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

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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 AR4<sup>1</sup>. 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 comparison<sup>2</sup>. 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 loss<sup>2</sup> (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<sup>3,4</sup>  
107 or laser altimetry<sup>5</sup>.

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 system<sup>6</sup>.

111        -The mass budget technique compares estimates of the net ice accumulation on the  
112 ice sheets with estimates of discharge across the grounding line<sup>7</sup> (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<sup>7,8</sup> studies use modelled snowfall fields from atmospheric

116 reanalysis data<sup>9,10</sup> 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)<sup>6</sup>.  
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 basins<sup>11</sup>.

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<sup>12,13</sup>, coupled with more robust GPS  
146 observations of the motion of exposed bedrock<sup>14</sup>, strongly suggest that Antarctic GIA-  
147 related bedrock motion peaks at about 5-6 mm yr<sup>-1</sup>. The resulting GIA models for  
148 Antarctica<sup>13,15</sup> 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<sup>13,16,17</sup>.

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 technique<sup>2</sup>. 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 data<sup>18</sup>. 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 balance<sup>19</sup>. 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  $-120 \text{ Gt}$   
178  $\text{yr}^{-1}$ , 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 <sup>120</sup>. 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<sup>-1</sup> 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 Antarctica<sup>2</sup>. For East Antarctica, a mass gain of +101 Gt yr<sup>-1</sup> for 2003-2008 has been  
194 proposed recently based on laser altimetry<sup>19</sup>, which is larger than the IMBIE GRACE  
195 estimate of +35 Gt yr<sup>-1</sup> and near the upper end of the laser altimetry estimates<sup>2</sup>.

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 approximation<sup>21</sup>. 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 approximation<sup>22</sup>. 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)<sup>21,23,24</sup>. 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)<sup>25,26</sup>. 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<sup>27,28</sup>.

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<sup>27,28</sup>) or adaptive mesh refinement<sup>29</sup> 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 interferometry<sup>30</sup>, surface elevation change through satellite and airborne

231 campaigns (IceBridge), and high-resolution bedrock and ice thickness measurements<sup>31</sup>.  
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<sup>27-29,32</sup>. 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<sup>33,34</sup>. Mechanisms behind this speedup are complex. Oceanic and/or atmospheric  
244 warming leads to ice-shelf thinning or disintegration<sup>35,36</sup>, which in turn may lead to loss of  
245 buttressing<sup>37</sup>, grounding line retreat and hence glacier speedup<sup>33</sup> (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 SLR<sup>38</sup>, support these mechanisms.

249 Major theoretical advances<sup>22</sup> 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 uncertainties<sup>39</sup>: 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 resolution<sup>39</sup>. 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 retreat<sup>29</sup>, or  
260 grounding-line stabilization approximately 25 km inland within 100 years<sup>37</sup>.

261 GIA also influences ice-sheet behaviour<sup>40,41</sup>. 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 Antarctica<sup>42</sup>. 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 positions<sup>22</sup>. Ignoring such processes can fundamentally alter model predictions relating to  
268 the stability of a marine-grounded ice sheet<sup>41</sup>.

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 lubrication<sup>43</sup>.  
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 flow<sup>44</sup>. This has been confirmed by observations<sup>45</sup>.  
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 <sup>46</sup>. 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<sup>47,48</sup>, 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 theory<sup>49</sup>. Two-dimensional generalizations of similar calving laws have been proposed in  
285 large-scale models<sup>50</sup>. 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 calving<sup>51</sup>. However, model applications based on this approach  
288 remain restricted to one-dimensional flowline models<sup>52</sup>, 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 mass<sup>53</sup>. 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 2012<sup>54</sup>. 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 years<sup>55</sup> 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 scale<sup>56</sup>.
- 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 snowpack<sup>57</sup>.
- 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 margin<sup>58</sup>, 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  $6\pm 2$  mm by 2100, with an upper bound of 45 mm when recurring forcing is applied<sup>59</sup>,  
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 2100<sup>60</sup>.  
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 7-15mm<sup>61</sup>.  
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<sup>53,61</sup>. According to a model inter-comparison<sup>62</sup>  
334 increased ice shelf melt rates of  $2 \text{ m yr}^{-1}$  lead to 27 mm SLR by 2100 (and 135 mm from a  
335 high melt rate of  $20 \text{ m yr}^{-1}$ ). 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<sup>56,58</sup>.

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 years<sup>53</sup>. 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 observed<sup>63</sup>. Secondly, little surface melt currently occurs and rising  
347 temperatures are not expected to significantly enhance surface melt in the next 100  
348 years<sup>53</sup>. Thirdly, an increase in snowfall is expected to be more significant due to  
349 atmospheric temperature rise, hence leading to an increase in SMB<sup>64</sup>. Here, the elevation  
350 feedback resulting from SMB changes is negative because the ice sheet is initially projected  
351 to thicken<sup>53</sup>, 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<sup>53,65</sup> produce  
355 ice-sheet thickening over East Antarctica and increased ice flux at the grounding line due to  
356 higher snowfall. However, both studies<sup>53,65</sup> fail to account for processes at the ice-sheet –  
357 ice shelf – ocean interface, such as grounding-line retreat or loss of buttressing<sup>39</sup>. 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 forcing<sup>62</sup>. 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 km<sup>36</sup>,  
363 while significant (100 km) grounding line retreat was reported elsewhere<sup>28</sup>. An alternative  
364 method based on probabilistic extrapolation of sustained glacier retreat from such  
365 numerical model output<sup>36</sup> leads to a SLR contribution of 130 mm by 2100<sup>66</sup>.

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<sup>-1</sup> <sup>67</sup>.  
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<sup>-1</sup> (1961-2003) and 0.77±0.22 mm yr<sup>-1</sup> (1993-2003), were based on  
373 extrapolation of sparse mass balance measurements made by the glaciological method<sup>1</sup>  
374 (Box 3). These values were later considered underestimates<sup>68</sup>, 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 approach<sup>68</sup>, to 0.79±0.34 mm yr<sup>-1</sup> (no value provided for 1993-2003). A later  
378 extrapolation-based global estimate<sup>69</sup>, with the novelty of allowing explicitly for glacier  
379 shrinkage, resulted in a lower estimate of 0.63 mm yr<sup>-1</sup> 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<sup>-1</sup> for 1993-2008 and 2000-2005,  
384 respectively<sup>67,70</sup>, compared with 0.97 and 0.95 mm yr<sup>-1</sup> for 1993-2006 and 2002-2006  
385 respectively<sup>69</sup>).

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 SLR<sup>71</sup>. 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<sup>-1</sup> during 2003-2010<sup>71</sup>,  
391 which is 30% and 47% lower than the two previous estimates that most closely match this

392 period (2002-2006<sup>69</sup> and 2005-2010<sup>72</sup>, 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 Inventory<sup>73</sup>. 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 reference<sup>74</sup>, 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 approach<sup>72</sup> and  
405 somewhat higher than the GRACE-based estimate<sup>71</sup>.

406         Ocean thermal expansion (OTE) is a major component of the SLR observed during  
407 the late 20<sup>th</sup> century<sup>67</sup>, and is projected to continue through the 21<sup>st</sup> century and beyond<sup>75</sup>.  
408 The IPCC AR4 found that OTE contributed ~25% of the observed SLR for 1961-2003 and  
409 ~50% for 1993-2003<sup>1</sup>. Time-varying biases in the ocean temperature data, however, were  
410 recently detected<sup>76</sup> and reduced. It is now understood that the percentages of SLR  
411 explained by OTE during the above periods are almost identical<sup>77</sup>, and so are higher for  
412 1961-2003 and lower for 1993-2003 than estimated in the IPCC AR4. A recent sea level  
413 budget<sup>67</sup> 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 rates<sup>77</sup> for OTE in the upper 700 m are  $0.71\pm 0.10$  mm yr<sup>-1</sup> for 1970-  
416 2011 and  $0.85\pm 0.20$  mm yr<sup>-1</sup> for 1993-2011, based on linear regression and time-variable  
417 uncertainties. Multi-decadal rates for the deep/abyssal ocean are very uncertain<sup>67</sup>, 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 rate<sup>78</sup> for 2005-2011 is  $0.60\pm 0.20$  mm yr<sup>-1</sup>, in close agreement with the change  
421 inferred from satellite altimetry and GRACE<sup>79</sup>. 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 Oscillation<sup>80</sup>.

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  $-0.08\pm 0.19$ <sup>67</sup> to  $0.10\pm 0.20$ <sup>81</sup> mm yr<sup>-1</sup>. A much larger  
429 (positive) contribution dominated by groundwater depletion has recently been  
430 suggested<sup>82</sup>, although this result is still controversial<sup>83</sup>.

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 altimetry<sup>67</sup>. 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 sheets<sup>2</sup> 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 2005-2010<sup>72</sup> (to illustrate this, we note that the GICs SLR  
441 contribution given in ref. 72 for 2000-2010 is  $1.38 \pm 0.21$  mm yr<sup>-1</sup>). 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.

444

## 445 **6.0 Conclusions and outlook**

446

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 GrIS<sup>2</sup>. 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 data<sup>84</sup>, 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<sup>-1</sup> sea-level equivalent by

461 IMBIE<sup>2</sup>), 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 need<sup>85</sup>. 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 proposed<sup>38</sup>,  
481 <sup>22</sup> and tested against more complete models<sup>39</sup>. 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 periods<sup>86</sup> 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**

507

508 EH coordinated the study, EH, FN & FP led the writing, and all authors contributed to the  
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510

511

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769 *The Cryosphere* **6**, 1497-1505 (2012).

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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 value<sup>87</sup> 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 <sup>-1</sup> )	
	1992/93-2008/11	2000/03-2009/11
GrIS + AIS <sup>2</sup>	0.59 ± 0.20	0.82 ± 0.16
GICs <sup>72,74</sup>	1.40 ± 0.16	0.71 ± 0.08
Ocean thermal <sup>77,87,88</sup>	1.10 ± 0.43	1.11 ± 0.80
Terrestrial water (1993-2008) <sup>67,81</sup>	0.02 ± 0.26	
<b>Sum of contributions</b>	<b>3.11 ± 0.56</b>	<b>2.66 ± 0.86</b>
Observed (1993-2008) <sup>67</sup>	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 2012<sup>2</sup> (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 estimates<sup>2</sup>  
815 (solid lines), and estimates by Sasgen and others<sup>16,20</sup> and King and others<sup>11</sup> (dashed lines),  
816 Zwally and others<sup>19</sup> (dot-dashed lines), Harig and Simons<sup>89</sup> and Ewert and others<sup>90</sup> (dotted  
817 lines).

818

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  $\pm 1$  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<sup>92,93</sup>. 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 change<sup>94</sup>

869 and for long-term modelling<sup>42</sup>. 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<sup>11,12,95</sup>. 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<sup>96,97</sup>. Second, lateral variations in Earth structure, as detected beneath Antarctica<sup>98</sup>,  
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 solutions<sup>99</sup>.

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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 zone<sup>22</sup> 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 metres<sup>39</sup>. 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 instability<sup>22</sup>. 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 bedrock<sup>100</sup>.

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 behaviours<sup>30</sup>.

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

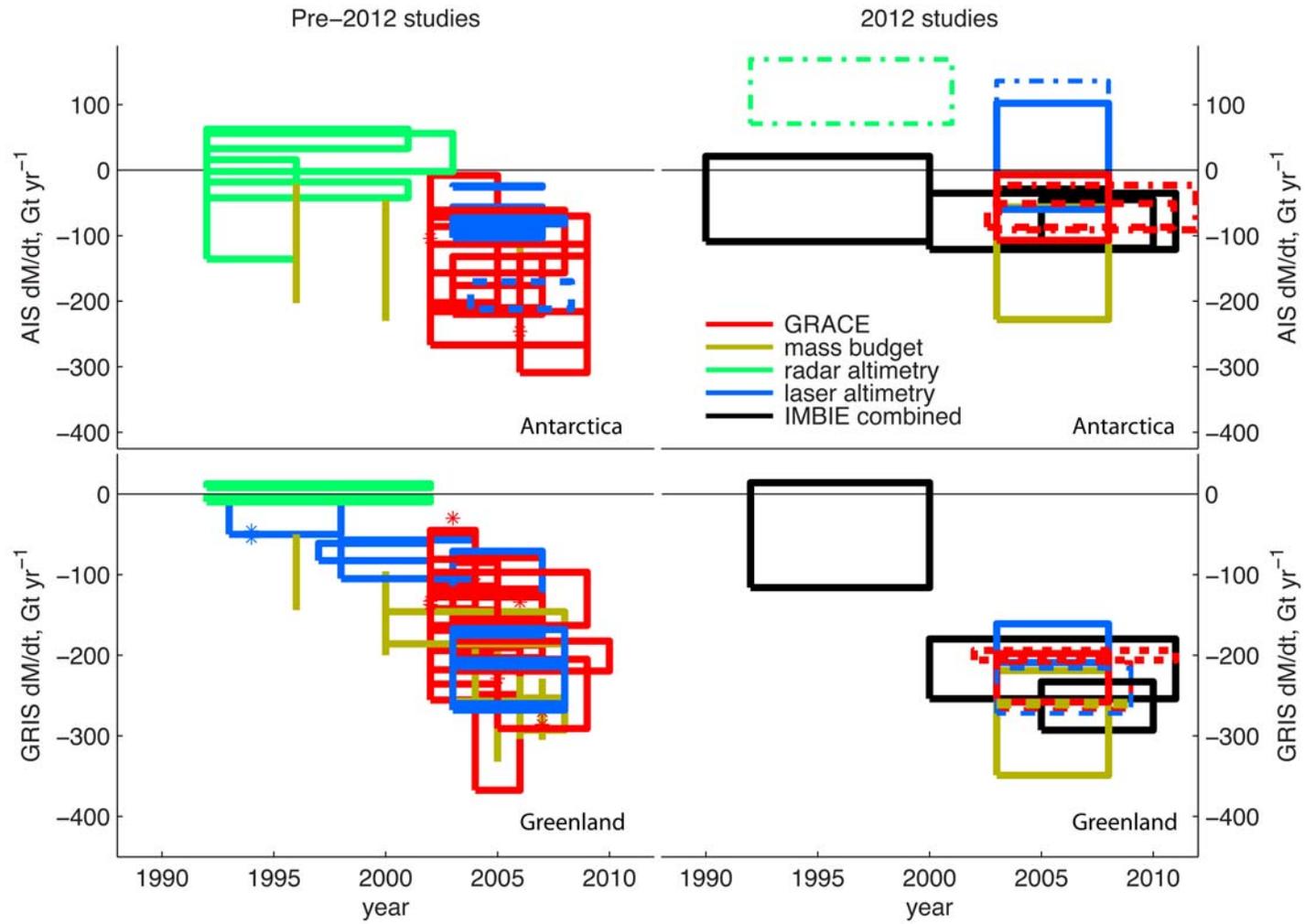
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933 GIC mass balance estimates by the *glaciological method* are based on extrapolation over the  
934 whole glacier surface of measurements of accumulation and ablation made in-situ at single  
935 points. These measurements include readings of surface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

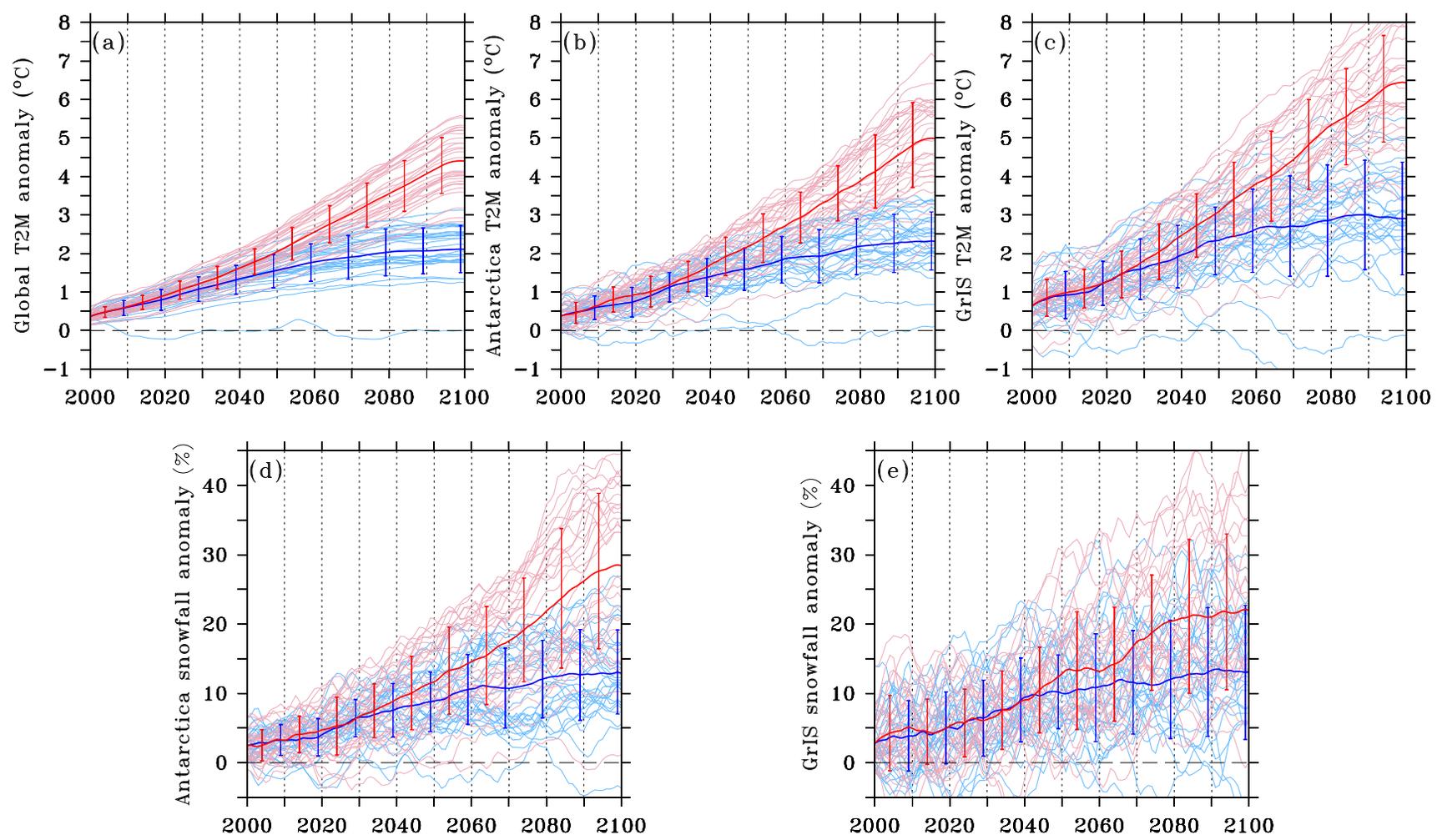
936 of density and accumulation in pits and shallow cores, and depth probing of the snow and firn,  
937 and shallow coring.

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



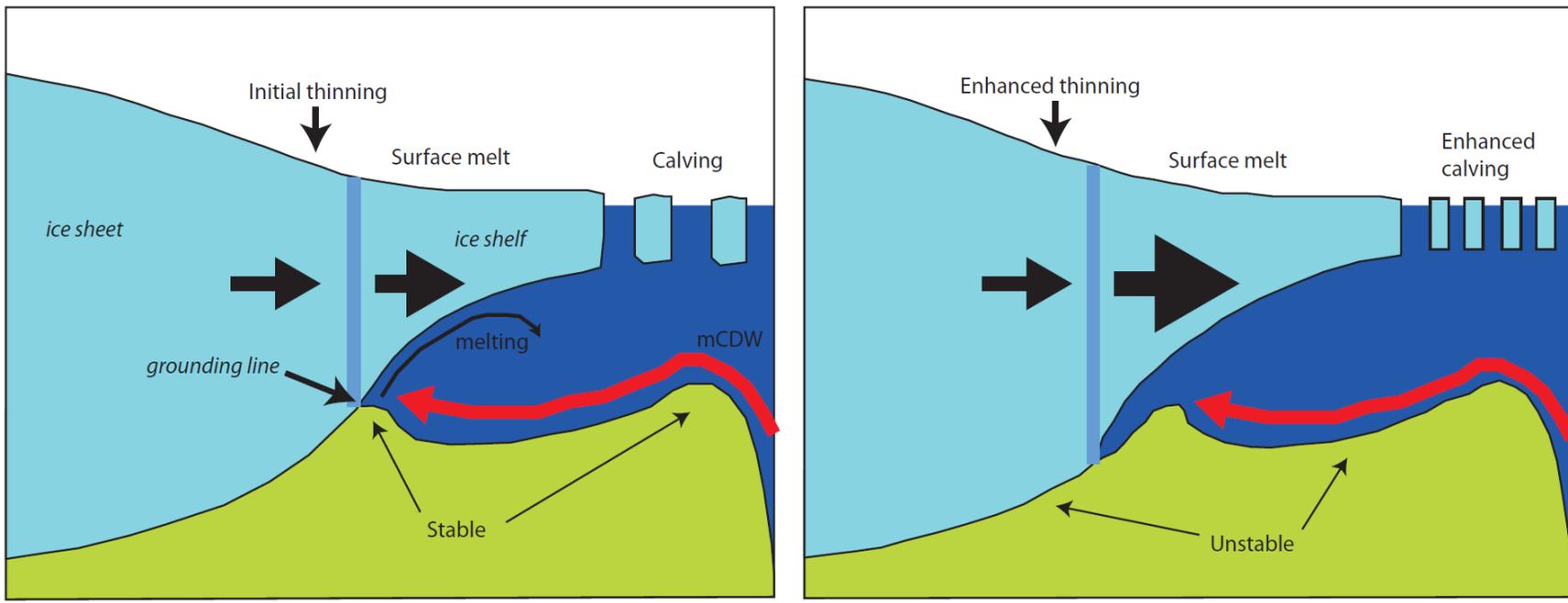
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