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1 **Ice sheet mass balance and climate change**

2

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4 Edward Hanna¹, Francisco J. Navarro², Frank Pattyn³, Catia M. Domingues⁴, Xavier
5 Fettweis⁵, Erik R. Ivins⁶, Robert J. Nicholls⁷, Catherine Ritz⁸, Ben Smith⁹, Slawek Tulaczyk¹⁰,
6 Pippa L. Whitehouse¹¹, H. Jay Zwally¹²

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34 ¹Department of Geography, University of Sheffield, UK

35 ²Universidad Politécnica de Madrid, Spain

36 ³Laboratoire de Glaciologie, Université Libre de Bruxelles, Belgium

37 ⁴Antarctic Climate and Ecosystems Cooperative Research Centre, Tasmania, Australia

38 ⁵Department of Geography, University of Liège, Belgium

39 ⁶Jet Propulsion Laboratory, California Institute of Technology, Pasadena, USA

40 ⁷Engineering and the Environment, University of Southampton, UK

41 ⁸Laboratoire de Glaciologie et Géophysique de l'Environnement, UJF – Grenoble 1 / CNRS,
42 France

43 ⁹Polar Science Center, Applied Physics Laboratory, University of Washington, USA

44 ¹⁰University of California-Santa Cruz, USA

45 ¹¹Department of Geography, Durham University, UK

46 ¹²NASA Goddard Space Flight Center, Cryospheric Sciences Laboratory, Greenbelt, USA

47 **Preface**

48

49 Since the 2007 Intergovernmental Panel on Climate Change Fourth Assessment Report
50 (IPCC AR4), both new observations of ice-sheet mass balance and improved computer
51 simulations of ice-sheet response to ongoing climate change have been published. While
52 Greenland is losing mass at an increasing pace, Antarctic loss is likely to be less than some
53 recently-published estimates. It remains unclear whether East Antarctica has been gaining
54 or losing mass over the last twenty years, and uncertainties in mass change for West
55 Antarctica and the Antarctic Peninsula remain large. We highlight the last six years of
56 progress and examine the key problems that remain.

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70 **1.0 Introduction**

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72 This review aims to synthesize key advances in monitoring and modelling of ice-sheet mass
73 balance since the IPCC AR4¹. Mass balance is defined as the net result of mass gains
74 (primarily snow accumulation) and mass losses (primarily melt-water runoff and solid ice
75 dynamical discharge across the grounding line). Surface mass balance (SMB) is the net
76 balance of mass gains and losses at the ice-sheet surface and does not include dynamical
77 mass loss. Efforts to determine ice-sheet mass balance using the three satellite geodetic
78 techniques of altimetry, interferometry, and gravimetry (see Section 2.1) have recently
79 been sharpened by carefully defining common spatial and temporal domains for inter-
80 comparison². Here we review the latest mass balance estimates for the Antarctic (AIS) and
81 Greenland (GrIS) ice sheets. New glacial isostatic adjustment (GIA) models, tested and
82 evaluated against Global Positioning System (GPS) data, have recently led to significant
83 downwards revision in GIA, and hence downwards revisions of gravimetric and altimetric
84 satellite estimates of Antarctic mass loss² (Box 1).

85 Since IPCC AR4, ice-sheet models are no longer constrained to using overly
86 simplified physics, allowing them to more accurately simulate the important coupling
87 between ice sheets, ice streams and ice shelves. This major advance has been accompanied
88 by improved model representation of the complex interactions of the ice-sheet with its bed,
89 the atmosphere and the ocean. For completeness we also discuss briefly the contributions
90 to sea-level rise (SLR) from other sources, namely glaciers and ice caps, thermal expansion
91 of the oceans and terrestrial water storage changes. Despite recent advances, improved
92 observations and predictions of ice-sheet response to climate change are as urgently

93 needed to feed into mitigation and adaptation models of ensuing SLR as they were at the
94 time of the AR4.

95

96 **2.0 Recent changes in ice sheet mass balance**

97

98 2.1 Comparison of mass balance estimates

99

100 One of the most sought-after but elusive goals in contemporary Earth sciences is to relate
101 the mass-balance state of the great ice sheets to observed SLR. A measure of this state
102 provides an unambiguous quantification of the ice-sheet system response to climate
103 change. Recent mass-change estimates have been derived from three categories of
104 techniques:

105 -Volumetric techniques determine changes in the volume of the ice sheet via
106 measurements of the height of the ice-sheet surface. These are based on radar altimetry^{3,4}
107 or laser altimetry⁵.

108 -Space gravimetric techniques derive changes in ice-sheet mass via repeated and
109 very accurate measurement of the Earth's gravity field by the Gravity Recovery and Climate
110 Experiment (GRACE) satellite system⁶.

111 -The mass budget technique compares estimates of the net ice accumulation on the
112 ice sheets with estimates of discharge across the grounding line⁷ (Box 2).

113 Each estimate relies on observational data that are unique to its own strategy, and
114 each strategy, therefore, has a unique set of sensitivities to the errors and biases in its data.

115 For example, mass budget^{7,8} studies use modelled snowfall fields from atmospheric

116 reanalysis data^{9,10} to estimate the mass input into glacier basins, while radar and laser
117 altimetry studies use the same fields to estimate the effective density of measured volume
118 changes. Thus mass budget estimates have a first-order sensitivity to errors in the
119 modelled mean accumulation rate, while radar and laser altimetry estimates have only
120 limited sensitivity to errors in fluctuations in the accumulation rate.

121 Similarly, GRACE and radar and laser altimetry studies require the effects of GIA-
122 related vertical bedrock motion (Box 1) to be accurately removed. Such vertical motion
123 could be misinterpreted as ice-mass change by the GRACE satellites or ice-thickness change
124 by radar and laser altimeters, and a GIA correction must therefore be applied. This
125 correction is a small percentage (~5%) of the total elevation change typically measured by
126 altimeters; however, the GIA correction applied to GRACE data can be of the same order of
127 magnitude as the signal due to contemporary ice-mass change (because of the density
128 contrast between ice and the solid Earth). As a result, ambiguities in the GIA correction
129 dominate GRACE sources of error in Antarctica (this is not as much of a problem for
130 Greenland where the GIA correction is a much smaller fraction of the total mass change)⁶.
131 Accurate quantification of the GIA signal is therefore crucial; small differences between
132 models can alter the sign of the ice-mass change deduced from GRACE for individual
133 drainage basins¹¹.

134 Published estimates of rates of Greenland and Antarctic ice-sheet mass change
135 obtained using the above methods show a large spread of values for the last two decades
136 (Figure 1). Some of this spread is due to technical differences and some is due to different
137 measurement epochs. However, in the last year, estimates have begun to give a more
138 coherent picture for both Antarctica and Greenland. For Greenland, the trend of increasing

139 mass loss [due to both SMB decrease and ice-to-ocean discharge increase] is clear, while
140 some of the large mass loss estimates for Antarctica have been discarded. We describe
141 some of the improvements in techniques and analysis below.

142

143 2.2 Reduced uncertainties

144

145 Recent assessments of mass-balance history^{12,13}, coupled with more robust GPS
146 observations of the motion of exposed bedrock¹⁴, strongly suggest that Antarctic GIA-
147 related bedrock motion peaks at about 5-6 mm yr⁻¹. The resulting GIA models for
148 Antarctica^{13,15} deliver less than half the mass corrections implied by previous models. At
149 the same time, processing of GRACE data has become more consistent between groups as
150 the time series lengthens. Estimates using the latest models show moderate, if increasing,
151 decadal mass losses for Antarctica^{13,16,17}.

152 In the IMBIE (Ice-sheet Mass Balance Inter-comparison Exercise) project,
153 researchers recently compiled average sets of mass-balance estimates for common time
154 periods for both the Antarctic and Greenland ice sheets, using the latest data, with multiple
155 groups deriving estimates with each technique². An important technical change helped
156 reduce the difference among techniques: unlike previously published mass budget
157 estimates that extrapolated mass changes from surveyed to unsurveyed basins, the IMBIE
158 mass budget estimates use radar altimetry data to demonstrate that unsurveyed areas have
159 near-zero rates of mass change, giving, on average, less mass loss. Other extrapolation
160 techniques can give a more positive Antarctic balance for the same data¹⁸. Similarly,
161 including the most recent GIA estimates for Antarctica brought GRACE estimates closer to

162 the radar and laser altimetry estimates. The IMBIE estimates are simple averages of all
163 measurements, and the discordancy that remains among methods (between radar and
164 laser altimetry, for example) is not fully understood.

165 Figure 1 shows that the disparity of recent mass balance results among different
166 techniques – primarily from IMBIE – is considerably reduced from that seen before. There
167 tend to be systematic differences between the results from different techniques, with the
168 mass budget method giving the most negative estimate for both ice sheets, laser altimetry
169 the most positive, and GRACE in between. IMBIE radar altimetry estimates cover only the
170 sub-peninsular part of Antarctica, and give rates of mass change consistent with those from
171 GRACE. The techniques agree in sign, and roughly in magnitude, for Greenland, and there is
172 considerable basin scale spatial fidelity revealed in the inter-comparisons. Greenland had
173 small contributions to SLR in the 1990s ($-51 \pm 65 \text{ Gt yr}^{-1}$) but was recently (2005-10) losing
174 mass at $-263 \pm 30 \text{ Gt yr}^{-1}$. ($362.5 \text{ Gt yr}^{-1} = 1 \text{ mm yr}^{-1}$ sea-level equivalent.) The situation for
175 Antarctica is less clear, with one estimate showing a significant positive mass balance¹⁹. An
176 unweighted average of the estimates indicates that Antarctica, which was in a state of
177 weakly negative balance in the 1990s, is now losing mass at a rate between -45 and -120 Gt
178 yr^{-1} , with large dynamic losses in West Antarctica partially offset by SMB gains in East
179 Antarctica.

180 For Greenland, an independent group of researchers compared laser altimetry, mass
181 budget, and GRACE estimates over the 2003-09 ICESat period: the mass budget estimate
182 gave the maximum loss rates at $-260 \pm 53 \text{ Gt yr}^{-1}$ and GRACE the minimum, at $-238 \pm 29 \text{ Gt yr}^{-1}$
183 ¹²⁰. On a basin-by-basin basis, agreement between the mass budget method and other
184 techniques provides validation for the practice of partitioning mass-balance change

185 between discharge and SMB components, demonstrating that in the northern part of
186 Greenland, the dominant cause of mass change was atmospheric in origin, while in the
187 southern part it was ice dynamics.

188 The new, reconciled IMBIE GRACE estimates of whole Antarctic mass balance are
189 now largely in agreement with one another, with 30-50 Gt yr⁻¹ spreads between the largest
190 and smallest 2003-08 rates. Previously published GRACE values show spreads around
191 twice as large for similar time periods. In the Antarctic Peninsula and West Antarctica, the
192 IMBIE estimates from laser altimetry and GRACE are in good agreement, in contrast to East
193 Antarctica². For East Antarctica, a mass gain of +101 Gt yr⁻¹ for 2003-2008 has been
194 proposed recently based on laser altimetry¹⁹, which is larger than the IMBIE GRACE
195 estimate of +35 Gt yr⁻¹ and near the upper end of the laser altimetry estimates².

196

197 **3.0 Recent advances in ice sheet modelling**

198

199 3.1 Key improvements and future challenges

200

201 Significant improvements in ice-sheet modelling have been made since the IPCC AR4,
202 motivated by the need to understand ongoing changes and by the challenge to make more
203 realistic projections for the next few centuries. The primary improvements concern
204 mechanical approximations made to the ice flow equations. The very first generation of ice-
205 sheet models was based on the shallow ice approximation²¹. Such models assume that all
206 resistance to flow is provided by shear-stress gradients in the vertical, which is valid for
207 creeping ice-sheet flow, but not when other ice-dynamical features such as ice streams and

208 ice sheet/ice shelf coupling come into play in ice-sheet evolution. More recent ice-sheet
209 models now include horizontal stress gradients, and can be classified into three categories
210 of increasing complexity and computational cost. Ice shelf/stream models are based on the
211 shallow-shelf approximation²². They include horizontal stress gradients, but neglect the
212 vertical shear stresses (which is valid for rapid ice flow at low basal traction). Hybrid
213 models use some combination of solutions from the shallow-ice approximation (to account
214 for the vertical shearing component of flow within grounded ice) and the shallow-shelf
215 approximation (to account for the horizontal stress coupling taking place in ice shelves or
216 regions of rapid sliding)^{21,23,24}. More elaborate higher-order models treat the vertical
217 dimension more rigorously, with the only approximation being the hydrostatic assumption
218 (pressure at any point in the ice is due only to the weight of the ice above it and not due to
219 ice flow)^{25,26}. Finally, a few models solve the equations of motion without neglecting any
220 terms. These models, called “Full Stokes”, have recently demonstrated their ability to
221 perform century-timescale simulations applied to a whole ice sheet^{27,28}.

222 Spatial resolution of models is the second aspect that has been improved. Hardly any
223 model is now run with a spatial grid size greater than 20 km, but this resolution is still not
224 high enough to resolve ice streams, which are often only a few kilometres wide. Moreover,
225 grounding line migration and calving require sub-kilometre resolution. Unstructured grids
226 (for finite element models^{27,28}) or adaptive mesh refinement²⁹ are two strategies that have
227 proven efficient at treating this difficulty with acceptable computational cost.

228 Finally, a third improvement has been enabled through satellite and ground-based
229 observations, such as the quantification of surface velocities and velocity change from
230 satellite interferometry³⁰, surface elevation change through satellite and airborne

231 campaigns (IceBridge), and high-resolution bedrock and ice thickness measurements³¹.
232 Ice-sheet model behaviour is highly dependent on initial and boundary conditions and face
233 the difficulty that drag at the ice-bed interface is poorly known. Inverse methods have now
234 been successfully implemented in ice-sheet models to infer the basal drag map that
235 provides a good agreement between observed and simulated surface velocities. This
236 procedure is becoming standard in the spin-up that is required for establishing an
237 optimum initial state^{27-29,32}. All the above refinements enable models to reproduce present-
238 day observed ice-sheet flow speeds, which is a major improvement since AR4.

239

240 3.2 Grounding lines, sliding and calving

241

242 Warming-induced ice-shelf loss has caused major glaciers and ice streams of Antarctica to
243 speed up^{33,34}. Mechanisms behind this speedup are complex. Oceanic and/or atmospheric
244 warming leads to ice-shelf thinning or disintegration^{35,36}, which in turn may lead to loss of
245 buttressing³⁷, grounding line retreat and hence glacier speedup³³ (Box 2). Observations
246 from the Antarctic Peninsula and the Amundsen Sea Embayment in West Antarctica, e.g.
247 Pine Island and Thwaites Glaciers, which are currently the main contributors of the AIS to
248 SLR³⁸, support these mechanisms.

249 Major theoretical advances²² in understanding grounding-line motion and stability
250 show that in the absence of buttressing (see Box 2), grounding lines retreat unstably on an
251 upward-sloping bed (in the direction of ice flow). Analytical solutions are now available to
252 test and verify marine ice-sheet models, so that the numerical error associated with
253 predicting grounding-line motion can be reduced significantly to the level of parameter

254 uncertainties³⁹: models that attempt to account for grounding-line dynamics should
255 incorporate horizontal stress transmission across the grounding line, so that the grounded
256 ice sheet realistically feels the influence of floating ice (Box 2). Furthermore, the grounding
257 line needs to be resolved at a sufficiently high spatial resolution³⁹. Such developments have
258 been made recently and applied to Pine Island Glacier, where a small increase in sub-ice
259 shelf melting has been shown to result in either unstable grounding line retreat²⁹, or
260 grounding-line stabilization approximately 25 km inland within 100 years³⁷.

261 GIA also influences ice-sheet behaviour^{40,41}. Effects such as Earth deformation in
262 response to ocean loading, and perturbations to the shape of the sea surface in response to
263 the redistribution of both internal and surface masses, including changes to the mass of the
264 ice sheet itself, play a key role in governing the behaviour of a marine-grounded ice sheet,
265 such as West Antarctica⁴². GIA alters the water depth via spatially-varying perturbations to
266 both the ocean floor and the sea surface and this has a first-order effect on grounding line
267 positions²². Ignoring such processes can fundamentally alter model predictions relating to
268 the stability of a marine-grounded ice sheet⁴¹.

269 Ice flow across the grounding line is equally controlled by inland basal hydrological
270 conditions and processes that govern basal sliding and sediment deformation. A wide range
271 of observations over the Greenland ice sheet suggests that surface meltwater reaches the
272 bed by fracture and drainage through moulins, which is likely to affect basal lubrication⁴³.
273 Recent work has shown that it is not simply mean surface melt but an increase in water
274 input variability that drives faster ice flow⁴⁴. This has been confirmed by observations⁴⁵.
275 However, more recent work supports the original contention that increased meltwater
276 leads to increases in basal sliding, but that the effect is much smaller than originally

277 thought because of buffering by subglacial drainage system evolution ⁴⁶. Given the available
278 evidence, the representation of basal sliding in large-scale ice sheet models still depends
279 largely on empirical parameterisations based on observations of seasonal variations in ice
280 flow.

281 Recent developments in the understanding of calving follow either fundamental
282 process approaches^{47,48}, leading to global calving laws relating thickness at the grounding
283 line/calving front to calving rate, or are based on stochastic modelling and fracture
284 theory⁴⁹. Two-dimensional generalizations of similar calving laws have been proposed in
285 large-scale models⁵⁰. More specific approaches take into consideration environmental
286 factors, relating surface meltwater runoff and sub-shelf melting to the widening of
287 crevasses and subsequent calving⁵¹. However, model applications based on this approach
288 remain restricted to one-dimensional flowline models⁵², due to the lack of data to resolve
289 the geometry of outlet glacier embayments at sufficiently high spatial resolution.
290 While improvements have been made over recent years, this lack of data hampers a
291 complete process-based evaluation of calving. In the near future, it is likely that models will
292 continue to rely on empirically-based parameterisations of calving.

293

294 **4.0 Future ice sheet changes**

295

296 For significantly warmer climates, both the Greenland and Antarctic ice sheets are
297 projected to lose mass⁵³. General circulation models (GCMs) generally project a small
298 increase of snowfall over both ice sheets (Figure 2(d,e)). However, the mass loss from
299 increasing surface melt will be dominant over the GrIS. For Antarctica, while the SMB is

300 projected to increase, there remain major uncertainties concerning the response of the
301 marine ice sheets and ice shelves to ocean forcing.

302 Surface melt already occurs over a large part of the GrIS during summer and reached a
303 new record in 2012⁵⁴. Therefore, rising temperatures will mainly impact mass loss through
304 increased surface melt in summer, and several positive feedbacks may accelerate this
305 surface mass loss:

- 306 • Polar amplification of global warming resulting from, among other processes, the sea-
307 ice extent decrease over the Arctic Ocean and its associated positive albedo feedback.
308 This process, already observed in recent years⁵⁵ and simulated by the Coupled Model
309 Inter-comparison Project Phase 5 (CMIP5) GCMs (see Figure 2c compared with Figure
310 2a), doubles the estimated uncertainties in projected near-surface temperature
311 anomalies for Greenland compared with those at the global scale⁵⁶.
- 312 • Positive snow albedo feedback over the ice sheet itself associated with the expansion
313 of the bare ice zone. This effect explains why the meltwater runoff increases
314 quadratically with rising summer temperatures: the albedo of bare ice (0.3-0.5) is
315 much less than that of melting snow (~0.7), and surface meltwater becomes more
316 likely to run off rather than percolating into deeper parts of the snowpack⁵⁷.
- 317 • Positive elevation feedbacks associated with the thinning of the ice sheet resulting
318 from the increasing surface melt and ice discharge. Significant thinning (up to 100 m)
319 of the ice sheet is projected along the ice-sheet margin⁵⁸, which should cause an
320 additional melt increase over this area (as ice moves to lower elevations, where it is
321 warmer).

322

323
324 Dynamical changes of the GrIS due to enhanced lubrication, calving and ocean warming still
325 remain difficult to predict. Higher-order ice flow modelling of observed retreat of GrIS
326 glaciers over the last decade and subsequent upscaling leads to a minimum additional SLR
327 of 6 ± 2 mm by 2100, with an upper bound of 45 mm when recurring forcing is applied⁵⁹,
328 while similar upscaling of realistic atmospheric and oceanic forcing of four GrIS glaciers
329 with a calving model leads to a maximum dynamic contribution of 40-85 mm by 2100⁶⁰.
330 This is still lower than previous estimates, but higher than when this retreat chronology is
331 implemented in a 3D higher-order model, leading to a dynamic contribution of 7-15mm⁶¹.
332 The reason for such low numbers is that due to the retreat of the ice sheet margin, calving
333 seems to decrease in relative importance^{53,61}. According to a model inter-comparison⁶²
334 increased ice shelf melt rates of 2 m yr^{-1} lead to 27 mm SLR by 2100 (and 135 mm from a
335 high melt rate of 20 m yr^{-1}). In response to SMB changes, ice-sheet model results are quite
336 consistent and most studies conclude that the largest uncertainty comes from the spread
337 among global climate models, which is amplified by some of the above-mentioned
338 feedbacks over Greenland^{56,58}.

339 For Antarctica, the amplification of the global climate modelling uncertainties is
340 smaller and the contribution of Antarctica to SLR is predicted to increase logarithmically
341 with rising global temperatures (as positive feedbacks become increasingly apparent later)
342 but with little change, and even perhaps a negative contribution, in the next 100-200
343 years⁵³. Firstly, polar amplification resulting from reduced sea-ice coverage seems to be
344 smaller than for the Arctic (see Figure 2b). However, a changing Antarctic Circumpolar
345 Current could potentially allow warmer water to penetrate into the coastal shelf regions of

346 Antarctica – as is observed⁶³. Secondly, little surface melt currently occurs and rising
347 temperatures are not expected to significantly enhance surface melt in the next 100
348 years⁵³. Thirdly, an increase in snowfall is expected to be more significant due to
349 atmospheric temperature rise, hence leading to an increase in SMB⁶⁴. Here, the elevation
350 feedback resulting from SMB changes is negative because the ice sheet is initially projected
351 to thicken⁵³, which is expected to affect its dynamics, especially on longer than centennial
352 time scales.

353 The response of ice-sheet dynamics is twofold, due to increased accumulation and to
354 higher ocean temperatures (in particular below the ice shelves). Two models^{53,65} produce
355 ice-sheet thickening over East Antarctica and increased ice flux at the grounding line due to
356 higher snowfall. However, both studies^{53,65} fail to account for processes at the ice-sheet –
357 ice shelf – ocean interface, such as grounding-line retreat or loss of buttressing³⁹. To date, a
358 continental-scale Antarctic ice-sheet model assessment taking into account those
359 fundamental processes is lacking, although one assessment – based on a wide variety of
360 model complexities – does report large inter-model variability in response to ocean
361 forcing⁶². Process-based modelling of parts of the WAIS, such as Pine Island Glacier, results
362 in a SLR contribution of 27 mm by 2100 for a modest grounding-line retreat of 25 km³⁶,
363 while significant (100 km) grounding line retreat was reported elsewhere²⁸. An alternative
364 method based on probabilistic extrapolation of sustained glacier retreat from such
365 numerical model output³⁶ leads to a SLR contribution of 130 mm by 2100⁶⁶.

366

367

368 **5.0 Other contributions to SLR**

369
370 The global average rate of SLR over the last few decades was about 2-3 mm yr⁻¹ ⁶⁷.
371 Estimates of the global contribution from glaciers and ice caps (GICs) to SLR in the IPCC
372 AR4, 0.50±0.18 mm yr⁻¹ (1961-2003) and 0.77±0.22 mm yr⁻¹ (1993-2003), were based on
373 extrapolation of sparse mass balance measurements made by the glaciological method¹
374 (Box 3). These values were later considered underestimates⁶⁸, due to the poor
375 representation in the glacier inventories of the GICs peripheral to Greenland and Antarctica
376 (PGICs): thus the 1961-2003 value was raised, based on a combined modelling and
377 observations approach⁶⁸, to 0.79±0.34 mm yr⁻¹ (no value provided for 1993-2003). A later
378 extrapolation-based global estimate⁶⁹, with the novelty of allowing explicitly for glacier
379 shrinkage, resulted in a lower estimate of 0.63 mm yr⁻¹ for 1961-2006 (no uncertainty
380 given). The extrapolation-based global estimates have been improved by the addition of
381 geodetic mass balances (Box 3) to the inventories of mass balance by the glaciological
382 method, which has resulted in consistently larger contributions to SLR, especially for the
383 most recent periods (e.g. 0.99±0.04 and 1.46±0.34 mm yr⁻¹ for 1993-2008 and 2000-2005,
384 respectively^{67,70}, compared with 0.97 and 0.95 mm yr⁻¹ for 1993-2006 and 2002-2006
385 respectively⁶⁹).

386 Satellite gravimetry, a method traditionally restricted to the large ice sheets, has
387 recently been used to estimate the global contribution of GICs to SLR⁷¹. GRACE data alone
388 do not have the resolution to separate the Greenland and Antarctic ice sheets from their
389 PGICs, but using an upscaling approach similar to that of ref. 68 has allowed one group to
390 estimate a global contribution from GICs to SLR of 0.63±0.23 mm yr⁻¹ during 2003-2010⁷¹,
391 which is 30% and 47% lower than the two previous estimates that most closely match this

392 period (2002-2006⁶⁹ and 2005-2010⁷², respectively). GRACE results for GICs, however, are
393 sensitive to the models used for calculating GIA, post-Little Ice Age isostatic rebound, and
394 surface- and ground-water mass transfer corrections.

395 The large uncertainties associated with the conventional extrapolation-based
396 methods mostly arise from the uneven representation of the glacierized regions in the
397 mass balance measurements and the incomplete knowledge of the PGICs, both in terms of
398 poorly known mass balances and inaccurate estimates of their area. The latter has greatly
399 improved with the recent release of the Randolph Glacier Inventory⁷³. A consensus
400 estimate combining GRACE, laser altimetry and the extrapolation-based method, using a
401 common inventory of glaciers and a common spatial and temporal reference⁷⁴, has very
402 recently enabled reconciliation of the disparate global estimates of wastage from GICs so
403 far available from the different techniques. The consensus value is $0.71 \pm 0.08 \text{ mm yr}^{-1}$
404 during 2003-2009, which is far lower than the extrapolation-based approach⁷² and
405 somewhat higher than the GRACE-based estimate⁷¹.

406 Ocean thermal expansion (OTE) is a major component of the SLR observed during
407 the late 20th century⁶⁷, and is projected to continue through the 21st century and beyond⁷⁵.
408 The IPCC AR4 found that OTE contributed ~25% of the observed SLR for 1961-2003 and
409 ~50% for 1993-2003¹. Time-varying biases in the ocean temperature data, however, were
410 recently detected⁷⁶ and reduced. It is now understood that the percentages of SLR
411 explained by OTE during the above periods are almost identical⁷⁷, and so are higher for
412 1961-2003 and lower for 1993-2003 than estimated in the IPCC AR4. A recent sea level
413 budget⁶⁷ indicates that OTE contributed ~40% of the observed SLR since 1970 and ~30%
414 since 1993. Warming in the upper 700 m of the ocean explains about 70-80% of the OTE

415 rates. Multi-decadal rates⁷⁷ for OTE in the upper 700 m are 0.71 ± 0.10 mm yr⁻¹ for 1970-
416 2011 and 0.85 ± 0.20 mm yr⁻¹ for 1993-2011, based on linear regression and time-variable
417 uncertainties. Multi-decadal rates for the deep/abyssal ocean are very uncertain⁶⁷, as these
418 are the most poorly sampled regions of the ocean. Since 2005, about 3,000 autonomous
419 Argo profiling floats have been monitoring the upper 2,000 m of the ocean. The Argo-based
420 OTE rate⁷⁸ for 2005-2011 is 0.60 ± 0.20 mm yr⁻¹, in close agreement with the change
421 inferred from satellite altimetry and GRACE⁷⁹. Although consistent with the rates estimated
422 for the multi-decadal periods, the OTE rate for 2005-2011 is unlikely to represent long-
423 term changes. Over such a short period, long-term changes can be easily obscured by more
424 energetic ocean variability, such as fluctuations in the phase of the El Niño Southern
425 Oscillation⁸⁰.

426 Recent estimates for total terrestrial water storage changes during 1993-2008,
427 which include dam retention, groundwater depletion and natural terrestrial storage
428 changes, give values ranging from -0.08 ± 0.19 ⁶⁷ to 0.10 ± 0.20 ⁸¹ mm yr⁻¹. A much larger
429 (positive) contribution dominated by groundwater depletion has recently been
430 suggested⁸², although this result is still controversial⁸³.

431 Table 1 summarizes the recent and current contributions to SLR calculated with the
432 methods discussed in this paper and compares their sum with the observed SLR from tide
433 gauges and satellite altimetry⁶⁷. OTE appears as the main current contributor to SLR,
434 closely followed by the large ice sheets, whose contribution is increasing, and the GICs. The
435 contribution from land-ice masses (ice sheets and GICs) could be slightly overestimated,
436 because only some of the methods in the consensus estimate for ice sheets² explicitly
437 exclude the PGICs (and thus the contribution from PGICs may have been double-counted).

438 Also, the apparent decrease in the contribution from the GICs between the two periods
439 (Table 1) is mostly a result of the different methods used, rather than a result of a lower
440 SMB observed during 2005-2010⁷² (to illustrate this, we note that the GICs SLR
441 contribution given in ref. 72 for 2000-2010 is 1.38 ± 0.21 mm yr⁻¹). Note that, for the most
442 recent period, there is a gap between the sum of contributions and the SLR observed from
443 tide gauges and satellite altimetry.

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445 **6.0 Conclusions and outlook**

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447 During the last 20 years, the AIS as a whole (East, West, and Antarctic Peninsula) has been
448 losing mass, and this is certainly true of the GrIS². There are still disagreements between
449 the numbers that come from the mass-balance retrieval techniques, particularly for East
450 Antarctica, demonstrating a need to better understand the errors of each method. For radar
451 altimetry, further assessment is needed of surface-density corrections and of short-term
452 corrections to ENVISat radar altimetry data⁸⁴, as more moderate estimates of rates of mass
453 change are possible using such corrections. For the mass budget method, NASA's Icebridge
454 project will provide airborne-radar-based improvements to SMB estimates, and radar-
455 sounding measurements of ice thickness at grounding lines will provide improved
456 discharge estimates. Gravimetry and laser altimetry will have, respectively, GRACE and
457 ICESat-2 follow-on missions (scheduled 2017 and 2016 launches) that will ideally provide
458 a decadal record of whole ice-sheet mass balance. However, it is unlikely that these
459 refinements will change the consensus picture emerging: while Antarctica as a whole is
460 losing mass slowly (assessed to be contributing 0.2 mm yr⁻¹ sea-level equivalent by

461 IMBIE²), Greenland, the Antarctic Peninsula, and parts of West Antarctica are together
462 losing mass at a moderate ($\sim 1 \text{ mm yr}^{-1}$ sea-level equivalent) rate today ($\sim 70\%$ of this
463 mass loss is from Greenland) and rates for each are becoming increasingly negative. For
464 the last decade, the collective sea-level contribution from the ice sheets is similar to those
465 from each of GICs and oceanic thermal expansion.

466 While the WAIS is most likely going to continue to contribute to SLR (although the
467 amount is poorly constrained), the sign of the contribution of the EAIS over the next
468 century is uncertain. From the standpoint of projecting global sea level through this
469 century and beyond, it is of fundamental importance to focus on improving ice-sheet
470 models, including representation of key processes and non-linear transitions. The concern
471 of policymakers rightly focuses on the possibility of extreme outcomes with their large
472 impact potential and adaptation need⁸⁵. This is particularly true for the cryosphere, which
473 non-linearly responds to rising temperatures because of several potential positive
474 feedbacks that may accelerate deglaciation. Improved knowledge of key ice-sheet
475 thresholds would support climate policy decisions. Continued observations of ice-sheet
476 processes and their implementation in ice-sheet models are crucial to ensure more
477 accurate sea-level projections.

478 Other key challenges include a need for up-scaling parameterisations to allow low-
479 resolution models, which run fast but with coarse meshes, to better represent crucial
480 processes. To date, parameterisations for grounding line migration have been proposed³⁸,
481 ²² and tested against more complete models³⁹. While advances have been made on the
482 theoretical level, process-based calving implemented in numerical flow models has, to date,
483 relied on parameterisations. Progress has been achieved in the spinup of ice-sheet models

484 so that initial states are closer to observations through the use of inversion techniques.
485 Nevertheless, the non-linearity of basal drag and its dependency on basal hydrology
486 remains a concern. Time-dependent evolution of basal drag is not yet fully implemented in
487 operational models, partly because subglacial hydrology models have not yet been fully
488 implemented and partly because the required data to calibrate spatially-dependent basal friction
489 laws are lacking. The recent release of velocity maps for various time periods⁸⁶ gives hope
490 that this problem will soon be tackled. A further vital step will be to couple improved ice-
491 sheet models with atmosphere/ocean models and GIA models to account for all the
492 feedbacks between the various physical systems at sufficiently high resolution. This will
493 need to be supported by targeted observations at an appropriate spatial and temporal
494 coverage.

495

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497

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506 **Author contributions**

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EH coordinated the study, EH, FN & FP led the writing, and all authors contributed to the writing and discussion of ideas.

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774 Table 1. Estimated recent and current contributions to SLR and observed SLR from tide
775 gauges and satellite altimetry. For Terrestrial Water Storage and Observed only the value
776 for the longer time span is given; the terrestrial water storage number is used for the sum
777 of contributions for both periods. For GICs we have taken, for 1993-2011, an update of that
778 in ref. 72, while for 2003-2009 we have used the value given in ref. 74. The value given for
779 Ocean Thermal Expansion combines a long-term abyssal value⁸⁷ with updates, for the
780 periods shown in the table, from an average of refs. 77 and 88, for the uppermost 700 m,
781 and from ref. 88, for 0-2000 m. The value given for terrestrial water storage is an average
782 of those in the references shown. The uncertainties given are the published errors from the
783 individual studies (usually standard deviations). When data from several sources are
784 combined, and for the sum of contributions, the quoted error is the square root of the sum
785 of the individual variances.

	SLR (mm yr ⁻¹)	
	1992/93-2008/11	2000/03-2009/11
GrIS + AIS ²	0.59 ± 0.20	0.82 ± 0.16
GICs ^{72,74}	1.40 ± 0.16	0.71 ± 0.08
Ocean thermal ^{77,87,88}	1.10 ± 0.43	1.11 ± 0.80
Terrestrial water (1993-2008) ^{67,81}	0.02 ± 0.26	
Sum of contributions	3.11 ± 0.56	2.66 ± 0.86
Observed (1993-2008) ⁶⁷	3.22 ± 0.41	

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807 **Figure captions**

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809 Figure 1. Summary of Antarctic and Greenland mass-rate estimates. In the studies
810 published before 2012² (left) and in 2012 (right) each estimate of a temporally-averaged
811 rate of mass change is represented by a box whose width indicates the time period studied,

812 and whose height indicates the error estimate. Single-epoch (snapshot) estimates of mass
813 balance are represented by vertical error bars when error estimates are available, and are
814 otherwise represented by asterisks. 2012 studies comprise IMBIE combined estimates²
815 (solid lines), and estimates by Sasgen and others^{16,20} and King and others¹¹ (dashed lines),
816 Zwally and others¹⁹ (dot-dashed lines), Harig and Simons⁸⁹ and Ewert and others⁹⁰ (dotted
817 lines).

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819 Figure 2. Comparison of projected global, Antarctic and Greenland surface air temperature
820 and snowfall anomalies to 2100. a) Anomaly of global mean 2m air temperature (T2m)
821 simulated by 30 GCMs from the CMIP5 data base. Values are with respect to 1970-1999 for
822 the RCP 4.5 (blue) and RCP 8.5 (red) scenarios. We refer to ref. 91 for more details about
823 the Representative Concentration Pathways (RCP) scenarios. The evolving ensemble means
824 are plotted as thick lines, with vertical bars representing ± 1 standard deviation for each decade. A
825 10-year running mean was used to smooth the curves. b) Same as a) but for Antarctica. The
826 land sea mask from each GCM is used to delimit Antarctica. c) Same as a) but for T2m over
827 GrIS. The T2m anomaly is taken over the area covering Greenland (60-85°N and 20-70°W)
828 and where surface elevation is higher than 1000 m a.s.l. d) Same as b) but for precipitation.
829 Anomalies are given in % with respect to the mean precipitation for 1970-1999. e) Same as
830 c) but for precipitation.

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832 Figure 3. Illustration of a marine ice sheet and its interaction with the ocean. Warm
833 modified Circumpolar Deep Water (mCDW) leads to melting at the grounding line, leading
834 to ice shelf thinning and grounding line retreat. Marine ice-sheet instability occurs when, in

835 the lack of buttressing, the grounding line retreats on an upward-sloping (in the direction
836 of the flow) bedrock: ice flux increases with thickness at the grounding line, leading to an
837 increased outflux to the ocean that may be compensated by further grounding-line retreat,
838 until a new downward-sloping bed (pinning point) is reached. Thinning of ice sheet and
839 shelf can also be caused by surface melt and increased calving.

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BOX 1: GIA models - recent developments

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Glacial isostatic adjustment (GIA) is the response of the solid Earth, including associated changes in planetary gravity and rotation, to past ice and ocean mass redistribution^{92,93}. The clearest observable effect of GIA is regional vertical rebound of the Earth's surface. Models of GIA are necessary for correcting measurements of present-day ice-mass change⁹⁴

869 and for long-term modelling⁴². The assimilation of glacial geological constraints on former
870 ice extent and geodetic constraints on rebound into GIA models is helping to reduce the
871 uncertainty associated with GIA, and hence estimates of ice-mass change^{11,12,95}. However,
872 several key challenges remain. First, ice extent and thickness changes during the last
873 millennium are poorly known, and typically not included in GIA models, despite the fact
874 that they can dominate the present-day rebound signal, especially in regions of low mantle
875 viscosity^{96,97}. Second, lateral variations in Earth structure, as detected beneath Antarctica⁹⁸,
876 also influence the GIA signal, but are not included in most models to date. Finally, the
877 limitations of the data used to tune GIA models mean that probabilistic approaches are now
878 being adopted to seek the most likely range of solutions⁹⁹.

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886 ***BOX 2: Marine ice sheets, grounding lines and buttressing***

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888 Marine ice sheets, such as the West Antarctic ice sheet, rest on bedrock that lies below sea
889 level. These grounded ice sheets are fringed by floating ice shelves. The grounding line is
890 the contact of the ice sheet with the ocean where the ice mass starts to float by buoyancy.

891 Ice from the grounded ice sheet is discharged across the grounding line into ice shelves,
892 from where icebergs break off, through a process called calving (Figure 3).

893

894 The migration of the grounding line is a result of the local balance between the masses of
895 ice and displaced ocean water. The grounding line advances if previously floating ice
896 becomes thick enough to ground, or retreats if previously grounded ice becomes thin
897 enough to float. Theory has demonstrated that in order to simulate grounding line
898 migration, it is necessary to include (horizontal) stress gradients across the grounding
899 zone²² and in order to resolve this numerically, a high spatial resolution is needed, either
900 by using a moving grid (following the grounding line directly) or by sub-sampling the grid
901 around the grounding line to hundreds of metres³⁹. This high resolution is necessary to
902 resolve horizontal stress gradients across a narrow boundary layer.

903

904 Ice discharge generally increases with increasing ice thickness at the grounding line. For a
905 bed sloping down toward the interior this may lead to unstable grounding line retreat, as
906 increased flux (e.g., due to reduced buttressing) leads to thinning and eventually floatation,
907 which moves the grounding line into deeper water where the ice is thicker. Thicker ice
908 results in increased ice flux, which further thins (and eventually floats) the ice, which
909 results in further retreat into deeper water (and thicker ice), etc. (Figure 3). This unstable
910 retreat is referred to as the marine ice-sheet instability²². However, the grounding line is
911 partially stabilized by the presence of ice shelves, which are either confined laterally
912 through embayments or otherwise stabilised by locally grounded features which they
913 enclose (e.g. pinning points). Both geometries transmit a back-force, or “buttressing”,

914 toward the grounded ice sheet, which may help to stabilise the grounding line against
915 unstable retreat down inland-sloping bedrock¹⁰⁰.

916

917 Thinning of ice shelves reduces drag at the margins and over pinning points, leading to
918 increased ice flow across the grounding line, causing grounding-line retreat until a new
919 stable point (e.g. upward sloping bedrock) is reached. The mechanisms described above
920 rely heavily on a precise knowledge of the geometry of the ice/ocean contact, which
921 explains why neighbouring outlet glaciers, in contact with the ocean, and subject to the
922 same atmospheric and oceanic forcing, may exhibit contrasting behaviours³⁰.

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931 ***BOX 3: Glaciological versus geodetic method***

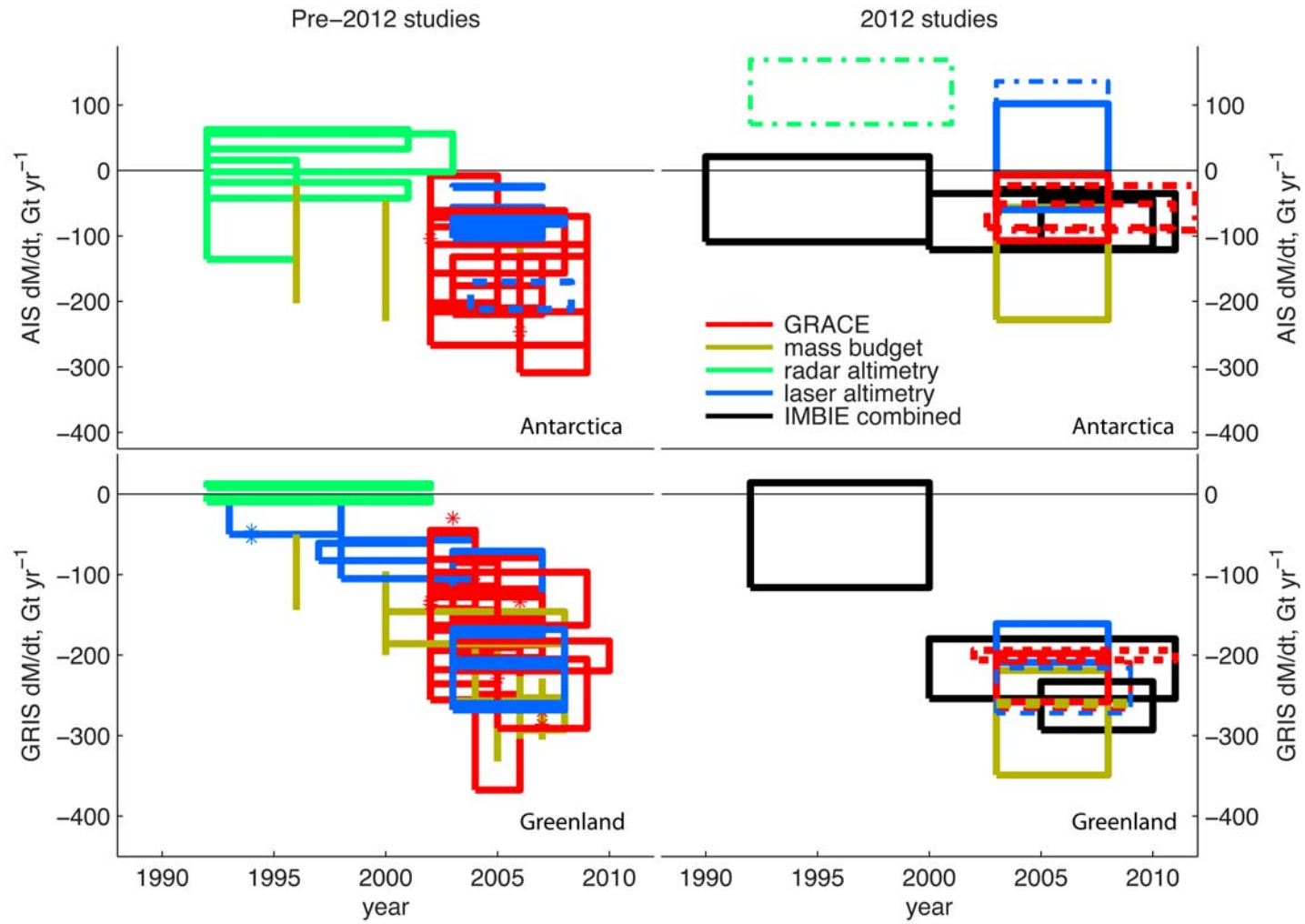
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933 GIC mass balance estimates by the *glaciological method* are based on extrapolation over the
934 whole glacier surface of measurements of accumulation and ablation made in-situ at single
935 points. These measurements include readings of surface elevation changes at stakes, sampling

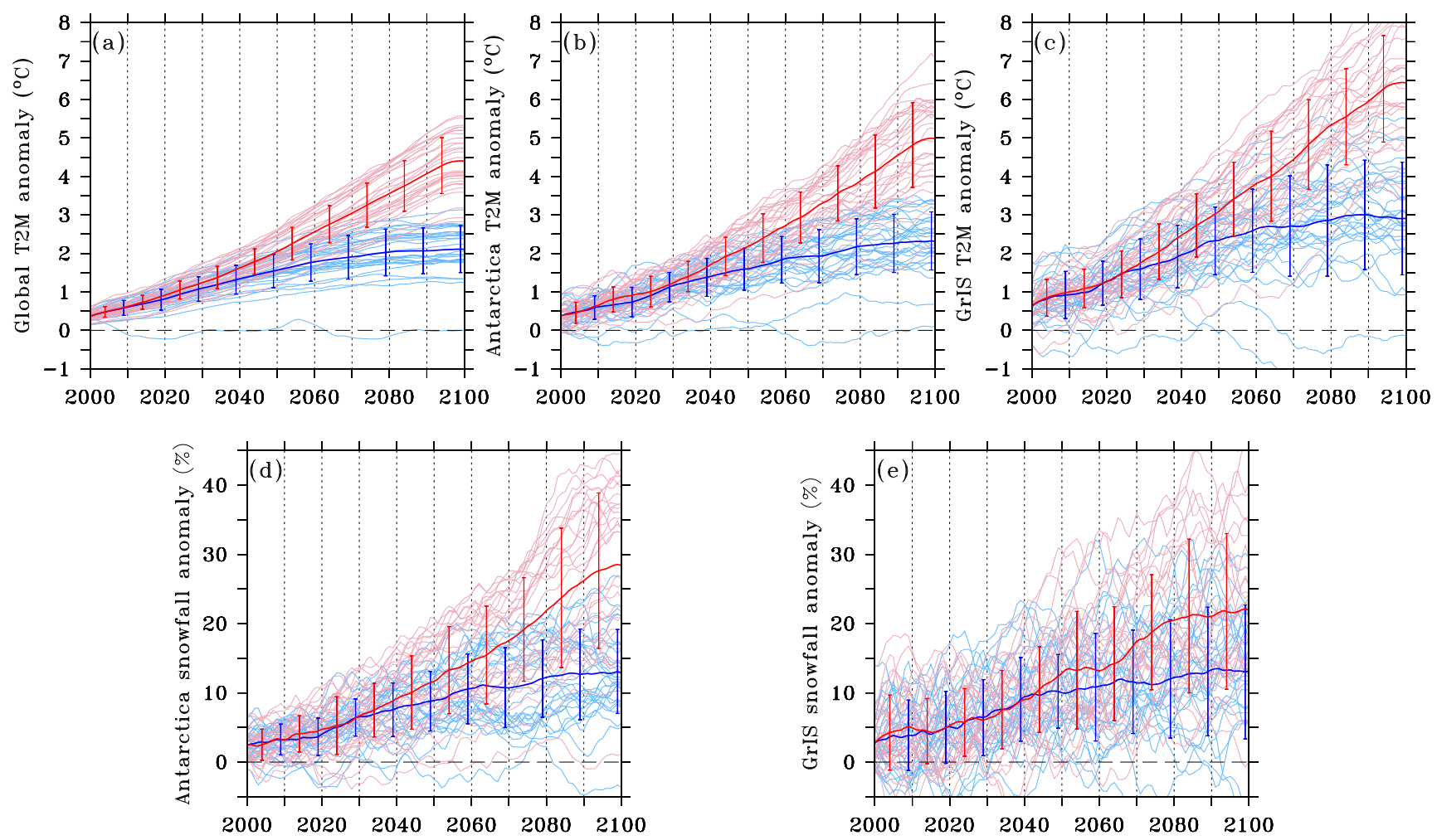
936 of density and accumulation in pits and shallow cores, and depth probing of the snow and firn,
937 and shallow coring.

938 Estimates by the *geodetic method* are based on repeated mapping of glacier surface
939 elevations to estimate the volume changes, from which the mass changes are calculated
940 using information about the density of the material and its time variations. The elevation
941 changes can be measured using different techniques, either from the glacier surface or,
942 more commonly, from airborne or satellite-borne sensors.

Figure 1



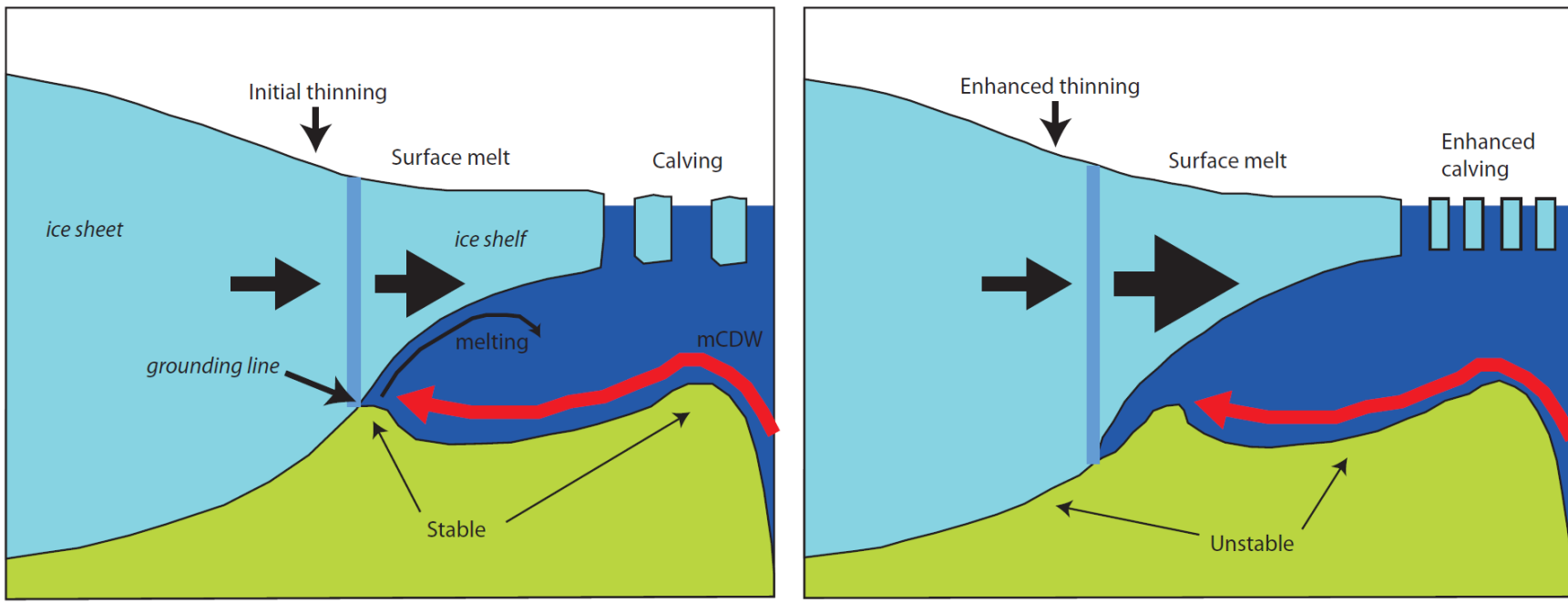
1 Figure 2
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