

Response of the East Antarctic Ice Sheet to Past and Future Climate Change

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Preface: The East Antarctic Ice Sheet (EAIS) contains the vast majority of Earth's glacier ice (~52 metres sea-level equivalent), but is often viewed as less vulnerable to global warming than the West Antarctic or Greenland ice sheets. However, some regions of the EAIS have lost mass over recent decades, prompting the need to re-evaluate its sensitivity to climate change. Here we review the EAIS's response to past warm periods, synthesise current observations of change, and evaluate future projections. Some marine-based catchments that underwent significant mass loss during past warm periods are currently losing mass, but most projections indicate increased accumulation across the EAIS over the 21st Century, keeping the ice sheet broadly in balance. Beyond 2100, high emissions scenarios generate increased ice discharge and potentially several metres of sea-level rise within just a few centuries, but substantial mass loss could be averted if the Paris Agreement to limit warming below 2°C is satisfied.

46 1. Introduction

47 Over recent decades, ice loss from Antarctica has exceeded mass gains and its contribution
48 to sea-level rise has accelerated¹⁻⁹. The largest imbalances are found in the West Antarctic
49 Ice Sheet (WAIS: Fig. 1d), which holds 5.3 m sea-level equivalent (SLE)¹⁰ and lost over 2,000
50 Gt of ice between 1992 and 2017, adding ~6 mm to global mean sea level¹. This imbalance is
51 attributed to warm ocean currents - modified Circumpolar Deep Water (CDW) - melting the
52 underside of floating ice shelves, causing marginal ice thinning, grounding line retreat and
53 increased ice discharge¹¹⁻¹⁸. Furthermore, the WAIS is marine-based, resting on topography
54 below sea level that deepens inland (Fig. 1e)¹⁰. In the absence of buttressing ice shelves¹⁶,
55 retreat could rapidly propagate inland via a feedback known as 'Marine Ice Sheet Instability'¹⁹.

56 The vulnerability of the WAIS was first recognised in the 1970s²⁰, prompting a huge
57 growth in research²¹. In comparison, much less work has focussed on the vulnerability of the
58 East Antarctic Ice Sheet (EAIS), which is an order of magnitude larger (52.2 m SLE)¹⁰ and
59 generally viewed as less sensitive to ocean-climate warming. This view stems from the fact
60 that large parts of the EAIS have persisted for millions of years²², recent mass balance
61 estimates tend towards equilibrium or show modest mass gains^{1,2,5,23,24} (Fig. 2), and most
62 model projections show low sensitivity to climate change over the next century²⁵. However,
63 some recent observations suggest the EAIS may be more sensitive than previously thought.
64 Although uncertainties are large and often obscure even the sign of any change, the latest
65 efforts to reconcile its mass balance from multiple methods^{1,2} have raised the possibility of
66 overall mass loss since ~2014 (Fig. 2). Furthermore, numerous studies^{4,5,6,8}, including those
67 reporting overall mass gain^{7,9,23,24}, detect clear signals of regional mass loss from some
68 marine-based catchments (e.g. Wilkes Land: Fig. 1e). Like the WAIS, mass loss has been
69 attributed to modified CDW proximal to major outlet glaciers^{4,26-30} that may also be susceptible
70 to Marine Ice Sheet Instability^{10,31}, driving ice sheet thinning^{5,7,11,32}, grounding line retreat<sup>17,33-
71 35</sup>, and the retreat^{30,36} and disintegration^{37,38} of floating ice tongues/shelves.

72 Of further concern is that marine-based sectors of the EAIS lost mass during past warm
73 periods³⁹⁻⁴² and some numerical modelling predicts significant sea-level contributions from
74 them over the coming centuries^{31,43-46}. In contrast, 'terrestrial' regions of the EAIS, grounded
75 on land well above sea level, are gaining mass through increased accumulation (e.g. Dronning
76 Maud Land: Fig. 1e), albeit with large inter-annual variability^{5-9,47,48}. A key issue is that
77 observational time-series of accumulation or ice discharge are generally too short to elucidate
78 whether recent trends are significant or represent natural variability in the ocean-climate
79 system⁴⁹⁻⁵¹, prompting a question of huge scientific and societal importance: what will happen
80 to the East Antarctic Ice Sheet? With an emphasis on marine-based versus terrestrial sectors,
81 we address this question by reviewing how the EAIS changed during past warm periods and

82 during deglaciation from the Last Glacial Maximum; synthesise current observations of
83 change; and then evaluate future projections through to 2500.

84

85 **2. Response to Past Warm Periods**

86 Since widespread glaciation of Antarctica at the Eocene-Oligocene Transition⁵² (34–33.5 Ma:
87 [Fig. 1g](#)), climatic changes have caused substantial advance and retreat of the EAIS^{53,54}. Early
88 to Mid-Miocene (24–14 Ma) sediment records in the Ross Sea, for example, provide evidence
89 for multiple orbitally-paced fluctuations in EAIS extent^{55,56}, recorded by erosional hiatuses
90 representing expansion⁵⁷, and sediment provenance and vegetation changes indicating parts
91 of East Antarctica were ice-free^{58,59}.

92 The largest reduction in EAIS volume during the past 20 million years occurred during
93 the Mid-Miocene Climatic Optimum (17.0–14.8 Ma), when average atmospheric CO₂
94 concentrations were around 600–800 ppm (ranging from 300–1400 ppm)⁶⁰ and sea surface
95 temperatures peaked at ~11–17°C off the Adélie Coast⁶¹ and ~6–10 °C in the Ross Sea⁵⁶.
96 Under these conditions, ice sheet modelling⁵⁷ can simulate tens of metres of SLE contribution
97 from East Antarctica, with mass loss focussed in the three main subglacial basins: the Aurora
98 (ASB), Wilkes (WSB) and Recovery (RSB) ([Fig. 1a](#)). Terrestrial sectors are also likely to have
99 retreated, but recent sediment provenance analysis from the central Ross Sea⁶² reveals that
100 far-field sea-level records⁶³ can be reconciled without substantial loss of terrestrial ice in East
101 Antarctica, consistent with coupled climate and ice sheet modeling⁵⁷. Notably, average Mid-
102 Miocene CO₂ concentrations could be reached by 2100⁶⁴; although orbital forcing was
103 stronger than present and global atmospheric temperatures (7–8°C)⁶⁵ were significantly
104 warmer than projected for 2100.

105 The most recent epoch when atmospheric CO₂ concentrations last exceeded 400 ppm
106 was the Pliocene (5.33–2.58 Ma)⁶⁶. Mid-Pliocene (~3.3–3.0 Ma) atmospheric temperatures
107 were ~2–4°C warmer than present⁶⁷ and global mean sea level was around 10–25 m higher<sup>68-
108 71</sup>. Given the combined volume of the WAIS and Greenland Ice Sheet (~12 m), and that their
109 mid-Pliocene minima were likely asynchronous⁷², most sea-level estimates require an EAIS
110 contribution^{21,54,71}.

111 Early work on the EAIS response to mid-Pliocene warming focussed on marine
112 diatoms in subglacial diamictites of the Sirius Group in the Transantarctic Mountains^{22,73}.
113 Ambiguity regarding the transport mechanism of these diatoms made it difficult to locate which
114 regions lost mass, but recent work⁷⁴ supports at least partial retreat of both the ASB and WSB.
115 Marine sediment records provide more direct evidence of substantial retreat of the WSB,
116 inferred from a shift in the provenance of fine-grained detrital sediment³⁹, contemporaneous

117 with a shift in marine productivity, indicating a reduction in local sea-ice coverage⁷⁵ and a
118 southward migration of the Southern Ocean Polar Front⁷⁶. Substantial retreat of the ASB is
119 also inferred from records of ice-rafted debris^{40,77}, and from erosional signatures beneath the
120 catchment of Totten Glacier, which could have contributed over 2 m SLE⁷⁸. Elsewhere,
121 evidence of elevated fjord temperatures and vegetated landscapes in the Transantarctic
122 Mountains suggests significant retreat of marine-terminating glaciers⁷⁹; and the Lambert
123 Glacier-Amery Ice Shelf system was highly sensitive to Southern Ocean warming⁸⁰.

124 In contrast, mid-Pliocene retreat of terrestrial sectors of the EAIS and/or the RSB is
125 largely unknown, due to a lack of empirical evidence. Some ice sheet modelling is able to
126 simulate retreat and thinning of the RSB^{44,46,81}, alongside the WSB and ASB (Fig. 1b), but it
127 has generally proved challenging to simulate mid-Pliocene retreat of marine-based
128 sectors^{54,82}. The amount of modelled retreat is sensitive to assumed pre-Pliocene ice sheet
129 configurations⁸¹, climate model forcing⁸³, and ice sheet model parameters⁸⁴, with those
130 simulating the most retreat (e.g. Fig. 1b) often requiring additional processes to enhance mass
131 loss^{44,54,81,85}, some of which are debated (e.g. Marine Ice-Cliff Instability, discussed below)⁸⁶.

132 Marine-based sectors were also vulnerable during the warmest interglacials of the
133 Pleistocene (2.58–0.017 Ma), with offshore evidence of mass loss in the WSB⁴¹. During
134 Marine Isotope Stage 11 (~400 ka), subglacial precipitates of opal and calcite⁴² suggest the
135 ice margin was ~700 km inland of its current position, potentially contributing ~3–4 m SLE
136 when global atmospheric temperatures were only 1–2°C warmer than present. Terrestrial
137 records from the Transantarctic Mountains⁸⁷ also indicate ice surface elevation fluctuations of
138 hundreds of metres during the Pleistocene, similar in magnitude to those during the Pliocene.

139 Mass loss during the last interglacial (Marine Isotope Stage 5: ~130–115 ka) is
140 equivocal²¹. Ice cores and glacio-isostatic adjustment modelling⁸⁸ suggest ice-surface
141 lowering is plausible in the WSB and ASB, with other work⁸⁹ placing an upper sea-level
142 contribution of 0.4–0.8 m from the WSB. Sea-level records⁷¹ require no more than a few
143 metres from Antarctica, which is more likely from the WAIS, but the EAIS cannot currently be
144 ruled out.

145

146 **3. The Last Deglaciation**

147 During the Last Glacial Maximum (27–20 ka), marine-based sectors of the EAIS expanded to
148 near the continental shelf edge⁹⁰ (Fig. 1c). Evidence of ice margin extent and subsequent
149 retreat is only available from some regions, but typically indicates deglaciation commencing
150 at ~19–18 ka (e.g. the Lambert-Amery system), with grounding lines reaching the mid-
151 continental shelf in some locations at ~15 ka (e.g. Ross Sea sector)⁹⁰⁻⁹². The maximum extent

152 in the Weddell Sea sector is less clear⁹³, with recent evidence⁹⁴ of an oscillating grounding
153 line position on the outer continental shelf until ~12 ka. A rapid rise in global sea level occurred
154 at ~14.6 ka (Meltwater Pulse 1a), but Antarctica's contribution was limited (≤ 1.3 m)⁹⁵; with little
155 direct evidence (e.g. ice surface elevation changes) for substantial changes in EAIS
156 geometry^{90,92,93}, although increased ice-rafted debris is recorded in the vicinity of the Weddell
157 Sea⁹⁶.

158 Most regions of the EAIS had retreated prior to the Holocene (~11.7 ka), but some
159 experienced a delayed response (e.g. Adélie Basin, Mac. Robertson Land, Prydz Bay)⁹⁰, or
160 even slightly thickened and advanced during deglaciation (e.g. Transantarctic Mountains)^{97,98}.
161 Furthermore, data from the Lambert-Amery system and Transantarctic Mountains indicate
162 most ice surface lowering occurred during the Early-Mid Holocene, continuing in some
163 locations into the last few millennia⁹⁹⁻¹⁰¹.

164 Bed topography influenced the retreat of marine-based sectors across the inner-
165 continental shelf^{91,102,103} and may explain some regional asynchronicity. Geomorphological
166 evidence on the sea-floor of the Mertz Trough, Prydz Channel and western Ross Sea, for
167 example, indicates marked accelerations in grounding line retreat across over-deepened
168 basins^{90,91,104}. Rapid retreat across over-deepenings in the Ross Sea was also associated with
169 hundreds of metres of ice surface lowering over several centuries^{99,101}. In contrast, retreat was
170 slowed by elevated bed topography¹⁰². Isostatic rebound in the Weddell and Ross Seas may
171 also have exerted a stabilising effect¹⁰⁵, but this process has not been explored elsewhere in
172 East Antarctica.

173 Ice sheet modelling indicates the greatest ice losses occurred in the major marine
174 embayments during deglaciation (e.g. Ross Sea, Weddell Sea and, to a lesser degree, Prydz
175 Bay), with mass loss elsewhere varying in both magnitude and timing depending on the
176 model¹⁰⁶⁻¹⁰⁸. Initial retreat was likely triggered by a combination of ocean, atmosphere and
177 sea-level changes²¹. In at least the Ross Sea, atmospheric conditions controlled the timing
178 and spatial pattern of early deglaciation¹⁰⁹. Meanwhile, terrestrial sectors of the interior EAIS,
179 and areas of the Transantarctic Mountains, likely thickened due to increased snowfall following
180 the Last Glacial Maximum^{97,106}. Oceanic warming became an increasingly important control
181 on marine-based retreat during deglaciation⁹⁰, driving a positive feedback whereby ice loss
182 freshened surface ocean waters, possibly reducing the formation of dense Antarctic bottom
183 waters, and facilitating incursions of modified CDW onto the continental shelf¹⁰⁸. Holocene ice
184 loss in the Ross Sea, for example, corresponds with ocean warming and the development of
185 ice shelf cavities and modified CDW intrusion^{102,109}.

186 Palaeo-data of grounding line retreat and ice sheet thinning during deglaciation provide
187 important context for modern-day observations (Fig. 3). The highest rates of grounding line
188 retreat exceeded 100 m yr⁻¹, possibly up to 800 m yr⁻¹ (Fig. 3a). Deglacial thinning rates from
189 the flanks of outlet glaciers were typically 0.06–0.76 m yr⁻¹, possibly up to several metres per
190 year (Fig. 3c). These time-averaged rates were sustained over several centuries and may
191 mask more extreme rates, but reveal that some present-day thinning (>1 m yr⁻¹) and retreat
192 rates (>200 m yr⁻¹) are comparable to the last deglaciation.

193

194 **4. Recent Ocean Conditions and Ice Dynamics**

195 Evidence and modelling from past warm periods clearly points to the enhanced sensitivity of
196 East Antarctica's three major marine basins (RSB, WSB, ASB). Terrestrial regions respond
197 mainly to atmospheric forcing, but these marine-based sectors respond to both atmospheric
198 and oceanic processes, potentially involving Marine Ice Sheet Instability. Hence, ocean
199 conditions and bathymetry on the continental shelf are important to understand in relation to
200 EAIS dynamics.

201 Around most of East Antarctica, strong easterly winds drive onshore Ekman transport
202 of cold fresh Antarctic Surface Water, yielding a 'fresh shelf' regime (Fig. 4a). This wind-driven
203 flow piles up cold fresh water over the continental shelf, inducing a down-welling circulation
204 and, via geostrophic adjustment, a strong Antarctic Slope Current¹¹⁰. In these locations, weak
205 cross-slope exchange across the Antarctic Slope Current yields a strong front separating cold
206 fresh waters from warm salty CDW offshore (Fig. 4c), limiting CDW intrusion. Elsewhere, a
207 'dense shelf' regime prevails in the Ross Sea, Adélie Coast, and around Prydz Bay (Fig. 4d).
208 Here, the overflow of dense shelf water (also referred to as High Salinity Shelf Water in some
209 sectors) is balanced in part by onshore CDW transport, although strong water-mass
210 transformation over the shelf cools these regions during winter¹¹¹. Poleward Ekman transport
211 of cold fresh surface water still occurs, but the Antarctic Slope Current is weaker and less of
212 a barrier to cross-shelf exchange. Finally, along the coast of Wilkes Land a limited 'warm shelf'
213 regime exists (Fig. 4e), where weaker easterly winds enable modified CDW intrusions closer
214 to the ice margin. Recent evidence of a localised warm shelf regime close to Shirase Glacier
215 (Fig. 1d), Dronning Maud Land, has also been detected¹¹², again enabled by weaker polar
216 easterlies.

217 Observations of shelf-water temperature trends are extremely sparse around East
218 Antarctica, with few multi-decadal measurements available (e.g. in the Ross Sea)^{113,114}.
219 Evidence for long-term shelf-water warming is therefore limited, but warm waters have been
220 detected close to several major outlet glaciers^{26-29,112,115}, coinciding with high basal melt rates

221 beneath ice shelves (Fig. 4b)^{12-15,116}. This can lead to ice shelf thinning and reduced
222 buttressing, increasing ice discharge¹⁸. Warm water entering ice shelf cavities can also form
223 basal channels, causing localised incision and further structural weakening¹¹⁷, including
224 transverse fractures associated with calving¹¹⁸.

225 One region where ocean forcing is impacting ice dynamics is Wilkes Land, overlying
226 the ASB and referred to as East Antarctica's 'weak underbelly'^{30,119}. A signal of mass loss has
227 emerged over the last three decades^{3-9,23}, with one study⁹ suggesting mass loss (-51 ± 80 Gt
228 yr^{-1} : 2016–2020) may be ten times higher than a decade ago. Observations²⁶⁻²⁸ have
229 confirmed modified CDW proximal to Moscow University and Totten glaciers (Fig. 4b). Totten's
230 catchment has been thinning and losing mass since the late-1970s, with numerous studies
231 attributing this to ocean forcing and wind-driven upwelling of modified CDW^{4-6,11,24,27,32-34,50,120-}
232 ¹²³. Its grounding line has been retreating since at least the 1990s^{17,33-34} (Fig. 3a) and, given
233 Totten's large catchment (3.9 m of SLE)³³ and high discharge (~ 70 Gt yr^{-1})⁴, these
234 observations are concerning. However, ice discharge may have slowed recently (2008–
235 2017)^{4,123}, and some variability in basal melting may reflect intrinsic oceanic processes⁵⁰.
236 Furthermore, the grounding line of both Totten and Moscow University glaciers sit on prograde
237 slopes extending 50–60 km up-ice¹⁰, suggesting that imminent Marine Ice Sheet Instability is
238 unlikely.

239 Elsewhere in Wilkes Land, glaciers entering Porpoise Bay have received much less
240 attention, but have experienced pronounced thinning⁵⁻⁷ and are sensitive to buttressing from
241 landfast sea-ice³⁷. Frost Glacier has the largest catchment (0.84 m SLE)⁴ and underwent
242 moderate thinning (< 0.5 m yr^{-1}) over recent decades^{5,7,11}, concomitant with grounding line
243 retreat¹⁷ (> 200 m yr^{-1} from 2010–2016) (Fig. 3b). Holmes Glacier is smaller (0.12 m SLE)⁴, but
244 the surface thinning (> 1 m yr^{-1}) is greater^{7,11}, perhaps driven by enhanced ice shelf basal
245 melting^{12,116} (Fig. 4b). Both glaciers merit monitoring given their large catchments, high
246 discharge, and sensitivity to ocean processes³⁷. Likewise, glaciers draining the ASB into
247 Vincennes Bay are largely unstudied, but lie proximal to some of East Antarctica's warmest
248 intrusions of modified CDW²⁹. Bond and Underwood glaciers have increased in flow speed
249 (2008–2016)^{3,4}; and the grounding line of Vanderford (Fig. 4b) retreated > 800 m yr^{-1} from
250 1996–2017⁴, the highest reported rate for East Antarctica (Fig. 3a).

251 Further west, Denman Glacier (Fig. 4b) holds ~ 1.5 m SLE⁴ in the ASB. Its grounding
252 line sits atop a deep canyon extending $> 3,500$ m below sea level^{10,35}. Both its grounded
253 ($17 \pm 4\%$) and floating ($36 \pm 5\%$) portions accelerated¹²⁴ from 1972–2017, accompanied by
254 surface thinning since at least the 1990s^{5,24,32}. Denman lost a lateral pinning point during its
255 last major calving event (1984)¹²⁴ and, from 1996–2017, the western part of its grounding line
256 retreated 5.4 km along a deep trough^{10,17,35}. Ice shelf melt rates of $> 45 \pm 4$ m yr^{-1} (2011–2014)

257 occur near the grounding line³⁵ (Fig. 4b), comparable to the highest rates in West Antarctica¹².
258 One study⁴ estimated mass loss from Denman's catchment equivalent to 0.5 mm of sea-level
259 rise from 1979–2017, second only to Totten (0.7 mm) in East Antarctica, but the drivers of any
260 imbalance are unclear given large uncertainties in mass input and limited changes in ice
261 discharge^{3,4}.

262 Whilst palaeo-records indicate that the neighbouring WSB retreated during past warm
263 periods, current observations provide limited evidence of change. Cook Glacier has attracted
264 attention due to its large size (~1.6 m SLE)⁴ and proximity to a retrograde bed-slope³¹. Its
265 western outlet lost its ice shelf between 1973 and 1989 and subsequently doubled in speed;
266 while the eastern outlet has accelerated since the 1970s³⁸. Observations reveal ice surface
267 thinning since at least the 1990s^{5,11,120-122}, and a small dynamic imbalance (0.2 mm to SLR:
268 1979–2017)⁴, albeit with large uncertainties. Periodic calving events have occurred at the
269 neighbouring Ninnis Glacier¹²⁵ (Fig. 2b), also deemed vulnerable to Marine Ice Sheet
270 Instability^{10,4}, and at Mertz¹²⁶, but without any dynamic response due to negligible buttressing.
271 Limited evidence of current change in glaciers draining the WSB is consistent with low basal
272 melt rates^{12-14,116} and a dense shelf regime (Fig. 4a), but ice shelf retreat/calving^{30,36,38} in the
273 1940s to 1980s is suggestive of warmer than present conditions. Hence, recent ocean forcing
274 in this region (difficult to measure due to high volumes of sea-ice/mélange), may not capture
275 the full range of inter-decadal variability.

276 East Antarctica's other major marine basin – the RSB – is drained by several large
277 glaciers with retrograde slopes (e.g. Recovery, Bailey, Slessor)¹²⁷ and may be highly
278 vulnerable to future ocean warming⁴⁵, but there is currently no evidence of changes in ice
279 dynamics^{3,4,17}. Elsewhere, few glaciers have been studied in East Antarctica's terrestrial
280 sectors, with no obvious changes reported. In Victoria and Oates Land, for example, numerous
281 glaciers have large, unconfined ice tongues that calve periodically, but with no trends in frontal
282 position or ice velocity since the 1970s^{30,36,125,128}. The large region encompassing Mac.
283 Robertson Land to Dronning Maud Land is characterised by ice sheet thickening^{5,6,11,24,32} due
284 to increased accumulation^{47,48} (Fig. 4a), with some evidence of grounding line advance¹⁷ and
285 most catchments gaining mass^{3,4}. Shirase – the fastest-flowing glacier in East Antarctica
286 (~2,500 m yr⁻¹) – experiences relatively high basal melt rates (7–16 m yr⁻¹)¹¹² and may have
287 thinned in the 1990s^{5,24}, but is currently in balance¹²⁹ or gaining mass⁴.

288 In summary, the vast majority of East Antarctic outlet glaciers show no discernible
289 change in velocity or discharge over recent decades^{3,4}, including those draining two of the
290 three marine basins (WSB, RSB). However, some glaciers draining the ASB in Wilkes Land
291 appear to be losing mass in response to ocean heat forcing, similar to glaciers in the WAIS,
292 and with potential for this to be sustained or even increase.

293

294 5. Recent Surface Mass Balance

295 The large spread in estimates of EAIS mass balance (Fig. 2) is derived largely from
296 uncertainties in mass input (surface mass balance: SMB), rather than ice discharge. The mean
297 annual EAIS SMB (1980-2018) over grounded ice has been estimated at +1,247 Gt yr⁻¹ from
298 MERRA-2 global reanalysis data⁷ and +1,290 Gt yr⁻¹ from the MAR regional atmospheric
299 climate model¹³⁰ (Fig. 5a). The SMB is dominated by snowfall, with other components (rainfall,
300 sublimation, blowing snow erosion/deposition, runoff) at least an order of magnitude
301 smaller¹³¹.

302 While most of East Antarctica is relatively dry with typical (interquartile) annual snowfall
303 ranging from 0.05–0.14 meters water equivalent¹³¹, the area is vast. Thus, atmospheric
304 variability (and snowfall) on time-scales from hours to decades⁵¹ is imprinted on overall mass
305 balance. Indeed, inter-annual variations in SMB (e.g. 1980–2018: $\sigma = 106$ and 91 Gt yr⁻¹ for
306 MERRA-2 and MAR, respectively; Fig. 5a) are comparable to the signal of overall mass
307 change (Fig. 2), highlighting the sensitivity of SMB to short-lived but extreme events and the
308 need for long observational time-series (>10s of years).

309 The absence of an EAIS-wide array of direct snow accumulation observations means
310 that assessments of SMB rely on atmospheric datasets, atmospheric reanalyses and regional
311 climate models. However, the lack of observations available for assimilation into global
312 reanalyses, and the regional climate models forced by those reanalyses, means that the
313 representation of atmospheric circulation over the EAIS is poorly constrained. This contributes
314 to the large spread in SMB estimates, exacerbating uncertainty in overall mass change. A
315 recent comparison of Antarctic SMB in eight regional climate models¹³² found the range in
316 basin-wide SMB varied from ~3% to ~40% of the model mean. Model differences were
317 greatest in the large, dry basins of Adélie and Victoria Land, whilst high accumulation basins
318 (e.g. Wilkes Land) showed more consensus.

319 Although the mean SMB of models varies substantially, they broadly agree on the
320 magnitude of inter-annual variability, as well as on recent trends^{131,133}. Both MERRA-2⁷ and
321 MAR¹³⁰ show no significant change in EAIS SMB over the last 40 years (<0.1% per year from
322 1980–2018: Fig. 5a). Shallow ice/firn core analyses^{133,134} indicate this forms part of a century
323 of no significant change (1901–2000 SMB trend = 0.1 ± 0.4 Gt yr⁻²). Over the period 1800–
324 2000, however, a trend of increased accumulation has been reported¹³³ (0.3 ± 0.1 Gt yr⁻²), but
325 substantial low-frequency variability increases the uncertainty, suggesting it may be
326 insignificant⁵¹. Furthermore, due to their relatively short (<20 years) observational record,
327 altimetry and gravimetry methods do not fully capture these decadal-to-century variations in

328 SMB, which can complicate the separation of SMB and ice dynamical change. Long-term SMB
329 variations, while relatively unconstrained, are also essential to correct altimetry records for
330 changes in firn air content.

331 Recently, ponding of meltwater on East Antarctic ice shelves has received
332 considerable attention¹³⁵⁻¹³⁹ due to its potential role in ice shelf collapse via hydro-
333 fracturing^{44,85,140-142}. Surface meltwater (streams, lakes, slush), found in the grounding zone of
334 numerous East Antarctic ice shelves¹³⁶, indicates insufficient porosity for drainage into the firn.
335 Where firn air content is high, meltwater drains into the firn and may refreeze¹⁴². Firn air
336 content can be approximated by a liquid-to-solid ratio, defined by the amount of surface melt
337 (and rainfall, rare over East Antarctica¹⁴³) divided by snowfall. Although subject to uncertainties
338 (particularly the liquid component in the marginal areas of the ice sheet), liquid-to-solid ratios
339 are relatively easy to compute and are <25% on most East Antarctic ice shelves (Fig. 5c),
340 indicating annual snowfall is >4 times larger than liquid water fluxes and that a porous firn
341 layer (10s m) accommodates storage/refreezing of summer meltwater. Notable exceptions
342 include the grounding zones of Amery Ice Shelf, with liquid-to-solid ratios up to 80% (similar
343 to Antarctic Peninsula ice shelves), and some eastern Dronning Maud Land ice shelves
344 (~40%). These ice shelves support high densities of supraglacial lakes^{136,137,139}, but their
345 physical confinement and thickness (e.g. Amery) protect them from widespread hydrofracture.
346 Indeed, ice shelf collapse via hydrofracturing is critically dependent on stress conditions, with
347 <1% of vulnerable ice shelf areas in East Antarctica currently supporting lakes¹⁴¹.

348 Given that surface melt has not significantly increased in any of East Antarctica's
349 drainage basins over the last 40 years (Fig. 5b) and that snowfall has remained broadly the
350 same, and increased over western East Antarctica, we suggest few ice shelves are currently
351 at risk from hydrofracture. However, climate projections indicate surface melt and rainfall
352 (especially on East Antarctic ice shelves), as well as snowfall (over the entire EAIS), will
353 increase in the next century¹⁴³⁻¹⁴⁹. This will increase the vulnerability of the northernmost ice
354 shelves^{130,142,147,149}, such as West and Shackleton^{130,140,141}. Shackleton, partially fed by
355 Denman Glacier, already hosts high densities of supraglacial lakes^{136,138}, experiences high
356 basal melt rates (Fig. 4b), and is considered most at risk^{141,149}.

357

358 **6. Future Projections**

359 Since the 2013 IPCC report¹⁵⁰, there has been significant progress in understanding the
360 uncertainties associated with modelling future ice sheet response in Antarctica. Here, we focus
361 on projections that partition the EAIS-only sea-level contribution at 2100, 2300 and 2500 (Fig.
362 6).

363 Using the IPCC (2013)¹⁵⁰ method gives median EAIS sea-level contributions of +0.5
364 to +0.8 cm at 2100 (Fig. 6a: 'IPCC 2013 updated'). Here, the dynamic response was
365 extrapolated from observations and does not vary with emissions scenario, and the SMB
366 response to warming was derived from climate models (recalculated here for Shared
367 Socioeconomic Pathways (SSPs), using temperature projections from the IPCC (2021)¹⁵¹).
368 More recent studies generate a wider range of projections with both negative and positive sea-
369 level contributions from the EAIS by 2100, some of which approach +15 cm or more under
370 very high emissions^{43,152,153} (Fig. 6a). The Ice Sheet Model Intercomparison Project (ISMIP6)
371 for the sixth phase of the Coupled Model Intercomparison Project (CMIP6) represents the
372 most comprehensive and up-to-date synthesis of these projections^{25,148,152,154}, using eleven ice
373 sheet models forced by six CMIP5¹⁴⁹ and four CMIP6²⁵ climate models. Experiments include
374 high and low emission scenarios (RCP8.5/SSP5-85 and RCP2.6/SSP1-26, respectively), a
375 range of parameter values governing the sensitivity of ice shelf basal melting to ocean
376 temperatures¹⁵⁵, and various scenarios of ice shelf collapse driven by atmospheric forcing¹⁴⁴.
377 Overall, ISMIP6 gives an EAIS-only sea-level contribution ranging from -7 to +15 cm at 2100
378 (Fig. 6a).

379 A major uncertainty is the balance between the SMB input (influenced by the choice of
380 climate model) and dynamic losses (largely influenced by the choice of basal melt sensitivity
381 to ocean warming). A comparison of ISMIP6 simulations driven by two different global climate
382 models under RCP8.5, for example, can change the sign of overall mass balance (Fig. 6a:
383 'ISMIP6: GCM1' versus 'GCM 2'). A similar effect is seen when comparing ISMIP6 simulations
384 using two distributions of the parameter governing basal melting: one derived from the
385 Antarctic average, and the other from a high-melt region proximal to Pine Island Glacier, WAIS
386 (Fig. 6a: 'ISMIP6: mean melt' versus 'high melt').

387 The ISMIP6 ensemble was unavoidably relatively small (344 simulations from 14
388 modelling groups) and unevenly sampled. Recently, statistical emulation was used¹⁵² to
389 resample the uncertainties, giving median projections from the EAIS at 2100 (+1.5 to +2.6 cm)
390 that are 2-3 cm higher than the original ensemble means (Fig. 6a: 'ISMIP: all' versus 'ISMIP6
391 emulator'). This is partly due to the greater weight given to high basal melt values¹⁵², and partly
392 due to the updated IPCC (2021) projections, which have a mean increase of +1.1 cm arising
393 mostly from the estimated response to pre-2015 climate change¹⁵¹. The influence of the basal
394 melt sensitivity can also be seen in the dynamic-only contributions from the Linear Antarctic
395 Response to basal melting Model Intercomparison Project phase 2 (LARMIP-2)¹⁵⁶,
396 recalculated here with IPCC (2021)¹⁵¹ temperature projections (Fig. 6b). The ISMIP6 emulator
397 projects similar 5th to 95th percentiles to LARMIP-2, but much higher medians (+9 to +10 cm),
398 under a 'risk-averse' scenario¹⁵² (Fig. 6b): a subset of climate and ice sheet models that lead

399 to high mass loss via high basal melting and ice shelf disintegration. The SMB contribution to
400 sea level is not modelled by LARMIP-2 but is expected to be negative, lowering the total sea-
401 level contribution (LARMIP-2's EAIS region is also smaller than other studies). After adding
402 the estimated SMB input (median -2 to -5 cm SLE), the IPCC (2021)¹⁵¹ found that differences
403 between LARMIP-2¹⁵⁶ and the ISMIP6 emulator¹⁵² could largely be explained by different
404 assumptions about basal melt sensitivity, and combined the two for the main assessment with
405 a 'p-box' approach (mean of the two individual medians gives the assessed median; outer
406 edges of the individual 17-83rd percentiles gives the outer edges of the assessed *likely* (17-
407 83%) ranges). Taking the same 'p-box' approach with LARMIP-2 and ISMIP6 here gives the
408 combined median EAIS contributions of +1 to +3 cm by 2100 across all scenarios, with 5th
409 percentiles of -3 to -5 cm, and 95th percentiles of +15 to +17 cm (SSP1-2.6), +16 to +19 cm
410 (SSP2-4.5), and +20 to +25 cm (SSP5-8.5).

411 Even higher projections at 2100 have been made by incorporating the proposed
412 'Marine Ice Cliff Instability' ^{44,85}, which involves the collapse of deep ice cliffs at the grounding
413 line, initiated by ice shelf disintegration (Fig. 6b: lower section). This mechanism has been
414 added to one ice sheet model^{44,85}, motivated by theoretical considerations and observations
415 of ice cliff calving mechanics, and for this model to be able to simulate the highest potential
416 Pliocene sea level contributions (Fig. 1b). The inclusion and representation of an ice-cliff
417 instability are debated^{68,86,151,157-159} and the timing of ice shelf disintegration is highly
418 uncertain¹⁵¹. Indeed, projections using marine ice-cliff instability⁴⁶ are extremely sensitive to
419 warming, with negative contributions under low and medium emissions, but a 95th percentile
420 of +38 cm under very high emissions (Fig. 6b). Expert elicitation¹⁶⁰ does not explicitly define
421 contributing processes, but encompasses the full range of model projections under very high
422 emissions and is narrower for low emissions (Fig. 6b). Both marine ice-cliff instability and
423 expert elicitation were assessed by the IPCC (2021)¹⁵¹ as *low confidence* projections - that
424 could nevertheless not be ruled out - and were combined with the main projections in a p-box
425 approach. Taking a similar approach here gives a *low confidence* 95th percentile of +47 cm
426 SLE from the EAIS at 2100 under SSP5-8.5.

427 Few scenario-dependent EAIS projections are available beyond 2100. Maximum
428 contributions under low and medium emissions are +0.6 m SLE at 2300 (Fig. 6c) and +0.7 m
429 at 2500 (Fig. 6d). This suggests the EAIS contribution to sea-level would be well under +1 m
430 over the next few centuries if emissions follow current Nationally Determined Contributions,
431 which are lower than the medium scenario (SSP2-4.5)¹⁶¹; and less than +0.5 m under low
432 emissions with a median warming of <2°C (SSP1-2.6 95th percentile at 2300 is 2.2°C)¹⁵¹.

433 Under very high emissions, model projections show a wide range, with the EAIS
434 contributing -0.08 to +3.0 m SLE at 2300 (Fig. 6c, upper panel) and +1.0 to +5.4 m at 2500

435 (Fig. 6d), although the upper bounds would halve (1.3 m at 2300 and 2.3 m at 2500) when
436 excluding the study⁴⁴ that the IPCC (2019)⁶⁸ deemed to over-estimate ice shelf collapse. *Low*
437 *confidence* projections using marine ice-cliff instability⁴⁶ and from expert elicitation¹⁶⁰ are even
438 higher (Fig. 6c, lower), with 95th percentiles of +4.7 m and +3.9 m at 2300, respectively,
439 although most of the elicited distribution is far lower (83rd percentile 0.2 m SLE). Taking the
440 IPCC (2021)¹⁵¹ p-box approach to combine these gives a *low confidence* 95th percentile
441 approaching +5 m at 2300. Such high emissions are becoming increasingly less probable, as
442 they would greatly exceed those predicted for Nationally Determined Contributions under the
443 Paris Agreement and other pledges (e.g. net zero emissions by mid-late century)¹⁶¹.

444 Spatial patterns of modelled mass loss consistently highlight the vulnerability of East
445 Antarctica's marine-based sectors, but the magnitude and rate of ice loss is model-dependent.
446 Multi-century simulations^{43,44,46,148,162,163} typically show grounding line retreat and mass loss in
447 the ASB (Fig. 1f), followed by the WSB and RSB, although the latter shows high sensitivity to
448 ocean warming in some studies^{45,163}. Notably, most models do not include recent discoveries¹⁰
449 of over-deepened subglacial topography that might exacerbate Marine Ice Sheet Instability,
450 such as the deep trough connecting Denman Glacier to the ASB. Given that large parts of
451 East Antarctica remain unsurveyed¹⁰, there may be undiscovered over-deepenings upstream
452 of grounding lines or unknown bathymetric troughs with potential to carry warm waters towards
453 the ice margin: both could increase mass loss beyond current expectations.

454 Future ocean conditions will exert a critical influence on ice discharge via basal melting
455 and ice shelf buttressing¹²⁻¹⁸. However, coupled global ocean-climate models do not resolve
456 important processes, such as circulation within sub-ice-shelf cavities, tidal flows and eddies,
457 and gradients across the Antarctic Slope Current. This leaves global models with baseline
458 biases in hydrographic properties over the continental shelf and shelf-break, although multi-
459 model means are more realistic¹⁶⁴. Insights can be gained from examining climate projections
460 in terms of future surface atmospheric warming, sea-ice melt, wind and ocean circulation
461 changes, and shelf-water hydrography. Projected atmospheric circulation changes include an
462 on-going poleward shift and strengthening of the Southern Hemisphere westerly jet across all
463 seasons¹⁶⁵, and a weakening of the coastal easterlies during austral summer and autumn,
464 particularly around East Antarctica¹⁴⁶. These wind changes are likely to weaken the Antarctic
465 Slope Current, enabling enhanced CDW intrusions onto the shelf^{166,167}, particularly around
466 Wilkes Land and west to Prydz Bay. Projected surface warming and the addition of meltwater
467 also enhances vertical stratification over the shelf, reducing or shutting down dense shelf water
468 formation¹⁶⁸ and leaving the deep-shelf waters warmer. Meltwater input from ice shelves may
469 also create a positive feedback, with additional freshening driving sub-ice-shelf warming,
470 leading to further melt^{169,170}, as hypothesised for the last deglaciation¹⁰⁸. Sea-ice loss also

471 reduces albedo over the ocean, driving further warming¹⁷¹ and increasing the vulnerability of
472 outlet glaciers to ice shelf/tongue collapse³⁷. However, climate models have typically struggled
473 to reproduce Antarctic sea-ice trends¹⁷², which are improved when ice shelf melt¹⁷³ and
474 improved representations of sea-ice motion¹⁷⁴ are included.

475

476 **7. Lessons from the Past Inform the Future**

477 Evidence from the palaeo-record and numerical modelling highlight the sensitivity of East
478 Antarctica's major marine basins (the ASB, WSB and RSB) to past warm periods, including
479 significant ice loss during the early to Mid-Miocene (24–14 Ma), and a multi-metre sea-level
480 contribution during the mid-Pliocene (5.3–2.6 Ma), when atmospheric CO₂ concentrations
481 were comparable to present-day. Retreat of the WSB during more recent interglacials (marine
482 isotope stage 11) further highlights its sensitivity to modest warming scenarios (1.5–2°C).
483 During the last deglaciation, however, there were only limited changes to the ASB and WSB,
484 with grounding line retreat focussed in the marine embayments of the Ross and Weddell Seas
485 that connect the WAIS and EAIS (Fig. 1c). Here, retreat has been linked to a positive feedback
486 driven by ocean warming, whereby meltwater input freshened surface waters, facilitating
487 increased incursions of modified CDW onto the continental shelf, continuing into the
488 Holocene¹⁰⁸. This mechanism, coupled with evidence of Marine Ice Sheet Instability across
489 major over-deepenings during deglaciation, illustrates a plausible scenario for destabilising
490 some major East Antarctic outlet glaciers over the next few centuries (e.g. Denman).
491 Furthermore, there are signs that some glaciers draining the ASB in Wilkes Land are currently
492 losing mass, with grounding line retreat and ice surface thinning rates comparable to, and
493 sometimes exceeding, millennial-scale rates of change during the last deglaciation (Fig. 3);
494 and raising the possibility that a longer-term dynamic response to ocean forcing is underway.
495 However, the WSB and RSB currently show limited evidence of change; even though the latter
496 is deemed most vulnerable to future ocean warming⁴⁵ and may already be exposed to periodic
497 intrusions of modified CDW¹⁷⁵ that could increase later this century¹⁷⁶.

498 Current understanding is therefore insufficient to determine if and when specific
499 thresholds of instability might be reached in East Antarctica's three marine-based sectors.
500 Indeed, there is no single EAIS response, or time-scale of response, and estimates of overall
501 mass balance may obscure emerging trends of mass loss from some catchments.
502 Furthermore, recent and future trends in SMB, dominated by snowfall, are subject to extreme
503 inter-annual variability and large uncertainties. These uncertainties, together with limited data
504 to inform models of glacio-isostatic adjustment and corrections for firn air content, lead to
505 satellite-based estimates of EAIS mass balance with large uncertainties.

506 Future work should continue to target early-warning signs of dynamic imbalance in the
507 three major marine basins, such as ice surface thinning propagating upstream from retreating
508 grounding lines, together with ice flow acceleration and ice shelf thinning. There remains an
509 urgent need to understand the sensitivity of basal melting to ocean temperatures, and for more
510 detailed observations of continental shelf bathymetry, bedrock topography proximal to, and
511 up-ice from, current grounding lines, and improved observations of sub-shelf cavities and
512 oceanic processes. These observations should be supplemented with more widespread and
513 systematic palaeo-campaigns on East Antarctic continental shelves to constrain the sensitivity
514 of catchments to past ocean-climate forcing (e.g. the RSB), including rates of change and
515 potential tipping points. Such data can inform numerical modelling, where multi-model and
516 perturbed parameter ensembles are required to improve the robustness of multi-century
517 projections.

518 Despite current uncertainties, surface melt and rainfall (particularly on ice shelves),
519 and snowfall (over the entire EAIS), will increase this century. By combining the 'ISMIP6
520 emulator' with 'LARMIP-2 updated' (Fig. 6b) and an intermediate IPCC (2021) SMB estimate,
521 we find the EAIS makes a small positive contribution to sea level (+2 cm) at 2100, but with a
522 wide range depending on scenario (5th to 95th percentiles: -4 cm to +16-22 cm), and with upper
523 bounds driven by high basal melt sensitivity to warming. *Low confidence* projections, due to
524 limited evidence, reach +0.47 m at 2100 under very high emissions (Fig. 6b). If warming
525 continues beyond 2100, sustained by high emissions, evidence from the palaeo-record, recent
526 observations, and numerical modelling projections (albeit derived from a small number of
527 studies) point to significant potential contributions to global mean sea level from marine-based
528 sectors, reaching ~1-3 m or more by 2300 (Fig. 6c) and ~2-5 m by 2500 (Fig. 6d). Catchments
529 most at risk drain the ASB in Wilkes Land (Frost, Holmes, Totten, Vanderford, Denman,
530 Moscow University), and the WSB in George V Land (Cook, Ninnis), with the RSB also
531 potentially vulnerable. Crucially, if the Paris Agreement to limit warming to well below 2°C
532 above pre-industrial is satisfied, significant mass loss could be reduced or averted (Fig. 6c, d:
533 SSP1-2.6/RCP2.6), with the EAIS sea-level contribution remaining below +0.5 m at 2500.
534 Even under emissions similar to Nationally Determined Contributions which exceed this
535 temperature target (Fig 6c, d: SSP2-4.5/RCP4.5), East Antarctica's contribution to sea-level
536 rise would remain well below +1 m over the coming centuries. The fate of the world's largest
537 ice sheet remains very much in our hands.

538

539

540 **Author Contributions:**

541 CRS developed the idea for the paper and all authors provided input on its initial contents and
542 structure. CRS drafted Section 1. GJGP and SSRJ drafted Section 2, with contributions from
543 MJB and TvdF. RSJ drafted Section 3 with contributions from MJB. CRS and BWJM drafted
544 Section 4, with contributions from MHE and AF. JTML and BM drafted Section 5 with input
545 from MAK. CR and TLE drafted Section 6, with contributions from MHE. CRS drafted Section
546 7 with input from TLE. All authors provided comments and edits on all sections of the paper.
547 GJGP produced Fig. 1, with input from CRS. PLW produced Fig. 2 with input from CRS, MAK
548 and RSJ. RSJ produced Fig. 3, with input from BWJM and CRS. AF, MHE and BWJM
549 produced Fig. 4. JTML and BM produced Fig. 5. TLE carried out the analysis and produced
550 Fig. 6 with input from CR.

551

552 **Competing Interests:**

553 The authors declare no competing interests.

554

555 **References:**

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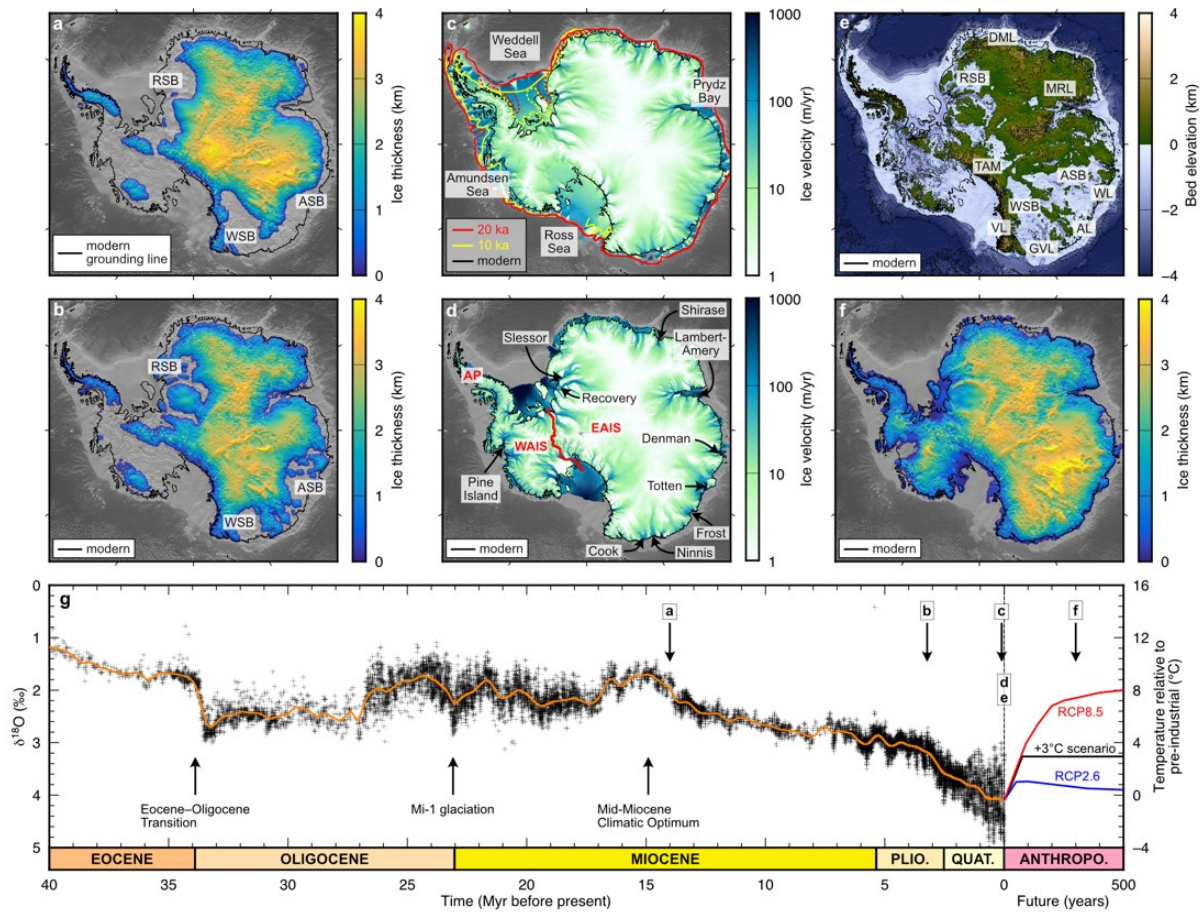
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977 **Acknowledgements:**

978 CRS, BWJM and SSRJ acknowledge funding from the Natural Environment Research Council
979 (NE/R000824/1). MAK, NJA, and MHE are supported by the Australian Research Council
980 Special Research Initiative, Australian Centre for Excellence in Antarctic Science (Project
981 Number SR200100008), and RSJ is supported by the Special Research Initiative, Securing
982 Antarctica’s Environmental Future (SR200100005). NJA (FT160100029), MHE
983 (DP190100494, LP200100406) and RSJ (DE210101923) also acknowledge funding from the
984 Australian Research Council. MHE and MAK also acknowledge support from the Centre for
985 Southern Hemisphere Oceans Research (CSHOR), a joint research centre between QNLM,
986 CSIRO, UNSW and UTAS. AF was supported by the Australian Antarctic Program Partnership
987 through funding from the Australian Government as part of the Antarctic Science Collaboration
988 Initiative. TLE was supported by the UK Natural Environment Research Council
989 (NE/T007443/1) and by the European Union’s Horizon 2020 research and innovation
990 programme under grant agreement No 869304, PROTECT contribution number 36. JTML
991 acknowledges support from the National Aeronautics and Space Administration (NASA),
992 award #80NSSC20K1123. MHE and AF thank Steve Rintoul for discussions on ocean data
993 coverage around East Antarctica. TLE thanks Gregory Garner and Robert Kopp for help with

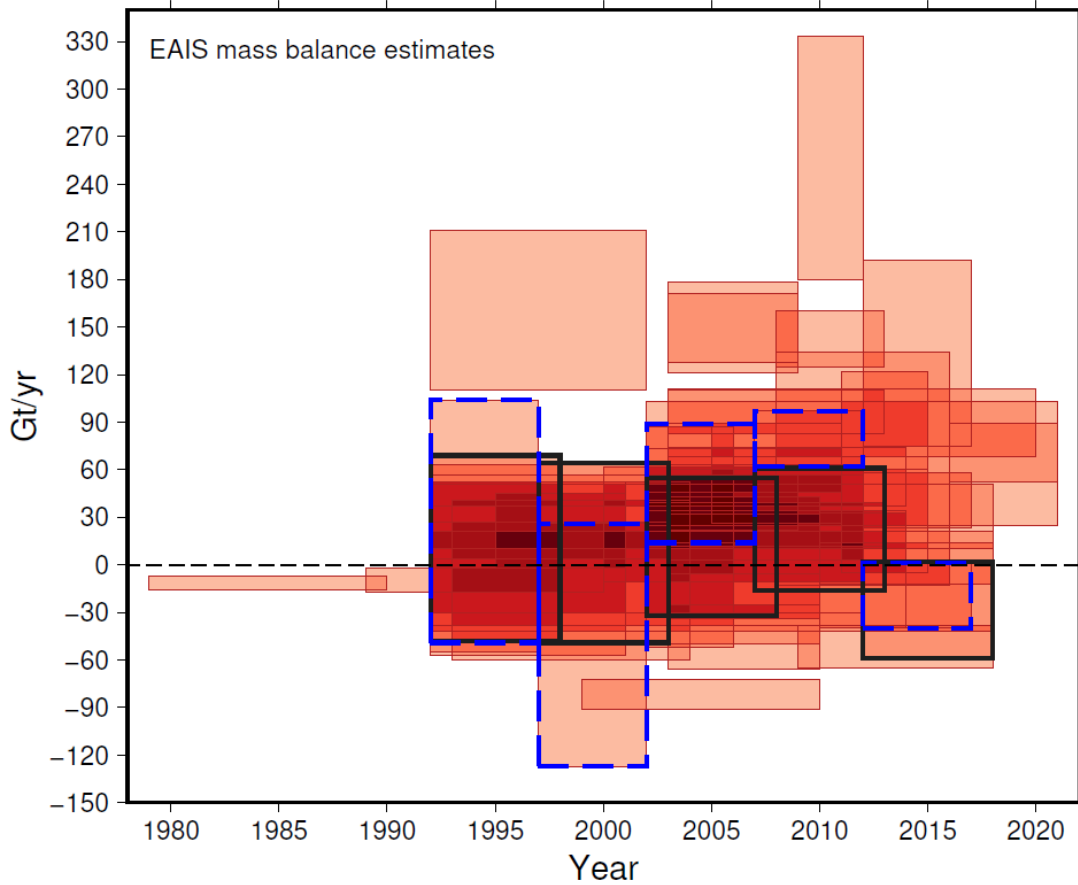
994 IPCC (2021) datasets. We thank the Editor, Michael White, together with Johann Klages, Ed
995 Gasson and three anonymous referees, who provided constructive reviews of the manuscript.
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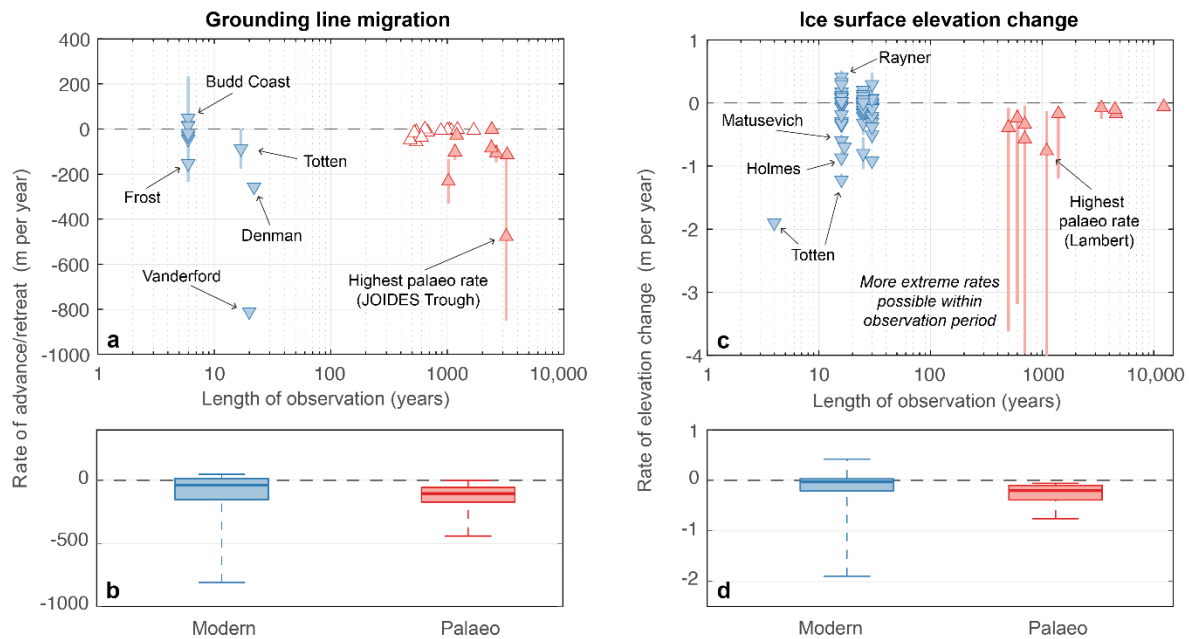
1000 **Figure 1: Grounding line extent and characteristics of the East Antarctic Ice Sheet (EAIS) at**
 1001 **selected times in the past, present and future. (a) Modelled ice thickness during the Mid-**
 1002 **Miocene**⁵⁷ and reconstructed Mid-Miocene palaeotopography¹⁷⁷ in greyscale, showing deglaciation of
 1003 West Antarctica and East Antarctica’s three major subglacial basins: the Recovery (RSB), Wilkes
 1004 (WSB) and Aurora (ASB). **(b) Modelled ice thickness during a warm mid-Pliocene interglacial**
 1005 **with hydrofracturing and ice-cliff calving physics enabled**⁴⁶ and reconstructed mid-Pliocene
 1006 palaeotopography¹⁷⁷ in greyscale. **(c) Modelled Last Glacial Maximum (20 ka) ice surface velocities**
 1007 **from a Parallel Ice Sheet Model ensemble best-fit reference simulation**¹⁷⁸ and RAISED consortium
 1008 **grounding lines at 20 ka (red) and 10 ka (yellow) inferred from empirical data**¹⁷⁹ (dashed lines depict a
 1009 RAISED scenario in the Weddell Sea that is now considered less likely⁹⁴). **(d) Present-day ice surface**
 1010 **velocities** derived from interferometric SAR phase mapping¹⁸⁰, with selected outlet glaciers labelled
 1011 together with EAIS, West Antarctic Ice Sheet (WAIS) and Antarctic Peninsula (AP). Note that we use
 1012 the standard definition of the EAIS as Antarctic drainage basin numbers 2-17 (e.g. refs ^{1, 24}). **(e)**
 1013 **Present-day Antarctic bed topography and Southern Ocean bathymetry** from BedMachine¹⁰ (AL =
 1014 Adélie Land; DML = Dronning Maud Land; GVL = George V Land; MRL = Mac. Robertson Land; TAM
 1015 = Transantarctic Mountains; WL = Wilkes Land; VL = Victoria Land). **(f) Modelled ice thickness at**
 1016 **2300 under a 3°C warming scenario**⁴⁶. **(g) Global benthic oxygen isotope curve through the**
 1017 **Cenozoic**¹⁸¹ with a 1 Myr-smoothed trend line (orange). The projected temperature of the end-member
 1018 RCP2.6 (blue) and RCP8.5 (red) future emissions scenarios are displayed to the year 2500. The ice
 1019 configurations shown in panels a–f are labelled along the timescale.

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1022 **Figure 2: Published estimates of the net mass balance of the East Antarctic Ice Sheet.** Each box
 1023 represents a single estimate of net mass balance with overlapping estimates indicated by darker
 1024 shading. The horizontal extent of each box represents the survey period. Most studies provide annual
 1025 data plotted from 1st January to 31st December for any given year. The vertical extent of each box
 1026 represents the stated uncertainties. Survey areas may vary slightly between different studies, but only
 1027 those that partition the net mass balance of the EAIS or a large part thereof are included (see [Source](#)
 1028 [Data](#) file for numeric values and references). Two recent attempts to reconcile data from multiple
 1029 methods are highlighted in black¹ and dashed blue² lines.
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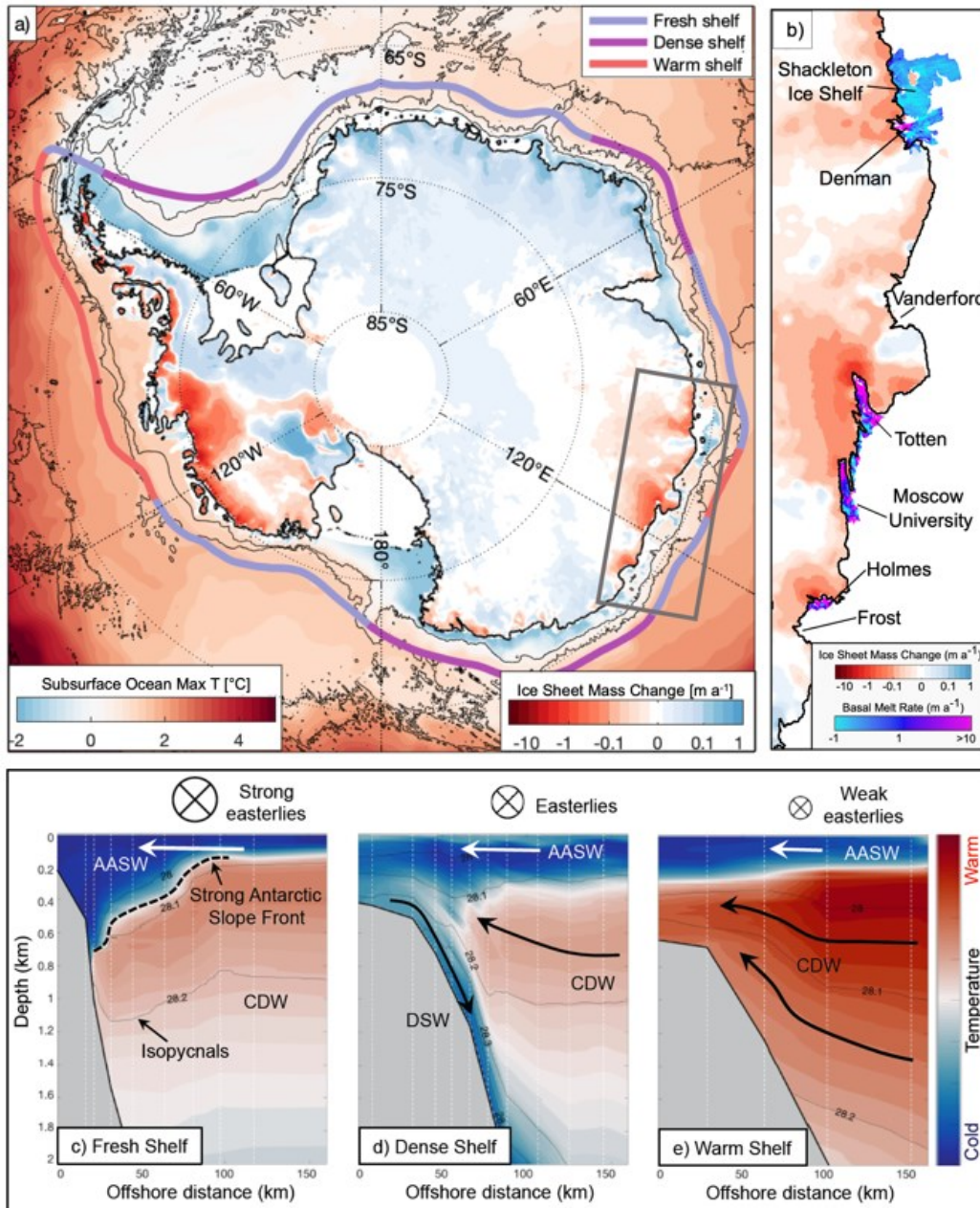
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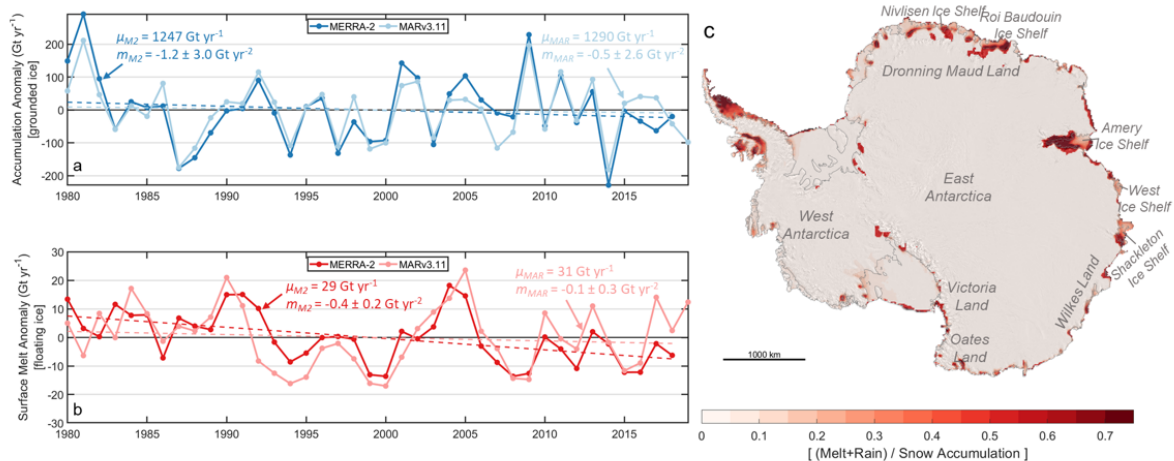
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Figure 3: Comparison between published estimates of modern and palaeo (last deglaciation) rates of grounding line migration and ice surface elevation change. (a) Rates of grounding line advance (positive) and retreat (negative) for modern (blue) and palaeo (red) estimates plotted against length of observations. For modern estimates, the triangle marker denotes the mean and the vertical line extends to the maximum possible advance or retreat value quoted in the study. For palaeo estimates, the triangle represents the mean and the vertical line represents the minimum-maximum range, where available. White triangles are minimum palaeo estimates based on the grounding line reaching its present-day position zero years ago. (b) Box and whisker plots for the range of modern and palaeo mean estimates of grounding line migration. The median and interquartile range is represented by the horizontal line and the box extent, respectively, while the range is shown by the vertical dashed line. (c) Rates of ice sheet thickening (positive) and thinning (negative) for modern (blue) and palaeo (red) estimates plotted against the length of observations. Modern rates from selected East Antarctic outlet glaciers are mean rates extracted from a 20 km x 20 km box immediately up-ice of the grounding line from three recently published altimetry studies⁵⁻⁷. Triangle markers and vertical lines represent the mean and published uncertainty range for the modern estimates, and the median and 95% confidence range for the palaeo estimates. (d) Box and whisker plots for the range of modern (mean) and palaeo (median) estimates of ice surface elevation change (as in 'c'). See [Source Data](#) for numeric values, uncertainties and references. Note that all palaeo-estimates are time-averaged rates for the period of observation and actual rates could have been lower/higher within the period.



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Figure 4: Modern oceanic conditions and characteristic shelf/slope regimes around East Antarctica in relation to recent ice sheet mass changes. (a) Oceanic colours show the 2005-2010 mean subsurface ocean potential temperature maximum from the Southern Ocean State Estimate¹⁸². Black lines indicate isobaths from ETOPO2v2 (ref. ¹⁸³), contoured every 2000 m from the 1000-m isobath; thick black line is the Antarctic continental coast. The thick coloured line parallel to the coast differentiates the three main oceanic shelf regimes (fresh shelf, dense shelf, warm shelf: from ref. ¹¹⁰). Continental colours represent data from a recent altimetry study⁷ of ice sheet elevation change (2003-2019), corrected for firm air content to reflect mass change. (b) Firn Air Content-corrected ice elevation change in Wilkes Land, as in (a) with location shown, together with time-averaged ice shelf basal melt rates (2010-2018) from ref. ¹¹⁶. Note the correspondence between mass loss and high basal melt rates. (c, d, e) Schematic latitude-depth transects indicating typical winds, subsurface ocean circulation, temperature and density structure in a (c) fresh shelf, (d) dense shelf and (e) warm shelf regime (modified from ref. ¹¹⁰). Colours represent temperature and black contours isopycnals of neutral density, with the bold black dashed line in (c) indicating the sharp density gradient across the Antarctic Slope Front. Cross-slope circulation is shown schematically with black and white arrows, and wind direction and strength by arrow tails going into the page. Water masses shown include Antarctic Surface Water (AASW), Circumpolar Deep Water (CDW), and Dense Shelf Water (DSW, also referred to High Salinity Shelf Water in some sectors).



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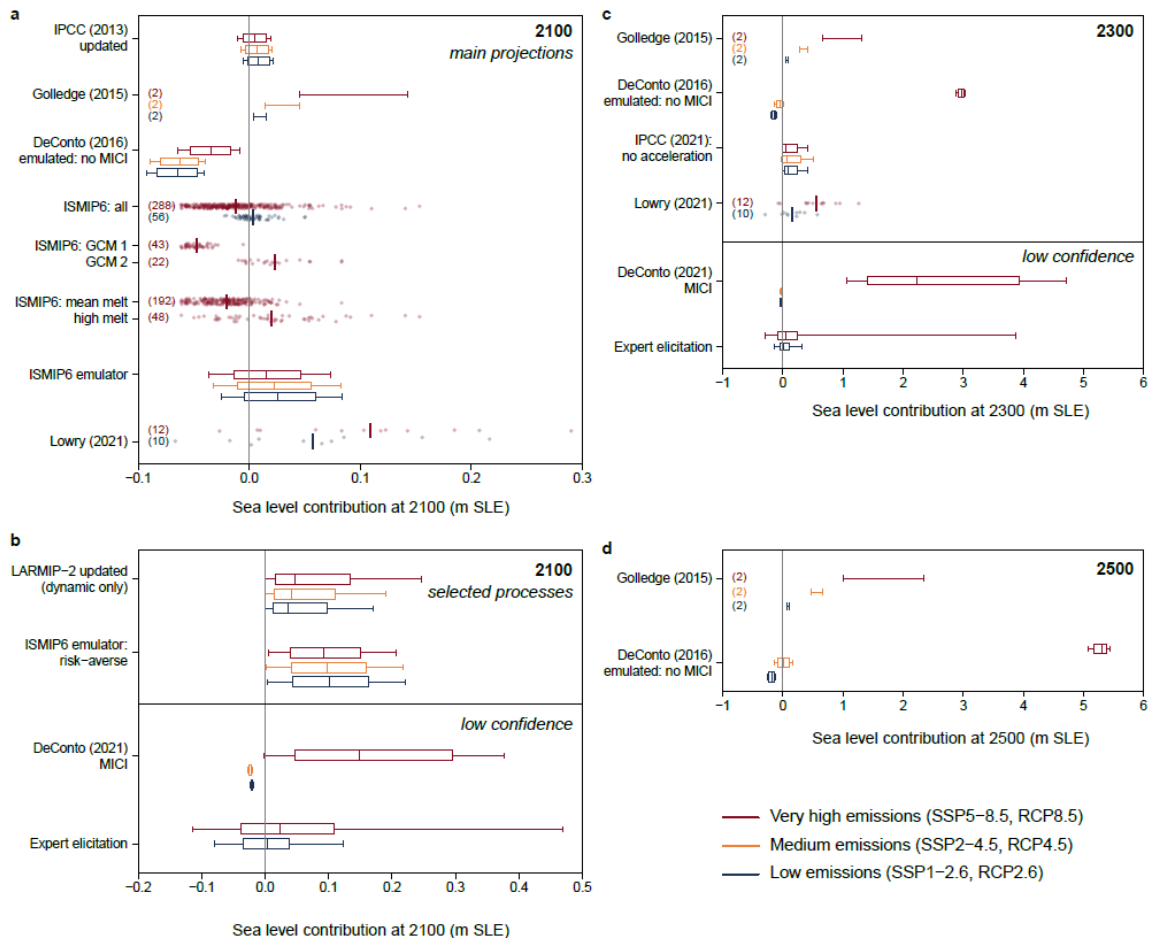
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Figure 5: Recent temporal and spatial trends in Antarctic snow accumulation and surface melt. (a) Annual snow accumulation rates integrated over the entire grounded EAIS expressed as an anomaly from the 39-year mean (1980–2018; $\mu_{M2} = 1,247 \text{ Gt yr}^{-1}$; $\mu_{MAR} = 1,290 \text{ Gt yr}^{-1}$; from ref. ⁷ and ¹³⁰, respectively). The 1980–2018 trends are displayed as dashed lines. (b) As in (a) but for annual surface melt rates over floating ice only ($\mu_{M2} = 29 \text{ Gt yr}^{-1}$; $\mu_{MAR} = 31 \text{ Gt yr}^{-1}$). (c) The average liquid-to-solid ratio from MERRA-2⁷ and MAR¹³⁰ over both the WAIS and EAIS (grounded and floating), where values approaching zero reflect areas of a thick, porous firn column capable of storing liquid water and those approaching one reflect areas with little to no pore space. See [Source Data](#) file for numeric values.



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Figure 6: Projected sea level contribution from the East Antarctic Ice Sheet at 2100, 2300 and 2500 under very high, medium and low emissions scenarios. (a) Projections at 2100 from: IPCC (2013) method, re-estimated for IPCC (2021)¹⁵¹; ref. ⁴³; emulated estimate of ref. ⁴⁴ without Marine Ice Cliff Instability (MICI) mechanism using method of ref.⁸⁶; ISMIP6 multi-model ensemble^{25,148,152}; subsets of ISMIP6 ensemble using climate forcing from two different Global Climate Models (CCSM4 and HadGEM2-ES), with mean sub-ice shelf melting; subsets of ISMIP6 ensemble using mean versus high sub-ice shelf melting treatment; emulated ISMIP6 projections¹⁵² re-estimated for IPCC (2021)¹⁵¹, including the addition of 0.09 mm/yr response to pre-2015 climate change; ref.¹⁵³, subtracting control simulation and adding the same pre-2015 response. **(b) Projections at 2100 for selected processes and 'low confidence' projections** from: LARMIP-2 dynamic-only contribution¹⁵⁶ re-estimated for IPCC (2021)¹⁵¹; emulated ISMIP6 'risk-averse' projections¹⁵² using a high sea-level subset of models and parameter values, with +1.1 cm contribution added to approximate re-estimation for IPCC (2021)¹⁵¹; with MICI enabled⁴⁶; expert elicitation¹⁶⁰. **(c) Projections at 2300** from: ref. ⁴³; emulated estimate of ref. ⁴⁴ without MICI, using method of ref. ⁸⁶; p-box of IPCC (2013) method¹⁵⁰ and dynamic-only contribution¹⁵⁶, extrapolated beyond 2100 with fixed rate mass loss from IPCC (2021)¹⁵¹; ref. ¹⁵³, subtracting control simulation; with MICI⁴⁶; expert elicitation¹⁶⁰. **(d) Projections at 2500** from: ref. ⁴³; emulated estimate of ref. ⁴⁴ without MICI using method of ref. ⁸⁶. Small dots show individual simulations, with short vertical lines showing ensemble means; whiskers without box show range of two simulations. Numbers of simulations are given in brackets. Central line and whiskers show median and 5-95% range; box shows 16%-84% for ref. ⁴⁴ or 17-83% otherwise. All relative to 1995-2014 baseline¹⁵¹ except for refs ^{43,44}, relative to 2000; ISMIP6 ensemble, relative to 2015; and ref. ¹⁵³ for 2105 and 2301, relative to 2025. All use identical climate forcing under Shared Socioeconomic Pathways (SSPs) from IPCC (2021)¹⁵¹, except for refs ^{43,44,46}, forced with regional climate model (RegCM3) under Representative Concentration Pathways (RCPs); ISMIP6 simulations, forced with various global climate models under RCPs and SSPs; IPCC (2021) no acceleration¹⁵¹, which has no climate dependence beyond 2100; and expert elicitation, where warming scenarios are interpreted as SSPs following ref.¹⁵¹. See [Source Data](#) file for numeric values.