights our current understanding of ice-dynamical processes that are key to
675 future predictions.

676 **Nowicki et al.**⁹² Outline of the new phase of ice-sheet model intercomparisons linked to
677 the Coupled Model Intercomparison Project CMIP6.

Figure 1: Annual mean surface mass fluxes (in Gt a^{-1}) as a function of global mean temperature anomaly with respect to the preindustrial era (1850-1920). (a) GrIS SMB, (b) GrIS runoff, (c) Antarctic SMB, (d) Antarctic surface melt. Red colours indicate model realizations of present-day ice sheets (RACMO2 and MAR forced by ERA reanalysis data). Blue colours indicate model realizations of future ice sheets. In panel (a) and (b), MAR is forced with CESM-CAM5 1.5 and 2.0 future scenarios (+1.5 and 2.0 °C w.r.t. preindustrial). In panel (c), RACMO2 is forced with a HadCM3 A1B scenario. In panel (d), CESM-CAM5 1.5 and 2.0 future scenarios include surface melt parametrized in terms of near-surface temperature⁴⁸. Trend lines are shown for future (blue) model realizations. Boxes delimit two standard deviations in temperature and SMB components over the present-day period (red boxes) and the stationary climate over 2061-2100 in the CESM-CAM5 1.5 (light blue boxes) and 2.0 (dark blue boxes) scenarios. None of these simulations include coupling to an ice-dynamical model.

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

Figure 3: MISI and MICI as main drivers for potential (partial) collapse of the Antarctic ice sheet. MISI (a) can lead to unstable retreat of grounding lines resting on retrograde bed slopes, a very common situation in Antarctica. MISI stems from a positive feedback loop between the increased (i) flux and (ii) ice thickness at the grounding line after the latter starts to retreat. MICI (b) is the result of collapse of exposed ice cliffs (after the ice shelf collapses due to hydro-fracturing) under their own weight. MISI applies for a retrograde slope bed, while MICI can also apply for prograde slopes. Both MISI and MICI are thus superimposed for retrograde slopes^{58,87}. The red colour qualifies the heat forcing exerted by the ocean against the ice shelf basal surface.

Figure 4: AIS stability as a function of the imposed regional annual mean temperature anomaly. Changes in SMB (a) and SLR contribution (b) for AIS relative to 2000 CE as simulated under spatially-uniform temperature increases that follow RCP trajectories to 2300 CE and then stabilize⁸¹. Colored lines denote different years (CE) data are averages of ‘high’ and ‘low’ scenarios, denoting two different grounding-line parametrisations. Grey shading shows approximate equivalent global mean temperature anomaly for an Antarctic mean temperature anomaly of 1.5-2.0 °C, accounting for polar amplification.

678 **Box 1: Projections of ice sheet mass loss**

679 **Projections of ice-sheets contribution** to SLR are established using ice flow models
680 that compute the evolution of ice sheets under given climate scenarios. Many of these
681 models were constructed to study the evolution of ice sheets across glacial-interglacial
682 cycles, and are not therefore ideally suited to making projections for this century.
683 Accordingly, the last decade has seen the modelling community repurpose these many
684 models, increasing the confidence in the skill of ice-sheet models (particularly
685 interaction with boundary conditions, such as ice/ocean and ice/bedrock), but they still
686 lag somewhat behind other areas of the climate system.

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

694 For low-emission scenarios and in the near term, the **initial state used by ice sheet**
695 **models is a key control** on the reliability of their projections because the anticipated
696 mass loss is relatively small in comparison to the total mass of the ice sheets. Two main
697 families of initialization strategies are currently employed. The first is **spin-up** of the
698 model over glacial-interglacial periods, which ensures that the internal properties of the
699 ice sheet are consistent with each other but which may have an inaccurate representation
700 of the ice sheets' contemporary geometry and velocity. The alternative is the

701 **assimilation** of satellite data, which may lead to inconsistencies in flow properties but
702 has a greatly improved representation of current geometry and surface velocity. These
703 two approaches lead to large differences in the initial conditions from which projections
704 are made and therefore create a significant spread in projected contributions to future
705 SLR, even when forced with similar datasets^{29,94}. Disentangling the impacts of natural
706 variability and forced climate change is also more difficult for these low emission
707 scenarios, but new model intercomparisons tend to focus on this aspect⁹⁵.

708 **Box 2: Climate commitment and tipping points**

709 For the long-term evolution of the ice sheets, on multi-centennial to multi-millennial
710 time scales, feedbacks with the atmosphere and ocean increase in importance. When
711 subjected to perturbed climatic forcing over this time scale, the ice sheets manifest large
712 changes in their volume and distribution. These changes typically occur with a
713 significant lag in response to the forcing applied, which leads to the concept of climate
714 commitment: changes that will occur in the long-term future, are committed to at a much
715 earlier stage⁹⁶. Because of the long residence time of CO₂ in the atmosphere, climate
716 change in coming decades will most probably last long enough to dictate ice sheet
717 evolution over centuries and millennia^{41,58,81,97}. Furthermore, the ice sheets are subject to
718 threshold behaviour in their stability, since a change in boundary conditions like climate
719 forcing can cause the current ice-sheet configuration to be unstable. Crossing this
720 so-called tipping point leads the system to equilibrate to a qualitatively different state⁹⁸
721 (by melting completely, for example). The existence of a tipping point implies that
722 ice-sheet changes are potentially irreversible — returning to a pre-industrial climate may
723 not stabilize the ice sheet once the tipping point has been crossed. A key concept here is
724 the timeframe of reversal, because many ice sheet changes may only be reversible over
725 e.g. a full glacial-interglacial cycle with natural rates of changes in climatic variables.
726 For both Greenland and Antarctica, tipping points are known to exist for warming levels
727 that could be reached before the end of this century^{58,81,99}. The unprecedented rate of
728 increase in GHGs over the Anthropocene leaves open the question of irreversible
729 crossing of tipping points. For example, it is possible that the expected future increase in
730 GHGs will prevent or delay the next ice sheet inception¹⁰⁰.