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## Mass balance of the ice sheets and glaciers – Progress since AR5 and challenges

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### ► To cite this version:

Edward Hanna, Frank Pattyn, Francisco Navarro, Vincent Favier, Heiko Goelzer, et al.. Mass balance of the ice sheets and glaciers – Progress since AR5 and challenges. *Earth-Science Reviews*, 2020, 201, pp.102976. <10.1016/j.earscirev.2019.102976>. <hal-03036869>

**HAL Id: hal-03036869**

**<https://hal.science/hal-03036869v1>**

Submitted on 16 May 2025

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1 **Mass balance of the ice sheets and glaciers – progress since AR5 and challenges**

2 **EARTH SCIENCE REVIEWS invited review/synthesis paper**

3 **30 September 2019 revised version**

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27 **Abstract.** Recent research shows increasing decadal ice mass losses from the Greenland and  
28 Antarctic Ice Sheets and more generally from glaciers worldwide in the light of continued  
29 global warming. Here, in an update of our previous ISMASS paper (Hanna et al., 2013), we  
30 review recent observational estimates of ice sheet and glacier mass balance, and their related  
31 uncertainties, first briefly considering relevant monitoring methods. Focusing on the response  
32 to climate change during 1992-2018, and especially the post-IPCC AR5 period, we discuss  
33 recent changes in the relative contributions of ice sheets and glaciers to sea-level change. We  
34 assess recent advances in understanding of the relative importance of surface mass balance  
35 and ice dynamics in overall ice-sheet mass change. We also consider recent improvements in  
36 ice-sheet modelling, highlighting data-model linkages and the use of updated observational  
37 datasets in ice-sheet models. Finally, by identifying key deficiencies in the observations and  
38 models that hamper current understanding and limit reliability of future ice-sheet projections,  
39 we make recommendations to the research community for reducing these knowledge gaps.  
40 Our synthesis aims to provide a critical and timely review of the current state of the science  
41 in advance of the next Intergovernmental Panel on Climate Change Assessment Report that is  
42 due in 2021.

## 1.0 Introduction

Major uncertainties in predicting and projecting future sea-level rise are due to the contribution of the two major ice sheets on Earth, Greenland and Antarctica (Pattyn et al., 2018). These uncertainties essentially stem from the fact that both ice sheets may reach a tipping point, in this context defined as (regionally) irreversible mass loss, with a warming climate and that the timing of the onset of such a tipping point is difficult to assess. This is particularly true for the Antarctic Ice Sheets (AIS), where two instability mechanisms potentially operate, allowing a large divergence in timing of onset and mass loss in model projections, while the Greenland Ice Sheet (GrIS) is also particularly susceptible to increased mass loss from surface melting and associated feedbacks under anthropogenic warming.

The Expert Group on Ice Sheet Mass Balance and Sea Level (ISMAS; <http://www.climate-cryosphere.org/activities/groups/ismass>) convened a one-day workshop as part of POLAR2018 in Davos, Switzerland, on 15 June 2018, to discuss advances in ice-sheet observations and modelling since the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC AR5). The talks and discussions are summarised here in an update of our previous review (Hanna et al., 2013) where we synthesised material from a similar workshop held in Portland, Oregon, USA, in July 2012. Here we focus, in the light of advances in the last six years, on what we need to know in order to make improved model projections of ice-sheet change. Apart from providing an update of recent observational estimates of ice-sheet mass changes, we also set this in a wider context of global glacier change. The paper is arranged as follows. In section (2) we discuss recent advances in ice-sheet observations, while section (3) focuses on advances in modelling and identifies remaining challenges – including links with observational needs - that need to be overcome in order to make better projections. Section (4) discusses recent and projected mass-balance rates for glaciers and ice caps, comparing these with recent ice-sheet changes, setting the latter in a broader context of global glacier change. Finally, in section (5) we summarise our findings and make key recommendations for stimulating further research.

## 2.0 Observational estimates of ice-sheet total and surface mass balance

In this section we summarise recent observation-based estimates of the total mass balance of the Antarctic and Greenland ice sheets, also considering changes in surface mass balance (SMB; net snow accumulation minus surface meltwater runoff) and – for marine-terminating glaciers – ice dynamics (solid ice dynamical discharge across the grounding line – the contact of an ice sheet with the ocean where the ice mass becomes buoyant and floats – and subsequent calving of icebergs) where appropriate (**Figure 1**). **Figure 2** shows mean SMB for the ice sheets for recent periods, while mean surface ice flow velocity maps can be found in Rignot et al. (2019) and Mouginot et al. (2019) (Fig. 1A in both papers). Satellite, airborne and in situ observational techniques and modelling studies have provided a detailed representation of recent ice-sheet mass loss and increases in ice melt and discharge (Moon et al., 2012; Enderlin et al., 2014; Bigg et al., 2014; Shepherd et al., 2012, 2018; Trusel et al. 2018; Rignot et al., 2019; Mouginot et al., 2019).

There are three main methods of estimating ice-sheet mass changes. Firstly, radar and laser altimetry (mainly using CryoSat, Envisat, ERA and ICESat satellites), which measure changes in height of the surface over repeat surveys that are interpolated over the surface area of interest to estimate a volume change which is converted into a mass change. This latter is typically done using knowledge or assumptions of the radar return depth and/or near-surface density. Alternatively Zwally et al. (2015) use knowledge of the accumulation-driven mass anomaly during the period of observation, together with the associated accumulation-driven

101 elevation anomaly corrected for the accumulation-driven firn compaction, to derive the total  
102 mass change and its accumulation- and dynamic-driven components Secondly, satellite  
103 gravimetry effectively weighs the ice sheets through their gravitational pull on a pair of  
104 orbiting satellites called GRACE (or, since May 2018, the subsequent GRACE Follow On  
105 mission). Thirdly, the mass budget or component method compares SMB model output with  
106 multi-sensor satellite radar observations of ice velocity across a position on or close to the  
107 grounding line, from which ice discharge can be inferred if the thickness and vertical velocity  
108 profile of ice at that point are also assumed/known. All three methods have their strengths and  
109 weaknesses (e.g. Hanna et al., 2013; Bamber et al., 2018). Altimetry and, especially,  
110 gravimetry, require accurate quantification of Glacial Isostatic Adjustment (GIA; Section 2.3)  
111 which contaminates the ice-sheet mass loss signals. Gravimetry is limited by a relatively  
112 short time series (since 2002) and low spatial resolution (~300 km) compared with the other  
113 methods but is the method that most directly measures mass change.

114 Altimetry surveys, which date relatively far back to the early 1990s, provide elevation  
115 changes that need to be converted into volume and then mass changes, requiring knowledge  
116 of near-surface density which is often highly variable and uncertain for ice sheets. In  
117 addition, radar altimeter surveys do not adequately sample relatively steeper-sloping ice-sheet  
118 margins and require correction for the highly-variable radar-reflection depth that has strong  
119 seasonal variations and interannual trends and complex interactions between linearly-  
120 polarized radar signals and the direction of the surface slope. Successful corrections have  
121 been developed and applied to radar altimeter data from ERS1 and ERS2 using crossover  
122 analysis data (Wingham et al., 1998; Davis and Ferguson, 2004; Zwally et al., 2005; Yi et al.,  
123 2011; Khvorostovsky, 2012) and to Envisat data using repeat track analysis and an advanced  
124 correction algorithm (Filament and Remy, 2012). However, the corrections applied by others  
125 to Envisat and CryoSat data have been questioned due to complex interaction of the cross-  
126 track linearly-polarized radar signal of Envisat and CryoSat with the surface slope that affects  
127 the highly-variable penetration/reflection depth (Zwally et al., 2016; Nilsson et al., 2016).  
128 Also, allowance must be made for firn-compaction changes arising from temperature and/or  
129 accumulation variations, especially in the context of a warming ice-sheet, which significantly  
130 affect surface elevation without mass change (e.g. Li and Zwally, 2015; Zwally et al., 2015).  
131 A number of the altimetry studies included here have used a regionally-varying, temporally  
132 constant effective density value to convert observed volume changes to mass change  
133 estimates. In many cases, a low effective density is assigned for inland areas, and a high  
134 effective density in coastal errors. Because in Greenland and much of Antarctica, coastal  
135 areas are thinning while inland areas are in neutral balance or thickening, this can produce  
136 negative biases in estimated ice-sheet mass-change rates if the changes in the interior are  
137 associated with long-term imbalance between ice flow and snow accumulation.

138 The mass-budget method involves subtracting two large quantities (SMB and  
139 discharge) and needs detailed and complete regional information on these components, which  
140 is recently available from satellite radar data for discharge. SMB cannot be directly measured  
141 at the ice-sheet scale but is instead estimated using regional climate models that are evaluated  
142 and calibrated using in-situ climate and SMB observations. These RCM/SMB models can  
143 have significant uncertainties in derived accumulation and runoff (of the order of 15%, e.g.  
144 Fettweis, 2018). Deriving discharge requires knowledge of bathymetry and the assumption of  
145 an internal velocity profile in order to determine ice flux across the grounding line, and there  
146 are also errors in determining the position of the grounding line. Further uncertainty arises in  
147 estimating the discharge from the areas where the ice velocity is not measured. Despite these  
148 significant uncertainties, an advantage of this method is that the mass change can be  
149 partitioned into its (sub-)components.

150

151 A more recent group use combinations of measurement strategies to minimize the  
152 disadvantages of each, such as by combining altimetric with gravimetric data (Sasgen et al,  
153 2019) or mass-budget data with gravimetric data (e.g. Talpe et al, 2017) to simultaneously  
154 estimate GIA rates and ice-sheet mass-balance rates. These studies typically report errors  
155 comparable to those reported by single-technique studies, but their results may be seen as  
156 more credible because they provide self-consistent solutions for the most important error  
157 sources affecting other studies.

158 A major international research programme called the Ice-sheet Mass Balance Inter-  
159 comparison Exercise (IMBIE; <http://imbie.org/>) has attempted to reconcile differences  
160 between these various methods, and its second phase IMBIE2 has recently reported an  
161 updated set of reconciled total mass balance estimates for Antarctica (Shepherd et al., 2018)  
162 and is shortly expected to update previous results for Greenland. However, despite recent  
163 improvements in coverage and accuracy, modern satellite-based records are too short for  
164 attribution studies aiming to separate the contributions from anthropogenic greenhouse gas  
165 warming signal and background climate variability to the contemporary mass loss (Wouters  
166 et al., 2013), and proxy data such as ice cores are therefore used to overcome this limitation.

167 We have compiled recent estimates of mass balance using available (at the time of  
168 writing) published references from 2014 to 2019 (**Figure 3**), in an update of Figure 1 in  
169 Hanna et al. (2013). Our new box plots clearly show continuing significant mass losses from  
170 both ice sheets, with approximately double the recent rate of mass loss for Greenland  
171 compared with Antarctica. However, the boxes tend to suppress the considerable interannual  
172 variability of mass fluctuations, e.g. the record loss of mass from the GrIS in 2012, and this  
173 shorter-term variability is strikingly shown by annually-resolved time series based on the  
174 mass-budget method [Figure 3 of Rignot et al. (2019) for Antarctica and Figure 3 of  
175 Mougnot et al. (2019) for GrIS].

176

## 177 *2.1 Antarctic ice sheets*

178

179 Recent work agrees on significant and steadily growing mass losses from the West Antarctic  
180 Ice Sheet (WAIS) and the Antarctic Peninsula but highlights considerable residual  
181 uncertainty regarding the recent contribution of the East Antarctic Ice Sheet (EAIS) to global  
182 sea-level rise (SLR) (Shepherd et al., 2018; Rignot et al., 2019). For Antarctica there is  
183 relatively little surface melt and subsequent runoff, and surface accumulation has been  
184 relatively stable, although recent reports show an increase in AIS snowfall (Medley and  
185 Thomas, 2019). In Antarctica, the main sustained mass losses are through ice dynamics,  
186 expressed as increased ice discharge across the grounding line. Mass loss through this  
187 mechanism occurs primarily through increased flow speeds of marine terminating glaciers in  
188 the Amundsen and Bellingshausen Sea sectors, which are sensitive to ocean warming,  
189 although superimposed on these relatively gradual changes there are significant short-term,  
190 i.e. interannual to decadal, SMB variations (Rignot et al., 2019). As a key output of the  
191 IMBIE2 project, Shepherd et al. (2018) built on Shepherd et al. (2012) by significantly  
192 extending the study period and reconciling the results of 24 independent estimates of  
193 Antarctic ice-sheet mass balance using satellite altimetry, gravimetry and the mass budget  
194 methods encompassing thirteen satellite missions and approximately double the number of  
195 studies previously considered. They found that between 1992-2017 the Antarctic ice sheets  
196 lost  $2725 \pm 1400$  Gt of ice, therefore contributing  $7.6 \pm 3.9$  mm to SLR, principally due to  
197 increased mass loss from the WAIS and the Antarctic Peninsula. However, they also found  
198 that EAIS was close to balance, i.e.  $5 \pm 46$  Gt yr<sup>-1</sup> averaged over the 25 years, although this  
199 was the least certain region, attributed to its enormous area and relatively poorly constrained  
200 GIA (Section 2.3) compared with other regions. Shepherd et al. (2018) found that WAIS

201 mass loss steadily increased from  $53\pm 29$  Gt yr<sup>-1</sup> for 1992-1996 to  $159\pm 26$  Gt yr<sup>-1</sup> during  
202 2013-2017, and that Antarctic Peninsula mass losses increased by 15 Gt yr<sup>-1</sup> since 2000,  
203 while the EAIS had little overall trend in mass balance during the period of study. The overall  
204 reconciled sea-level contribution from Antarctica rose correspondingly from 0.2 to 0.6 mm  
205 yr<sup>-1</sup>. These authors also reported no systematic Antarctic SMB trend, and they therefore  
206 attributed WAIS mass loss to increased ice discharge. Of particular concern is the case of  
207 ongoing grounding line retreat in the Amundsen Sea in West Antarctica, as well as basal melt  
208 of ice shelves through polynya-related feedbacks, e.g. in the Ross Sea (Stewart et al., 2019).

209 Rignot et al. (2019) used the mass budget method to compare Antarctic snow  
210 accumulation with ice discharge for 1979-2017, using improved, high-resolution datasets of  
211 ice-sheet velocity and thickness, topography and drainage basins and modelled SMB. Within  
212 uncertainties their total mass balance estimates for WAIS and the Antarctic Peninsula agreed  
213 with those of Shepherd et al. (2018) but they derived a  $-57\pm 2$  Gt yr<sup>-1</sup> mass balance for East  
214 Antarctica for 1992-2017, compared with the  $+5\pm 46$  Gt yr<sup>-1</sup> for the same period derived in  
215 IMBIE2. Possible reasons for this difference include uncertainties in ice thickness and  
216 modelled SMB in the mass budget method, together with further uncertainties in the IMBIE-  
217 2 EAIS mass estimates arising from volume to mass conversions within the altimetry data  
218 processing and significantly uncertain GIA corrections when processing GRACE data.  
219 Zwally et al. (2015) found significant EAIS mass gains of  $136 \pm 50$  Gt yr<sup>-1</sup> for 1992-2001  
220 from ERS radar altimetry and  $136 \pm 28$  Gt yr<sup>-1</sup> for 2003-2008 based on ERS radar altimetry  
221 and ICESat laser altimetry, dynamic thickening of  $147 \pm 55$  Gt yr<sup>-1</sup> and  $147 \pm 34$  Gt yr<sup>-1</sup>  
222 respectively, and accumulation-driven losses of  $11 \pm 6$  Gt yr<sup>-1</sup> in both periods with respect to  
223 a 27-year mean. They attributed the dynamic thickening to a long-term dynamic response  
224 arising from a 67-266% increase in snow accumulation during the Holocene, as derived from  
225 six ice cores (Siegert, 2003), rather than contemporaneous increases in accumulation.  
226 However, because the results of Zwally et al. (2015) differ from most others, they have been  
227 questioned by other workers (Scambos and Shuman, 2016; Martín-Español et al., 2017),  
228 although see Zwally et al. (2016) for a response. Bamber et al. (2018) describe “reasonable  
229 consistency between [EAIS mass balance] estimates” if they discount the outlier of Zwally et  
230 al. (2015). Notwithstanding, as highlighted by Hanna et al. (2013) and Shepherd et al. (2018)  
231 and clearly shown here in **Figure 3** which clearly shows ‘outliers’ on both sides of the  
232 IMBIE-reconciled means, disparate estimates of the mass balance of East Antarctica, which  
233 vary by  $\sim 100$  Gt yr<sup>-1</sup>, have not yet been properly resolved. Furthermore, the range of  
234 differences does not appear to be narrowing with time, which indicates a lack of advancement  
235 in one or more of the mass-balance determination methods.

## 236 237 *2.2 Greenland Ice Sheet*

238  
239 According to several recent estimates, the GrIS lost  $257\pm 15$  Gt yr<sup>-1</sup> of mass during 2003-  
240 2015 (Box et al., 2018),  $262\pm 21$  Gt yr<sup>-1</sup> during 2007-2011 (Andersen et al., 2015),  $269\pm 51$   
241 Gt yr<sup>-1</sup> during 2011-2014 (McMillan et al., 2016),  $247$  Gt yr<sup>-1</sup> of mass – representing 37% of  
242 the overall land ice contribution to global sea-level rise – during 2012-2016 (Bamber et al.  
243 2018), and  $286\pm 20$  Gt yr<sup>-1</sup> during 2010-2018 (Mouginot et al., 2019). A slightly greater mass  
244 loss of  $308\pm 12$  Gt yr<sup>-1</sup> based on GRACE gravimetric satellite data for 2007-2016 was given  
245 by Zhang et al. (2019). Some of the difference between these numbers can be attributed to  
246 different methods considering either just the contiguous ice sheet or also including  
247 disconnected peripheral glaciers and ice caps, the latter being the case for GRACE-based  
248 estimates. However, GrIS mass loss approximately quadrupled during 2002/3 to 2012/13  
249 (Bevis et al., 2019). The GrIS sea-level contribution over 1992-2017 was approximately one

250 and a half times the sea-level contribution of Antarctica (Box et al., 2018). However this kind  
251 of average value masks very significant interannual variability of  $\pm 228 \text{ Gt yr}^{-1}$ , and even 5-  
252 year mean values can vary by  $\pm 102 \text{ Gt yr}^{-1}$ , based on 2003-2016 data; for example recent  
253 annual mass losses ranged from  $>400 \text{ Gt}$  in 2012 (a record melt year caused by jet-stream  
254 changes, e.g. Hanna et al., 2014) to  $<100 \text{ Gt}$  just one year later (Bamber et al., 2018).

255 McMillan et al. (2016) found that high interannual (1991-2014) mass balance  
256 variability was mainly due to changes in runoff of  $102 \text{ Gt yr}^{-1}$  (standard deviation,  $\sim 28\%$  of  
257 the mean annual runoff value) with lesser contributions from year-to-year snowfall variations  
258 of  $\sim 61 \text{ Gt yr}^{-1}$  ( $\sim 9\%$  of the mean snowfall value) and solid ice discharge of  $\sim 20 \text{ Gt yr}^{-1}$  ( $\sim 5\%$   
259 of the mean annual discharge). Their interpretation of transient mass changes was supported  
260 by Zhang et al. (2019) who attributed big short-term ( $\sim 3$ -year) fluctuations in surface mass  
261 balance to changes in atmospheric circulation, specifically the Greenland Blocking Index  
262 (GBI; Hanna et al. 2016), with opposite GBI phases in 2010-2012 (highly positive GBI) and  
263 2013-2015 (less blocked Greenland). Also, in the MODIS satellite record since the year 2000,  
264 Greenland albedo was relatively high from 2013-2018 after reaching a record low in 2012  
265 (Tedesco et al., 2018). The relatively low GrIS mass loss in 2013-14 was termed the “pause”  
266 (Bevis et al., 2019). However, Zhang et al. (2019) inferred an acceleration of  $18 \pm 9 \text{ Gt yr}^{-2}$  in  
267 GrIS mass loss over 2007-2016. Given this pronounced recent short-term variability, for  
268 example the recent slowdown of rapid mass loss increases in the 2000s and very early 2010s,  
269 such trends should only be extrapolated forward with great caution.

270 Greenland mass loss is mainly driven by atmospheric warming, and – based on ice-  
271 core-derived melt information and regional model simulations – surface meltwater runoff  
272 increased by  $\sim 50\%$  since the 1990s, becoming significantly higher than pre-industrial levels  
273 and being unprecedented in the last 7000 years (Trusel et al., 2018). Enderlin et al. (2014)  
274 found an increasingly important role of runoff on total mass annual losses during their 2000-  
275 2012 study period and concluded that SMB changes were the main driver of long-term  
276 (decadal or longer) mass loss.

277 However, just five marginal glacier near-termini regions, covering  $<1\%$  of the GrIS  
278 by area were responsible for 12% of the net ice loss (McMillan et al., 2016), highlighting the  
279 potentially important role and sensitivity of ice dynamics; these authors alongside Tedesco et  
280 al. (2016) also found an atmospheric warming signal on mass balance in the northernmost  
281 reaches of the ice sheet. Taking a longer perspective from 1972-2018, using extended  
282 datasets of outlet glacier velocity and ice thickness, improved bathymetric and gravity  
283 surveys and newly-available high resolution SMB model output, Mougintot et al. (2019)  
284 reported that dynamical losses from the GrIS have continuously increased since 1972,  
285 dominating mass changes except for the last 20 years, estimating that over this longer period  
286  $66 \pm 8\%$  of the overall mass losses were from dynamics and  $34 \pm 8\%$  from SMB. They  
287 concluded that dynamics are likely to continue to be important in future decades, apart from  
288 the southwest where runoff/SMB changes predominate, and that the northern parts of GrIS –  
289 where outlet glaciers could lose their buttressing ice shelves – are likely to be especially  
290 sensitive to future climate warming.

291

### 292 *2.3 Glacial Isostatic Adjustment*

293

294 Processes associated with GIA must be accounted for when quantifying contemporary ice-  
295 sheet change (Shepherd et al., 2018) and also when predicting the dynamics of future change  
296 (Adhikari et al., 2014; Gomez et al., 2015; Konrad et al., 2015). Specifically, ongoing  
297 changes to the height of the land surface and the shape of Earth’s gravitational field, in  
298 response to past ice-mass change, will bias gravimetry- and altimeter-based measurements of  
299 contemporary ice mass balance and alter the boundary conditions for ice sheet dynamics. Due

300 to density differences between the ice sheet and the solid Earth, the impact of GIA on  
301 gravimetry measurements will be 4-5 times greater than the impact on altimetry  
302 measurements (Wahr et al., 2000).

303 Numerical models can be used to estimate the geodetic signal associated with GIA  
304 (Whitehouse et al., 2012; Ivins et al., 2013; Argus et al., 2014) or it can be inferred via data  
305 inversion (Gunter et al., 2014; Martín-Español et al., 2016; Sasgen et al., 2017). Both  
306 approaches would benefit from better spatial coverage of GPS observations of land  
307 deformation, while the first approach strongly depends on past ice sheet change, for which  
308 constraints are severely lacking, particularly across the interior of the Greenland and  
309 Antarctic ice sheets. Both approaches also typically rely on the assumption that mantle  
310 viscosity beneath the major ice sheets is spatially uniform and high enough that the signal due  
311 to past ice-mass change is constant in time. However, recent work has revealed regions in  
312 both Greenland and Antarctica where mantle viscosity is much lower than the global average  
313 (e.g. Nield et al., 2014; Khan et al., 2016; Barletta et al., 2018; Mordret, 2018). This has two  
314 important implications. First, in regions where upper mantle viscosity is less than  $\sim 10^{19}$  Pa s  
315 the response to recent (decadal to centennial) ice-mass change will dominate the GIA signal,  
316 and may not be steady in time. In such regions a time-varying GIA correction, which  
317 accounts for both the viscous and elastic response to contemporary ice-mass change, should  
318 be applied to gravimetry, altimetry and other geodetic observations. Secondly, since GIA acts  
319 to reduce the water depth adjacent to a shrinking marine-based ice sheet, this can act to slow  
320 (Gomez et al., 2010) or reverse (Kingslake et al., 2018) the rate of ice loss, with the  
321 stabilising effect being stronger in regions with low upper mantle viscosity (Gomez et al.,  
322 2015; Konrad et al., 2015). To better understand the behaviour and likely future of marine-  
323 based ice masses it will be necessary to quantify the spatially-varying strength of this  
324 stabilising effect and account for feedbacks between GIA and ice dynamics within a coupled  
325 modelling framework (e.g. Pollard et al., 2017; Gomez et al., 2018; Larour et al., 2019;  
326 Whitehouse et al., 2019).

327

### 328 **3.0 Recent advances and challenges in modelling including links with observational** 329 **needs**

330

#### 331 *3.1 Modelling ice-sheet instabilities*

332

333 The marine ice-sheet instability (MISI; **Figure 4**) hypothesises a possible collapse of West  
334 Antarctica as a consequence of global warming. This process, first proposed in the 1970s  
335 (Weertman, 1974; Thomas and Bentley, 1978), was recently theoretically confirmed and  
336 demonstrated in numerical models (Schoof, 2007; Pattyn et al., 2012). It arises from thinning  
337 and eventually flotation of the ice near the grounding line, which moves the latter into deeper  
338 water where the ice is thicker. Thicker ice results in increased ice flux, which further thins  
339 (and eventually floats) the ice, resulting in further retreat into deeper water (and thicker ice)  
340 and so on. This instability is activated when the bedrock deepens toward the interior of the  
341 ice sheet, i.e., a retrograde bed slope, as is the case for most of the West Antarctic ice sheet.  
342 The possibility that some glaciers, such as Pine Island Glacier and Thwaites Glacier, are  
343 already undergoing MISI has been suggested (Rignot et al., 2014; Christianson et al., 2016).  
344 Thwaites Glacier is currently in a less-buttressed state, and several simulations using state-of-  
345 the-art ice-sheet models indicate continued mass loss and possibly MISI or MISI-like  
346 behaviour even under present climatic conditions (Joughin et al., 2014; Nias et al., 2016;  
347 Seroussi et al., 2017). However, rapid grounding line retreat due to MISI or MISI-like  
348 behaviour remains highly dependent on the subtleties of subglacial topography (Waibel et al.,

349 2018) and feedbacks associated with GIA (section 2.3), limiting the predictive behaviour of  
350 the onset of MISI. In other words, geography matters.

351 The marine ice cliff instability (MICI) hypothesises (**Figure 4**) collapse of ice cliffs  
352 that become unstable and fail if higher than  $\sim 90$  m above sea level, leading to the rapid  
353 retreat of ice sheets during past warm (e.g., Pliocene and last interglacial) periods (Pollard et  
354 al., 2015; DeConto and Pollard, 2016). MICI is a process that facilitates and enhances MISI  
355 once the ice shelf has completely disappeared but can also act alone, for instance where the  
356 bed is not retrograde (which prevents MISI). MICI relies on the assumption of perfect plastic  
357 rheology to represent failure. Cliff instability requires an a priori collapse of ice shelves and  
358 is facilitated by hydro-fracturing through the increase of water pressure in surface crevasses  
359 which deepens the latter (Bassis and Walker, 2012; Nick et al., 2013; Pollard et al., 2015).  
360 Whether MICI is necessary to explain Pliocene sea-level high stands has been questioned  
361 recently (Edwards et al., 2019).

362 The introduction of MICI in one ice-sheet model (DeConto and Pollard, 2016) has  
363 profoundly shaken the modelling community, as the mechanism potentially results in future  
364 sea-level rise estimates of almost an order of magnitude larger compared with other studies  
365 (Figure 5 and Table 1). While projected contributions of the Antarctic ice sheet to sea-level  
366 rise by the end of this century for recent studies hover between 0 and 0.45 m (5%-95%  
367 probability range), the MICI model occupies a range of 0.2-1.7 m (Figure 5a). The  
368 discrepancy is even more pronounced for 2300, where the MICI results and other model  
369 estimates no longer agree within uncertainties. Edwards et al. (2019) discuss in detail the  
370 results of DeConto and Pollard (2016), related to cliff collapse but also the sensitivity of the  
371 driving climate model that overestimates surface melt compared to other CMIP5 models.  
372 MICI is a plausible mechanism and is observed on tidewater and outlet glaciers in Greenland  
373 and the Arctic. However, whether and how it applies to very large outlet glaciers of the  
374 Antarctic ice sheet will require further scrutiny. Evidence from paleo-shelf breakup in the  
375 Ross Sea shows that ice-sheet response may be more complicated, including significant lags  
376 in the response of grounding line retreat (Bart et al., 2018). In order to accurately model ice-  
377 sheet instabilities, motion of the grounding line must be accurately represented. International  
378 model inter-comparisons of marine ice-sheet models (MISMIP; MISMIP3d) greatly  
379 improved those models in terms of representing grounding-line migration numerically by  
380 conforming them to known analytical solutions (Pattyn et al., 2012, 2013). These numerical  
381 experiments demonstrated that in order to resolve grounding-line migration in marine ice-  
382 sheet models, a sufficiently high spatial resolution needs to be applied, since membrane  
383 stresses need to be resolved across the grounding line to guarantee mechanical coupling. The  
384 inherent change in basal friction occurring across the grounding line – zero friction below the  
385 ice shelf – requires high spatial resolution (e.g.,  $< 1$  km for Pine Island Glacier; Gladstone et  
386 al., 2012) for an accurate representation of grounding-line migration. Therefore, a series of  
387 ice-sheet models have implemented a spatial grid refinement, mainly for the purpose of  
388 accurate data assimilation (Cornford et al., 2015; Gillet-Chaulet et al., 2012; Morlighem et  
389 al., 2010), but also for further transient simulations where the adaptive mesh approach  
390 enables the finest grid to follow the grounding-line migration (Cornford et al., 2013, 2016).  
391 These higher spatial resolutions of the order of hundreds of meters in the vicinity of  
392 grounding lines also pose new challenges concerning data management for modelling  
393 purposes (Durand et al., 2011).

### 394 395 *3.2 Model initialisation, uncertainty and inter-comparison*

396  
397 Despite major improvements in ice-sheet model sophistication, major uncertainties still  
398 remain pertaining to model initialisation as well as the representation of critical processes

399 such as basal sliding and friction, ice rheology, ice damage (such as calving and MICI) and  
400 sub-shelf melting. New developments in data assimilation methods led to improved  
401 initialisations in which the initial ice-sheet geometry and velocity field are kept as close as  
402 possible to observations by optimising other unknown fields, such as basal friction coefficient  
403 and ice stiffness (accounting for crevasse weakening and ice anisotropy; Arthern and  
404 Hindmarsh, 2006; Arthern and Gudmundsson, 2010; Cornford et al., 2015; MacAyeal, 1992;  
405 Morlighem et al., 2010, 2013). Motivated by the increasing ice-sheet imbalance of the  
406 Amundsen Sea Embayment glaciers over the last 20 years (Shepherd et al., 2018), and  
407 supported by the recent boom in satellite data availability, data-assimilation methods are  
408 progressively used to evaluate unknown time-dependent fields such as basal drag by using  
409 time-evolving states accounting for the transient nature of observations and model dynamics  
410 (Gillet-Chaulet et al., 2016; Goldberg et al., 2013, 2015, 2016).

411 Ensemble model runs equally improve the predictive power of models by translating  
412 uncertainty in a probabilistic framework. The use of statistical emulators thereby increases  
413 the confidence in sampling parameter space (Bulthuis et al., 2019) and helps to reduce  
414 uncertainties in ice dynamical contributions to future sea-level rise (Ritz et al., 2015;  
415 Edwards et al., 2019). Probability distributions for Antarctica are usually not Gaussian and  
416 have a long tail towards high values, especially for high greenhouse warming scenarios  
417 (**Figure 5** and **Table 1**).

418 An important step forward since the Fifth Assessment Report of the IPCC (IPCC,  
419 2013) is that process-based projections of sea-level contributions from both ice sheets are  
420 now organised under the Ice Sheet Model Intercomparison Project for CMIP6 (ISMIP6) and  
421 form an integral part of the CMIP process (Eyring et al., 2016; Nowicki et al., 2016; Goelzer  
422 et al., 2018a; Seroussi et al., 2019). ISMIP6 is working towards providing projections of  
423 future ice-sheet mass changes for the next Assessment Report of the IPCC (AR6). It has  
424 recently finished its first set of experiments focussing on the initial state of the ice sheets as a  
425 starting point for future projections (Goelzer et al., 2018a; Seroussi et al., 2019), which has  
426 seen an unprecedented return from ice-sheet modelling groups globally. With ISMIP6, the  
427 ice-sheet modelling community has engaged to evolve to new standards in availability,  
428 accessibility and transparency of ice-sheet model output data (e.g. Goelzer et al., 2018b),  
429 facilitating model-model and data-model comparison and analysis.

430 ISMIP6 has strengthened the links between the ice-sheet modelling community and  
431 other communities of global and regional climate modellers, ocean modellers and remote  
432 sensing and observations of ice, ocean and atmosphere.

433

### 434 *3.3 Ice sheet model-climate model coupling*

435

436 Fully coupled simulations based on state of the art AOGCMs and ISMs are an emerging field  
437 of active research (e.g. Fyke et al., 2014a; Fischer et al., 2014; Vizcaino et al., 2015; Reerink  
438 et al., 2016; Fyke et al., 2018). This development will help to improve our understanding of  
439 processes and feedbacks due to climate-ice sheet coupling in consistent modelling  
440 frameworks. However, coupling is challenging due to differences in resolution between  
441 climate and ice-sheet models, the computational expense of global climate models, and the  
442 need for advanced snow/firn schemes, etc. (a review of these challenges and recent advances  
443 is given by Vizcaino, 2014). ISMIP6 is also leading and supporting current coupled  
444 modelling efforts (Nowicki et al., 2016).

445 Coupling approaches between atmosphere/ice/ocean/sea ice for the Antarctic ice sheet  
446 have been considerably developed since the AR5 (Asay-Davis et al., 2017; Pattyn et al.,  
447 2017; Favier et al., 2017; Donat-Magnin et al., 2017) but there is still an important need to  
448 document the processes occurring at the interface between ocean and ice. Due to the

449 computational cost, these are limited to a single basin (Seroussi et al., 2017) or intermediate  
450 coupling for the whole ice sheet (Golledge et al., 2019). Observations are currently being  
451 developed to study the ocean characteristics below the ice shelves using autonomous  
452 underwater vehicle (AUVs) or remotely operated vehicle (ROVs) (Jenkins et al., 2010;  
453 Kimura et al., 2016; Nicholls et al., 2006) and should offer critical information for modellers.

454 For the Greenland ice sheet, coupled models have been applied to investigate several  
455 outstanding questions regarding ice-climate interaction, particularly on multi-century and  
456 multi-millennia timescales. Some examples of the topics already addressed include the  
457 impacts of meltwater on ocean circulation (Golledge et al., 2019), regional impact of ice-  
458 sheet area change (Vizcaino et al., 2008, 2010), effect of albedo and cloud change on future  
459 SMB (Vizcaino et al., 2014), and elevation-SMB feedback (Vizcaino et al., 2015). Ongoing  
460 work aims to include more interaction processes, such as the effects of ocean warming on ice-  
461 sheet stability (Straneo et al., 2013).

462 Due to their high computational cost, simulation ensembles (for ice-sheet parameters  
463 as well as climate forcing) are rare in coupled modelling. These ensembles are essential tools  
464 for the attribution of on-going mass loss and to constrain uncertainty in century projections.  
465 Vizcaino et al. (2015) compared 1850-2300 Greenland ice-sheet evolution with a coupled  
466 model forced with three different Representative Concentration Pathways (RCP2.6, RCP4.5  
467 and RCP8.5). For the historical and RCP8.5 scenarios, they performed a small ensemble (size  
468 three). They found a relatively high uncertainty from climate variability in the simulation of  
469 contemporary mass loss. However, this uncertainty was relatively small for the projections as  
470 compared with the uncertainty from greenhouse gas scenario.

471  
472 *3.4 Earth system/regional climate modelling and surface mass balance modelling: advances*  
473 *and challenges*

#### 474 475 3.4.1 General

476  
477 The accuracy of SMB model output naturally depends on observations that are available to  
478 evaluate the models. Recent efforts to collect, synthesise and quality-control in-situ  
479 observations of SMB over the AIS and GrIS have greatly improved our confidence in these  
480 measurements (Favier et al., 2013; Machguth et al., 2016; Montgomery et al., 2018), yet the  
481 observational density remains too low to estimate ice-sheet wide SMB based on interpolation  
482 of these data alone. Uncertainties remain especially large along the ice-sheet margins, where  
483 SMB gradients are steepest and data density lowest because of adverse climate conditions  
484 (Arthern et al., 2006; Bales et al., 2009). Moreover, most in-situ observations constitute an  
485 integrated measurement, providing little insight in SMB component partitioning and seasonal  
486 evolution. Suitable co-located meteorological observations enable time-dependent estimates  
487 of SMB and surface energy balance components such as snow accumulation, sublimation and  
488 melt (van den Broeke et al., 2004, 2011), but especially on the AIS surprisingly few  
489 (automatic) weather stations collect sufficient data to do so. In the GrIS ablation zone, the  
490 PROMICE automatic weather station (AWS) network has recently resolved this problem  
491 (Citterio et al., 2015).

492 Although their performance in simulating ice-sheet SMB is continually improving  
493 (Cullather et al., 2014; Vizcaino et al., 2014; Lenaerts et al., 2016; van Kampenhout et al.,  
494 2017), Earth System Models (ESMs) currently have insufficient (50-100 km) horizontal  
495 resolution in the atmosphere to properly resolve marginal SMB gradients, although  
496 downscaling via elevation classes (Lipscomb et al., 2013; Alexander et al., 2019; Sellevold et  
497 al., submitted), and upcoming variable-resolution ESMs may alleviate this. Moreover, as they  
498 do not assimilate observations, ESMs do not simulate realistic weather. Atmospheric

499 reanalyses have similar low resolution, although this is improved in the recently released  
500 ERA5 reanalysis, but do assimilate meteorological observations, and hence can be used to  
501 force regional climate models (RCMs) at their boundaries. As a result, RCMs provide  
502 reasonably realistic ice-sheet weather at acceptable resolutions: typically 25 km for the full  
503 AIS (van Wessem et al., 2018; Agosta et al., 2019) and 5 km for AIS sub-regions (van  
504 Wessem et al., 2015; Lenaerts et al., 2012; Lenaerts et al., 2018; Datta et al., 2019) and the  
505 GrIS (Lucas-Picher et al., 2012; Fettweis et al., 2017; van den Broeke et al., 2016). Further  
506 statistical downscaling to 1 km resolution is required to resolve SMB over narrow GrIS outlet  
507 glaciers (Noël et al., 2018a). The resulting gridded SMB products cover multiple decades  
508 (1979/1958-present for AIS/GrIS, respectively) at (sub-)daily timescales, allowing synoptic  
509 case studies at the SMB component level but also multidecadal trend analysis. RCM products  
510 also helped to extend ice-sheet SMB time series further back in time by guiding the  
511 interpolation between firn cores (Thomas et al., 2017; Box, 2013).

512 Further improvements are needed: RCMs struggle to realistically simulate (mixed-  
513 phase) clouds (van Tricht et al., 2016) and (sub-) surface processes, such as drifting snow  
514 (Lenaerts et al., 2017), bio-albedo (Stibal et al., 2017) and heterogeneous meltwater  
515 percolation (Steger et al., 2017). A powerful emerging observational technique for dry snow  
516 zones is airborne accumulation radar (Koenig et al., 2016; Lewis et al., 2017), which together  
517 with improved re-analyses products such as MERRA (Cullather et al., 2016) will further  
518 improve our knowledge of contemporary ice-sheet SMB.

519

### 520 3.4.2 Greenland

521

522 Despite considerable advances with RCMs and SMB models, there are significant remaining  
523 biases in absolute values between GrIS SMB simulations for the last few decades. However,  
524 these are expected to be at least partly reconciled through a new SMB Model Intercomparison  
525 Project (SMB\_MIP; Fettweis, 2018) which is standardising model comparisons and  
526 evaluation using in-situ and satellite data (e.g. Machguth et al., 2016). The results of this  
527 exercise should help to improve the models as well as inform on what are the more reliable  
528 model outputs. This exercise may help to resolve significant disagreement between model  
529 reconstructions of GrIS SMB, and especially accumulation, for the last 50-150 years (van den  
530 Broeke et al., 2017).

531 The elevation classes downscaling method has been applied to 1850-2100 GrIS SMB  
532 simulations in several studies with the Community Earth System Model (CESM): these  
533 encompass regional climate and SMB projections (Vizcaino et al., 2014), a freshwater  
534 forcing reconstruction and effect on ocean circulation (Lenaerts et al., 2015), the relationship  
535 between SMB variability and future climate change (Fyke et al., 2014b), and the time of  
536 emergence of an anthropogenic SMB signal from background SMB variability (Fyke et al.,  
537 2014c). The latter study assesses the point in time when the anthropogenic trend in the SMB  
538 becomes larger than the “noise”, and addresses an observational gap given the short records  
539 and/or limited density of remote-sensing/in-situ observations and high GrIS SMB variability  
540 (Wouters et al., 2013). Fyke et al. (2014c) identified a bimodal emergence pattern, with  
541 upward emergence (positive SMB trend) in the interior due to increased accumulation,  
542 downward emergence (negative SMB trend) in the margins due to increased ablation, and an  
543 intermediate area of no emergence due to compensating elevated ablation and accumulation.  
544 This study suggests the Greenland summit as an interesting area to monitor emergence, due  
545 to its high signal-to-noise ratio and resulting early emergence. This high ratio is due to low  
546 SMB variability from drier and colder conditions relative to the margins. These results should  
547 be revisited with further simulations, e.g., from an ensemble and/or multiple models.  
548 Additionally, they should be confronted with available observations of the recent strong SMB

549 decline to identify whether the models adequately represent the causes of this trend (e.g.,  
550 Greenland Blocking, Hanna et al., 2018).

551

### 552 3.4.3 Antarctica

553

554 Shepherd et al. (2018) reveal that present sub-decadal to decadal precipitation and SMB  
555 variations significantly dominate EAIS mass balance variability (Gardner et al., 2018)  
556 justifying the need for further SMB model improvements, validations, and inter-comparisons  
557 (Agosta et al., 2019; Favier et al., 2017). Thanks to observations, the inclusion of several key  
558 processes have been improved in models since AR5, including the roles of the stable  
559 atmospheric boundary layer (Vignon et al., 2017), drifting snow, (Amory et al., 2017; van  
560 Wessem et al., 2018) and supraglacial hydrology (Kingslake et al., 2015, 2017; Hubbard et  
561 al., 2016).

562 A persistent problem is that climate reanalyses used to force regional climate models  
563 still present biases (Bromwich et al., 2011), most noticeably in moisture transport (Dufour et  
564 al., 2019). Constraining atmospheric moisture and cloud microphysics with ground-based  
565 techniques in Antarctica [ceilometer, infrared pyrometer, vertically profiling precipitation  
566 radar (Gorodetskaya et al., 2015), polarimetric weather radar, micro rain radar, weighing  
567 gauges, multi-angle snowflake cameras (Grazioli et al., 2017a), etc.] is necessary to  
568 accurately model cloud evolution and precipitation. Ground-based estimates of cloud  
569 properties and precipitation are only obtained at a few sites, which calls for the use of  
570 distributed remote-sensing techniques to characterise Antarctic precipitation statistics and  
571 rates [e.g., Cloudsat products (Palerm et al., 2014)]. However, processes occurring within 1  
572 km above the surface remain undetected by satellite sensors. In this critical layer for SMB,  
573 sublimation impacts precipitating snowflakes (Grazioli et al., 2017b) and drifting snow  
574 particles (Amory et al., 2017; van Wessem et al., 2018), reducing surface accumulation and  
575 leading to potential feedbacks on atmospheric moisture (Barral et al., 2014). Thus  
576 continental-scale sublimation may be underestimated, suggesting mass balance and SMB  
577 agreement likely relies on some degree of error compensation in models (Agosta et al., 2019).

578 Recent progress has shown that an improved description of the atmospheric structure  
579 is needed during precipitation events; several studies present site-specific results on  
580 precipitation origins [precipitation from synoptic scale systems, hoar frost, diamond dust  
581 (Dittmann et al., 2016; Stenni et al., 2016; Schlosser et al., 2016)] and their impact on the  
582 local SMB. Synoptic-scale precipitation is known to control the inter-annual variability of  
583 accumulation in Dronning Maud Land (Gorodetskaya et al., 2014), Dome C, and Dome F  
584 (Schlosser et al., 2016) through high-intensity precipitation events, but continental-scale  
585 studies for Antarctica are still rare (Turner et al., 2019). High precipitation events are related  
586 to warm and moist air mass intrusions linked to mid-tropospheric planetary waves (Turner et  
587 al. 2016) that are connected with the main modes of atmospheric circulation variability at  
588 southern high-latitudes (Thompson et al., 2011; Turner et al., 2016; Nicolas et al., 2017;  
589 Bromwich et al., 2012). Low-elevation surface melt in West Antarctica (Nicolas et al., 2017;  
590 Scott et al., 2019) and on the Larsen ice shelves (Kuipers Munneke et al., 2018; Bozkurt et  
591 al., 2018) occurs during increased foehn events (Cape et al., 2015) and moisture intrusions  
592 favoured by large synoptic blockings (Scott et al., 2019). These melt-related moisture  
593 intrusions generally occur in the form of atmospheric rivers (Wille et al., 2019). However, the  
594 synoptic causes of these events are still poorly known. Moreover, the feedbacks between  
595 melting and albedo, which may be critical for processes prior to ice shelf collapse (Kingslake  
596 et al., 2017; Bell et al., 2018), are poorly observed in the field. Currently, there is a major gap  
597 between the large scale on which models and remote sensing typically operate (Lenaerts et  
598 al., 2016; Kuipers Munneke et al., 2018) and the local scale, especially regarding snow

599 erosion and redistribution (Amory et al., 2017). These latter processes typically occur at a  
600 decametre scale (Libois et al., 2014; Souverijns et al., 2018), which is not matched by space-  
601 and airborne microwave radar (e.g., between 4 and 6 GHz) or ground penetrating radar  
602 (GPR) (Fujita et al., 2011; Verfaillie et al., 2012; Medley et al., 2013, 2015; Frezzotti et al.,  
603 2007) observations on the kilometre scale that are used to evaluate regional climate models  
604 (Agosta et al., 2019; van Wessem et al., 2018).

605 Despite improvements in regional-scale models, assessing the future SMB of  
606 Antarctica will rely on our capability to produce accurate future projections of the moisture  
607 fluxes towards Antarctica, e.g. linked to changes in sea-ice cover (Bracegirdle et al., 2017;  
608 Krinner et al., 2014; Palerme et al., 2017), and the westerly circulation and atmospheric  
609 blocking patterns around Antarctica (Massom et al., 2004). These aspects are still poorly  
610 represented in CMIP5 simulations (Bracegirdle et al., 2017; Favier et al., 2016). To resolve  
611 this, bias corrections based on nudging approaches or data assimilation schemes have been  
612 proposed, in addition to ensemble approaches (Beaumet et al., 2019; Krinner et al., 2014,  
613 Krinner et al. 2019). To aid these efforts, paleo-climate information on the westerlies  
614 (Saunders et al., 2018), sea ice characteristics (Campagne et al., 2015), temperature (Jones et  
615 al., 2016), and SMB (Thomas et al., 2017) may be useful for constraining the models (Jones  
616 et al., 2016; Abram et al., 2014) and attributing SMB changes to anthropogenic warming.  
617 Emergence of this signal from the natural climate variability of Antarctica is currently  
618 expected between 2020-2050 (Previdi and Polvani, 2016).

619

#### 620 **4.0. Recent and projected mass-balance rates for glaciers and ice caps**

621

622 In this section we target valley glaciers or mountain glaciers and ice caps (<50,000 km<sup>2</sup>). We  
623 here review the advances, since the IPCC AR5, in the estimate of the contribution to SLR of  
624 wastage from these smaller glaciers and ice caps (henceforth, glaciers), as well as its  
625 projections to the end of the 21<sup>st</sup> century. At the time of AR5, the first consensus estimate of  
626 this contribution had just been published (Gardner et al., 2013), and it was estimated to be  
627  $259 \pm 28 \text{ Gt yr}^{-1}$  ( $0.94 \pm 0.08 \text{ mm yr}^{-1}$  SLE) for 2003–2009, including the contribution from the  
628 glaciers in the periphery of Greenland and Antarctica (henceforth, peripheral glaciers). For  
629 the longer period of 1993–2010, AR5 attributed 27% of the SLR to wastage from glaciers  
630 (Church et al., 2013). This was above the combined contribution of the ice sheets of  
631 Antarctica and Greenland (21%), despite the fact that global glacier volume is only ~0.6% of  
632 the combined volume of both ice sheets (Vaughan et al., 2013). Since then, the contribution  
633 to SLR from the ice sheets has accelerated, as discussed in earlier sections, which has  
634 resulted in a current dominance of the ice-sheet contribution despite the contribution from  
635 glaciers having also increased in absolute terms, as will be discussed in this section.

636

##### 637 *4.1 Methods used to estimate the global glacier mass balance*

638

639 For estimating the global mass balance of glaciers, in addition to the techniques already  
640 discussed for ice sheets, such as repeated altimetry (e.g. Moholdt et al., 2010), gravity  
641 observations (e.g. Luthcke et al., 2008), or the mass budget method (e.g. Deschamps-Berger  
642 et al., 2019), other methods are commonly used, which are sometimes variations of those  
643 mentioned above. Purely observation-based techniques include the extrapolation of both in-  
644 situ direct observations by the glaciological method and geodetic mass balance estimates  
645 (Cogley, 2009), as well as reconstructions based on glacier length changes (Leclercq et al.,  
646 2011, 2012, 2014). The glaciological method relies on point measurements of surface mass  
647 balance, which are then integrated to the entire glacier surface (Cogley et al., 2011). Such  
648 measurements are available for a reduced sample of <300 glaciers (Zemp et al., 2015) out of

649 more than 200,000 glaciers inventoried worldwide (Pfeffer et al., 2014), which introduces a  
650 bias when extrapolating to the whole glacierized area of undersampled regions (Gardner et al.,  
651 2013). The geodetic mass balance, in turn, is determined using volume changes from DEM  
652 differencing and then converting to mass changes using an appropriate assumption for the  
653 density (Huss, 2013). The reconstructions based on observed glacier length changes convert  
654 these, upon normalization and averaging to a global mean, to normalized global volume  
655 change. The latter is converted into global glacier mass change using a calibration against  
656 global glacier mass change over a certain period (Leclercq et al., 2011).

657 Finally, the modelling-based approaches for estimating past or current changes are  
658 mostly based on the use of climatic mass balance models forced by either climate  
659 observations or climate model output, calibrated and validated using surface mass-balance  
660 observations. As these techniques are based on a statistical scaling relationship, they are  
661 commonly referred to as statistical modelling, to distinguish them from the use of an RCM to  
662 estimate, directly, the surface mass balance of an ice mass. The latter works well for ice caps,  
663 but not for glaciers, due to their complex topography and corresponding micro-climatological  
664 effects (Bamber et al., 2018). Based on statistical modelling, an analysis of the processes and  
665 feedbacks affecting the global sensitivity of glaciers to climate change can be found in  
666 Marzeion et al. (2014a), while the attribution of the observed mass changes to anthropogenic  
667 and natural causes has been addressed by Marzeion et al. (2014b).

668

#### 669 *4.2 20th century and current estimates*

670

671 Much of the work done since AR5 has focused on improving the estimates for the reference  
672 period 2003-2009 (or some earlier periods), and on producing new estimates for more recent  
673 (or extended) periods. Both the reanalyses and the new estimates have been based on  
674 improvements in the number of mass balance or glacier length changes observations, and on  
675 the use of an increased set of gridded climate observations, and of more complete and  
676 accurate global glacier inventories and global DEMs. These improvements allowed Marzeion  
677 et al. (2015) to achieve the agreement, within error bounds, of the global reconstructions of  
678 the mass losses from glacier wastage for the periods 1961-2005, 1902-2005 and 2003-2009  
679 produced using the various methods available. In spite of the agreement at the global level,  
680 strong disagreements persisted for particular regions such as Svalbard and the Canadian  
681 Arctic, likely because of the omission of calving in the statistical models. Marzeion et al.  
682 (2017), using a yet more extended set of glaciological and geodetic measurements (Zemp et  
683 al., 2015), gave a global glacier mass-change rate estimate of  $-0.61 \pm 0.07$  mm SLE  $\text{yr}^{-1}$  for  
684 2003-2009 (including Greenland peripheral glaciers, but not those of the Antarctic  
685 periphery), obtained by averaging various recent GRACE-based studies (Jacob et al., 2012;  
686 Chen et al., 2013; Yi et al., 2015; Schrama et al., 2014) and several studies combining  
687 GRACE with other datasets (Gardner et al., 2013, and an update of it; Dieng et al., 2015;  
688 Reager et al., 2016; Rietbroek et al., 2016). The studies based on GRACE data consistently  
689 give less negative glacier mass balances than those obtained using other methods.  
690 Uncertainties in the GRACE-derived estimates remain important especially due to the small  
691 size of glaciers compared with the GRACE footprint of  $\sim 300$  km. Associated problems  
692 include the leakage of the gravity signal into the oceans, or the difficulty of distinguishing  
693 between mass changes due to glacier mass changes or to land water storage changes. In  
694 regional and global studies, however, the problem of the footprint and related leakage is not  
695 relevant, as individual glaciers need not to be resolved and GRACE has been shown to be  
696 effective in providing measurements of mass changes for clusters of glaciers (Luthcke et al.,  
697 2008). Uncertainties in the GIA correction also remain, and the effects of rebound from the  
698 Little Ice Age (LIA) deglaciation have to be accounted for.

699 Parkes and Marzeion (2018) have analysed the contribution to SLR from uncharted  
700 glaciers (glaciers melted away and small glaciers not inventoried) during the 21<sup>st</sup> century.  
701 Although they will play a minimal role in SLR in the future, the important finding is that their  
702 contribution is sufficient to close the historical sea-level budget, for which undiscovered  
703 physical processes are then no longer required.

704 Bamber et al. (2018) have updated the glacier mass-change rates presented in  
705 Marzeion et al. (2017) by adding new estimates of mass trends for the Arctic glaciers and ice  
706 caps and the glaciers of High-Mountain Asia and Patagonia, which together contribute to  
707 84% of the SLR from glacier wastage. They combine the most recent observations (including  
708 CryoSat2 radar altimetry) and the latest results from statistical modelling, as well as regional  
709 climate modelling for the Arctic ice caps (Noël et al., 2018b) and stereo photogrammetry for  
710 High-Mountain Asia (Brun et al., 2017). They find poor agreement between the estimates  
711 based on statistical modelling and all other methods (altimetry/gravimetry/RCM) for Arctic  
712 Canada, Svalbard, peripheral Greenland, the Russian Arctic and the Andes, which are all  
713 regions with significant marine- or lake-terminating glaciers, where statistical modelling,  
714 which does not account for frontal ablation, is expected to perform worse than the  
715 observational-based approaches. Bamber et al. (2018) also present pentadal mass balance  
716 rates for the period 1992-2016, which are shown in **Table 2** and clearly illustrate the increase  
717 in global glacier mass losses. If we add to the mass budget for the last pentad (2012-2016) in  
718 **Table 2** the mass budget of  $-33 \text{ Gt yr}^{-1}$  for the Greenland peripheral glaciers estimated by  
719 averaging the CryoSat and RCM values for 2010-2014 given in Table 1 of Bamber et al.  
720 (2018), and the mass budget of  $-6 \text{ Gt yr}^{-1}$  for the Antarctic peripheral glaciers over 2003-  
721 2009 estimated by Gardner et al. (2013), we get an estimate of the current global glacier  
722 mass budget of  $-266 \pm 33 \text{ Gt yr}^{-1}$  ( $0.73 \pm 0.09 \text{ mm SLE yr}^{-1}$ ).

723 The most recent studies to highlight are those of Zemp et al. (2019) and Wouters et al.  
724 (2019). The former is based on glaciological and geodetic measurements but uses a much-  
725 extended dataset (especially for the geodetic measurements), the most updated glacier  
726 inventory (RGI 6.0) and a novel approach. The latter combines, for each glacier region, the  
727 temporal variability from the glaciological sample with the glacier-specific values of the  
728 geodetic sample. The calibrated annual time series is then extrapolated to the whole set of  
729 regional glaciers to assess regional mass changes, considering the rates of area change in the  
730 region. The authors claim that this procedure has overcome the earlier reported negative bias  
731 in the glaciological sample (Gardner et al., 2013). Nevertheless, for large glacierised regions  
732 (e.g. RGI regions), large differences remain between different mass-loss estimates, for  
733 example in the Southern Andes where two recent studies have found reduced mass loss  
734 compared to Zemp et al. (2019) and Wouters et al. (2019) using differencing of digital  
735 elevation models (Braun et al., 2019; Dussaillant et al., 2019). However, the global glacier  
736 mass loss estimate by Zemp et al. (2019), of  $0.74 \pm 0.05 \text{ mm SLE yr}^{-1}$  during 2006-2016,  
737 excluding the peripheral glaciers ( $0.92 \pm 0.39 \text{ mm SLE yr}^{-1}$  if included), is still large compared  
738 to that by Bamber et al. (2018), of  $0.59 \pm 0.11 \text{ mm SLE yr}^{-1}$  for the same period, which is very  
739 similar to the most recent gravimetry-based estimate by Wouters et al. (2019), of  $0.55 \pm 0.10$   
740  $\text{mm SLE yr}^{-1}$ , again for the same period (from their Table S1). This estimate is an  
741 improvement over earlier ones, by using longer time series, an updated glacier inventory  
742 (RGI 6.0), the latest GRACE releases (RL06), which are combined in an ensemble to further  
743 reduce the noise, a new GIA model (Caron et al., 2018) and new hydrology models (GLDAS  
744 V2.1 (Rodell et al., 2004; Beaudoin and Rodell, 2016), and PCR-GLOBW 2 (Sutanudjaja et  
745 al., 2018)) to remove the signal from continental hydrology.

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#### 749 4.3 Projected estimates to the end of the 21<sup>st</sup> century

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751 Among the post-AR5 studies on projected global estimates of mass losses by glaciers to the  
752 end of the 21<sup>st</sup> century, we highlight those of Radić et al. (2014), Huss and Hock (2015) and  
753 Marzeion et al. (2018), together with the main results from the recent model intercomparison  
754 by Hock et al. (2019). An account of other pre- and post-AR5 (up to 2016) projections can be  
755 found in the review by Slangen et al. (2017). While the first two mentioned projections share  
756 many common features (glacier inventory, global climate models and emission scenarios, a  
757 temperature-index mass balance model, similar climate forcing for the calibration period and  
758 similar global DEMs), they have two remarkable differences. First, Radić et al. (2014) rely on  
759 volume-area scaling for the initial volume estimate and to account for the dynamic response  
760 to modelled mass change, while Huss and Hock (2015) derive the initial ice-thickness  
761 distribution using the inverse method by Huss and Farinotti (2012), and the modelled glacier  
762 dynamic response to mass changes is based on an empirical relation between thickness  
763 change and normalized elevation range (Huss et al., 2010). Second, the Huss and Hock  
764 (2015) model accounts for frontal ablation of marine-terminating glaciers, dominated by  
765 calving losses and submarine melt. The results by Radić et al. (2014) suggest SLR  
766 contributions of  $155\pm 41$  (RCP4.5) and  $216\pm 44$  (RCP8.5) mm, similar to the projections of  
767 Marzeion et al. (2012), and to the projections of Slangen and van de Wal (2011) updated in  
768 Slangen et al. (2017). However, the more updated and complete model by Huss and Hock  
769 (2015) predicts lower contributions, of  $79\pm 24$  (RCP2.6),  $108\pm 28$  (RCP4.5), and  $157\pm 31$   
770 (RCP8.5) mm. Of these glacier mass losses, ~10% correspond to frontal ablation globally,  
771 and up to ~30% regionally. In both models, the most important contributors to SLR are the  
772 Canadian Arctic, Alaska, the Russian Arctic, Svalbard, and the periphery of Greenland and  
773 Antarctica. Both models are highly sensitive to the initial ice volume. Regarding Marzeion et  
774 al. (2018), while they use basically the same statistical model as in Marzeion et al. (2012,  
775 2014a,b, 2015, 2017), the use of a newer version (5.0) of the RGI, as well as updated DEMs  
776 and SMB calibration datasets, led to lower SLR contributions from glacier wastage to the end  
777 of the 21<sup>st</sup> century, similar to those by Huss and Hock (2015): 84 [54–116] (RCP2.6), 104  
778 [58–136] (RCP4.5) and 142 [83–165] (RCP8.5) mm (the numbers in brackets indicate the  
779 fifth and ninety-fifth percentiles of the glacier model ensemble distribution).

780 A recent intercomparison of six global-scale glacier mass-balance models,  
781 GlacierMIP (Hock et al., 2019), has provided a total of 214 projections of annual glacier mass  
782 and area, to the end of the 21<sup>st</sup> century, forced by 25 GCMs and four RCPs. Global glacier  
783 mass loss (including Greenland and Antarctic peripheries) by 2100 relative to 2015, averaged  
784 over all model runs, varies between  $94\pm 25$  (RCP2.6) and  $200\pm 44$  (RCP8.5) mm SLE. Large  
785 differences are found between the results from the various models even for identical RCPs,  
786 particularly for some glacier regions. These discrepancies are attributed to differences in  
787 model physics, calibration and downscaling procedures, input data and initial glacier volume,  
788 and the number and ensembles of GCMs used.

789 Although only a regional study, the modelling by Zekollari et al. (2019) is a good  
790 example of one of the lines of improvements expected for the future generation of models for  
791 projecting the future evolution of glaciers. Zekollari et al. (2019) have added ice dynamics to  
792 the model by Huss and Hock (2015), in which glacier changes are imposed based on a  
793 parameterization of the changes in surface elevation at a regional scale. The inclusion of ice  
794 dynamics results in a reduction of the projected mass loss, especially for the low-emission  
795 scenarios such as RCP2.6, and this effect increases with the glacier elevation range, which is  
796 typically broader for the largest glaciers.

797 The contribution from glaciers to SLR is expected to continue to increase during most  
798 of the 21<sup>st</sup> century. Note e.g. that the projections by Huss and Hock (2015) give average rates,

799 over their 90-yr modelled period, between  $0.88\pm 0.27$  and  $1.74\pm 0.34$  mm SLE  $\text{yr}^{-1}$ , depending  
800 on the emission scenario, which are larger than the current rates. However, this contribution  
801 is expected to decay as the total ice volume stored in glaciers becomes smaller as the low-  
802 latitude and low-altitude glaciers disappear and those remaining become confined to the  
803 higher latitudes and altitudes. The projections by Huss and Hock (2015) yield a global glacier  
804 volume loss of 25–48% between 2010 and 2100, depending on the scenario. In parallel, the  
805 contribution from the ice sheets is increasing (e.g. Shepherd et al., 2013, 2018; this paper),  
806 and thus the sea-level rise caused by mass losses from land ice masses will more and more be  
807 dominated by losses from the ice sheets (**Table 3**).

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## 809 **5.0 Summary and outlook**

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811 Never before have there been so much new observational, especially satellite, data for  
812 assessing the state of mass balance of ice sheets and glaciers and their sensitivity to ongoing  
813 climate change. However, the usable satellite record is still relatively short in climate terms.  
814 One of the main remaining challenges is that satellite observations date back only 2-3  
815 decades, which is a very short period for the reference and evaluation of century-scale  
816 projections. Therefore, further extension of the ice-sheet satellite record into the past, for  
817 example through revised processing of earlier albeit lower quality observations following the  
818 method of Trusel et al. (2018), would greatly inform modellers. Also in the same line, and for  
819 the sake of ice-sheet mass and regional climate change detection and attribution, model  
820 evaluation and improved projections, the maintenance and extension of current automatic  
821 weather stations (e.g. Hermann et al., 2018; Smeets et al. 2018) across the ice sheets is of key  
822 interest, with particular emphasis on energy balance stations able to quantify melt energy.

823 Our review highlights that, despite recent efforts, significant discrepancies remain  
824 with respect to absolute mass balance values for the EAIS, and so further studies are  
825 recommended to resolve this matter. Compared to the AIS, for the GrIS, there is a higher  
826 level of agreement, but absolute values vary by  $\sim 100\text{--}300$  Gt  $\text{yr}^{-1}$  between recent years. These  
827 significant fluctuations are mainly due to SMB variability (precipitation and runoff) that are  
828 in turn linked to fluctuations in atmospheric circulation. Ice dynamics may also have an  
829 important role to play in future changes of the GrIS, especially in regions away from the  
830 southwest, and the relative contributions of SMB and dynamics to future mass change remain  
831 unclear.

832 Continued monitoring is vital to resolve these open questions. Apart from ensuring  
833 the continuity of key satellite data provided by missions including GRACE Follow On  
834 (gravimetry) and ICESat2 (altimetry), and carrying out more frequent (annual)  
835 comprehensive inter-comparison assessments of ice-sheet mass balance, the cryospheric and  
836 climate science communities need to enhance existing collaborations on improving regional  
837 climate model and SMB simulations of Antarctica and Greenland (SMB\_MIP being a key  
838 example), and also make further significant improvements to GIA models, as these are some  
839 of the key sources of residual uncertainty underlying current ice-sheet mass balance  
840 estimates.

841 Recent advances in ice-sheet models show major improvements in terms of  
842 understanding of physics and rheology and model initialization, especially thanks to the  
843 wealth of satellite data that has recently become available. However, recent model  
844 intercomparisons (Goelzer et al., 2018a; Seroussi et al., 2019) still point to large process and  
845 parameter uncertainties. Nevertheless, new techniques need to be further explored to improve  
846 initialization methods using both surface elevation and ice velocity changes, allowing for  
847 improved understanding of underlying friction laws and rheological conditions of marine-  
848 terminating glaciers (e.g. Gillet-Chaulet et al., 2016; Gillet-Chaulet, 2019). Given that marine

849 outlet glaciers are especially sensitive to small-change topographic variations, multi-  
850 parameter ensemble modelling and the use of novel emulation methods to evaluate  
851 uncertainty will become an essential tool in ice-sheet modelling. There is a corresponding  
852 need to acquire additional high resolution subglacial topography data to help with  
853 predictions. Several paleo-studies have also emphasized the importance of subglacial  
854 topography in controlling grounding zone location. Jamieson et al. (2012), Batchelor and  
855 Dowdeswell (2015), and Danielson and Bart (2019) all demonstrate that the post-LGM  
856 Antarctic grounding line preferentially stabilized in regions where there are vertical or lateral  
857 topographic restrictions. Meanwhile, in recognition of the remaining limitations of ice-sheet  
858 models, despite significant recent progress, alternative novel approaches including structured  
859 expert judgment are useful to assess the likely impact of ongoing ice-sheet melt on SLR. For  
860 example, Bamber et al. (2019) indicate that a high-emissions greenhouse warming scenario  
861 gives a not insignificant chance of a total  $>2$  m SLR by 2100.

862 Regarding glaciers other than the ice sheets, in spite of recent improvements the  
863 observational database needs to be further extended in space and time. As suggested by Zemp  
864 et al. (2019), emphasis should be on closing data gaps in: 1) regions where glaciers dominate  
865 runoff during warm/dry seasons (tropical Andes and Central Asia), and 2) regions expected to  
866 dominate the future glacier contribution to SLR (Alaska, Arctic Canada, the Russian Arctic  
867 and Greenland and Antarctica peripheries). ICESat-2 and GRACE follow-on missions are  
868 likely to have revolutionary impacts on our knowledge of the mass changes of glaciers and  
869 ice caps, though GIA corrections and LIA deglaciation effects still have room for  
870 improvement. ICESat-2 especially, with its multiple laser beams and precise repeat-track  
871 pointing capability, has the potential to revolutionise our knowledge of mass changes on  
872 small glaciers worldwide. However, there is an unfortunate conflict that is seriously limiting  
873 ICESat-2 collection of precise repeat-track data globally. The current mission operation for  
874 ICESat-2 has systematic off-nadir pointing outside of polar regions to provide denser  
875 mapping of vegetation biomass for a vegetation inventory, despite the fact that such data is  
876 also being collected by the GEDI laser altimeter on the International Space Station. After one  
877 year of ICESat-2 vegetation-inventory mapping, it would be advisable that the mission  
878 operation plan be changed to precise-repeat track pointing to reference tracks globally for  
879 studies of mass changes of glaciers and ice caps, which will also provide improved vegetation  
880 measurements for studies of seasonal and interannual vegetation changes. DEM differencing  
881 from sub-metre resolution optical satellites such as Quickbird, WorldView and Pléiades will  
882 play a key role in geodetic mass-balance estimates (Kronenberg et al. 2016; Melkonian et al.,  
883 2016; Berthier et al., 2014). The discrepancy between the GlacierMIP mass-change  
884 projections from the various models, even under identical emission scenarios, calls for further  
885 standardized intercomparison experiments, where common glacier inventory version, initial  
886 glacier volume, ensemble of GCMs and RCP emission scenarios are prescribed for all models  
887 (Hock et al., 2019). Finally, projections of future contributions to SLR will benefit from  
888 inclusion in the models of ice dynamics, as done by Zekollari et al. (2019).

889

## 890 **Acknowledgements**

891

892 The authors are grateful to WCRP CliC, SCAR and IASC for sponsoring the ISMASS  
893 workshop in Davos, Switzerland, on 15 June 2018 that led to this paper. FN received funding  
894 from grant CTM2017-84441-R of the Spanish State Plan for R&D. KB acknowledges  
895 support from the Fonds de la Recherche Scientifique de Belgique (F.R.S.-FNRS). HG  
896 received funding from the programme of the Netherlands Earth System Science Centre  
897 (NESSC), financially supported by the Dutch Ministry of Education, Culture and Science  
898 (OCW) under grant no. 024.002.001. MV acknowledges support from the European Research

899 Council ERC-StG-678145-CoupledIceClim. EH thanks Jay Zwally for permission to  
900 reproduce Figure 1, and Holly Garner for help with final checking.

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2183 **Table 1.** Probabilistic projections (5th, 25th, 50th, 75th and 95th percentiles) of Antarctic  
 2184 sea-level contribution at 2300 (in metres) under RCP8.5. Colour legend: **L14**: Simulations by  
 2185 Levermann et al. (2014), **G15**: Simulations by Golledge et al. (2015), **DP16**: Simulations by  
 2186 DeConto and Pollard (2016), **DP16BC**: Bias-corrected simulations by DeConto and Pollard  
 2187 (2016), **B19S**: Simulations with Schoof’s parameterisation by Bulthuis et al. (2019), **B19T**:  
 2188 Simulations with Tsai’s parameterisation by Bulthuis et al. (2019), **E19MICI**: Simulations  
 2189 with MICI by Edwards et al. (2019).  
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	5%	25%	50%	75%	95%
<b>L14</b>	0.30	0.64	1.06	1.75	3.54
<b>G15</b>	1.61	2.07	2.28	2.50	2.96
<b>DP16</b>	6.86	7.35	9.05	11.09	11.25
<b>DP16BC</b>	6.94	7.37	9.05	11.08	11.27
<b>B19S</b>	0.27	0.61	1.04	1.47	1.81
<b>B19T</b>	0.59	1.16	1.85	2.55	3.12
<b>E19MICI</b>	7.08	8.28	8.90	9.51	10.71

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2218 **Table 2.** Pentad mass balance rates for all glaciers and ice caps, excluding the peripheral  
 2219 glaciers of Greenland and Antarctica. Modified from Bamber et al. (2018). The contributions  
 2220 from the peripheral glaciers are here excluded because in Bamber et al. (2018) the peripheral  
 2221 glacier contributions are included in those of the corresponding ice sheet because most data  
 2222 sources (many of them from GRACE) do not separate the peripheral glacier contributions.  
 2223 For reference, the mass-change rates during 2003-2009, according to Gardner et al. (2013),  
 2224 were of  $-38 \pm 7$  Gt yr<sup>-1</sup> ( $0.10 \pm 0.02$  mm SLE yr<sup>-1</sup>) for the Greenland peripheral glaciers, and of  
 2225  $-6 \pm 10$  Gt yr<sup>-1</sup> ( $0.02 \pm 0.03$  mm SLE yr<sup>-1</sup>) for the Antarctic peripheral glaciers. According to  
 2226 Zemp et al. (2019), the contributions during 2002-2016 were of  $-51 \pm 17$  Gt yr<sup>-1</sup> ( $0.14 \pm 0.05$   
 2227 mm SLE yr<sup>-1</sup>) for Greenland periphery and  $-14 \pm 108$  Gt yr<sup>-1</sup> ( $0.00 \pm 0.30$  mm SLE yr<sup>-1</sup>) for the  
 2228 Antarctic periphery.  
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Pentad	1992-1996	1997-2001	2002-2006	2007-2011	2012-2016
Gt yr <sup>-1</sup>	$-117 \pm 44$	$-149 \pm 44$	$-173 \pm 33$	$-197 \pm 30$	$-227 \pm 31$
mm SLE yr <sup>-1</sup>	$0.32 \pm 0.12$	$0.42 \pm 0.12$	$0.48 \pm 0.09$	$0.55 \pm 0.08$	$0.63 \pm 0.08$

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2264 **Table 3.** Estimated contributions to sea-level rise by glaciers and by ice sheets over different  
 2265 recent periods. The data sources are indicated. The percentages indicate the relative  
 2266 contributions of the glaciers and of the ice sheets with respect to the total contribution from  
 2267 the landed ice masses.  
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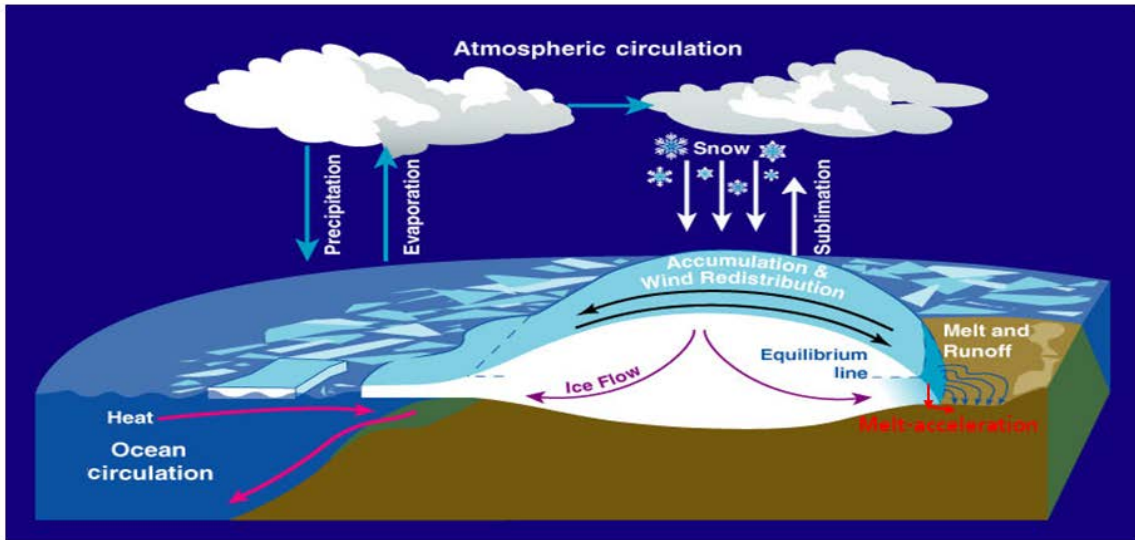
	1993-2010 Church et al. (2013) (IPCC AR5)		2003/05-2009/10 Gardner et al. (2013) Shepherd et al. (2012)		2012-2016 modified from Bamber et al. (2018)	
	mm SLE yr <sup>-1</sup>	%	mm SLE yr <sup>-1</sup>	%	mm SLE yr <sup>-1</sup>	%
Glaciers	0.86	59	0.72	43	0.73 <sup>a</sup>	40 <sup>a,b</sup>
Ice sheets	0.60	41	0.95	57	1.10 <sup>a,b</sup>	60 <sup>a,b</sup>

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 2270 <sup>a</sup> Including the contributions from the peripheral glaciers of Greenland and Antarctica.  
 2271 <sup>b</sup> If the more recent estimate for the Antarctic Ice Sheet by Shepherd et al. (2018) for 2012-  
 2272 2017 were taken instead of that by Bamber et al. (2018) for 2012-2016, the contribution from  
 2273 the ice sheets would increase to 1.29 mm SLE yr<sup>-1</sup> and the relative contributions would be of  
 2274 36% for glaciers and 64% for ice sheets.  
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2308 **Figures**

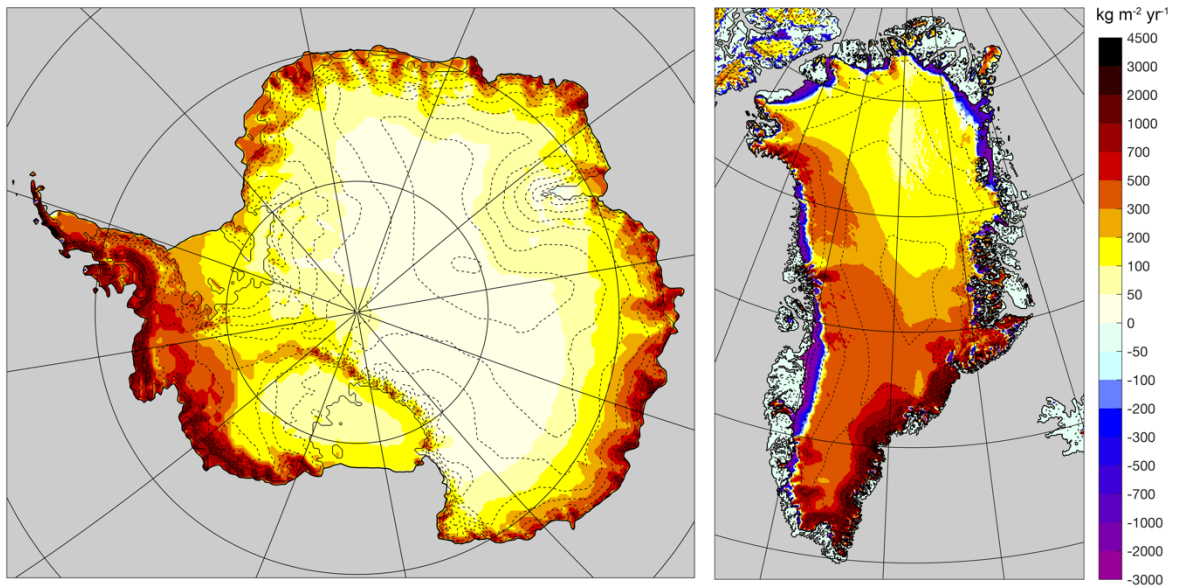
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2310 **Figure 1.** The main processes affecting the mass balance and dynamics of ice sheets. Mass  
2311 input from snowfall is balanced by losses from surface meltwater runoff, sublimation and  
2312 dynamical mass losses (solid ice discharge across the grounding line). Surface melting is  
2313 highly significant for Greenland but for Antarctic grounded ice is very small and subject to  
2314 refreezing. Interaction with the ocean occurs at the undersides of the floating ice shelves and  
2315 glacier tongues, and consequent changes in thickness affect the rate of ice flow from the  
2316 grounded ice. Reproduced from Zwally et al. (2015) with the permission of Jay Zwally.



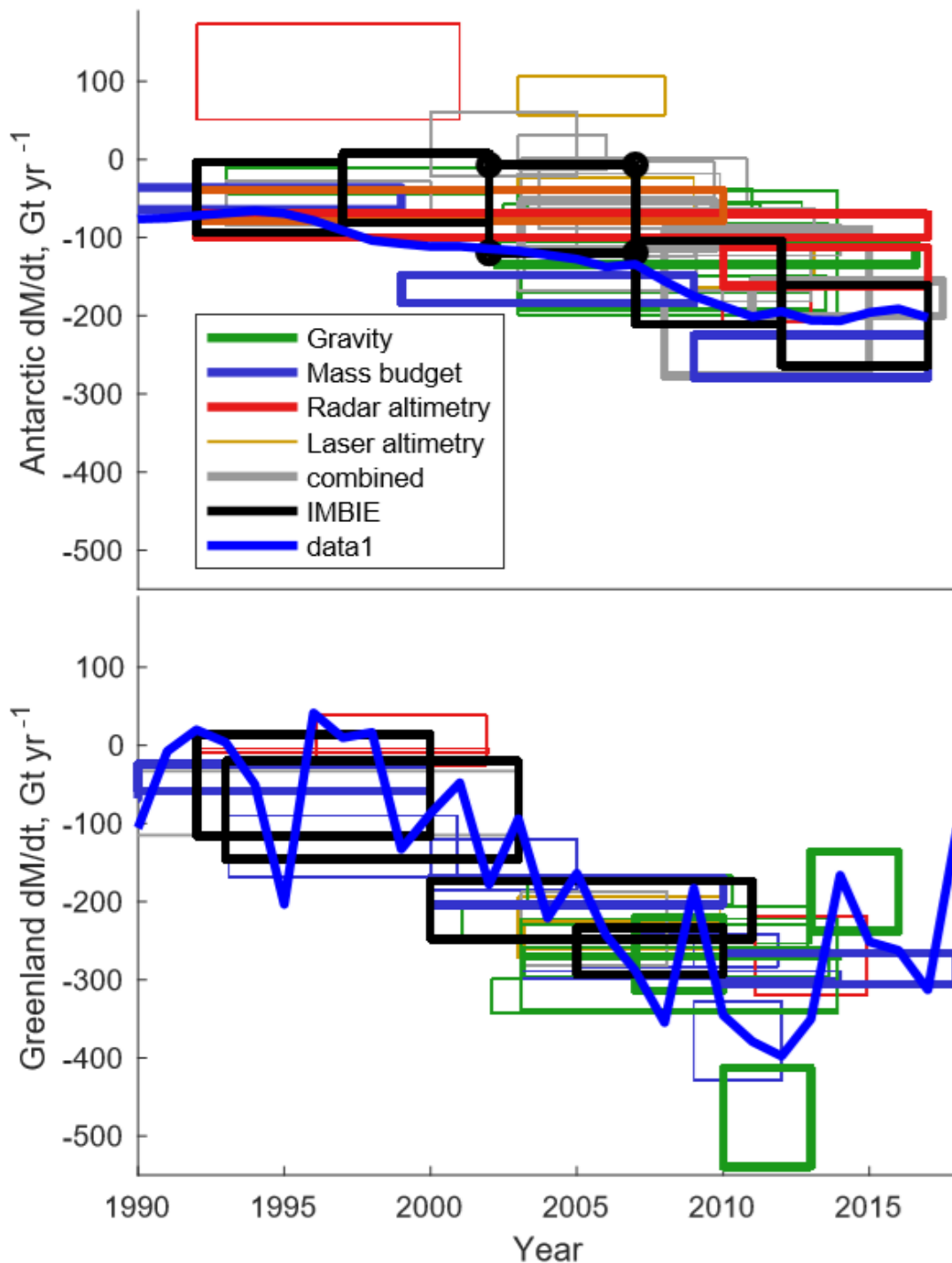
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2335 **Figure 2.** Surface mass balance (averaged over the period 1989-2009) of the Antarctic ice  
2336 sheets (left) and the Greenland Ice Sheet (right) from the regional climate model  
2337 RACMO2.3p2 in  $\text{kg m}^{-2} \text{yr}^{-1}$  (van Wessem et al., 2018; Noël et al., 2018a). Elevation contour  
2338 levels (dashed) are shown every 500 m.



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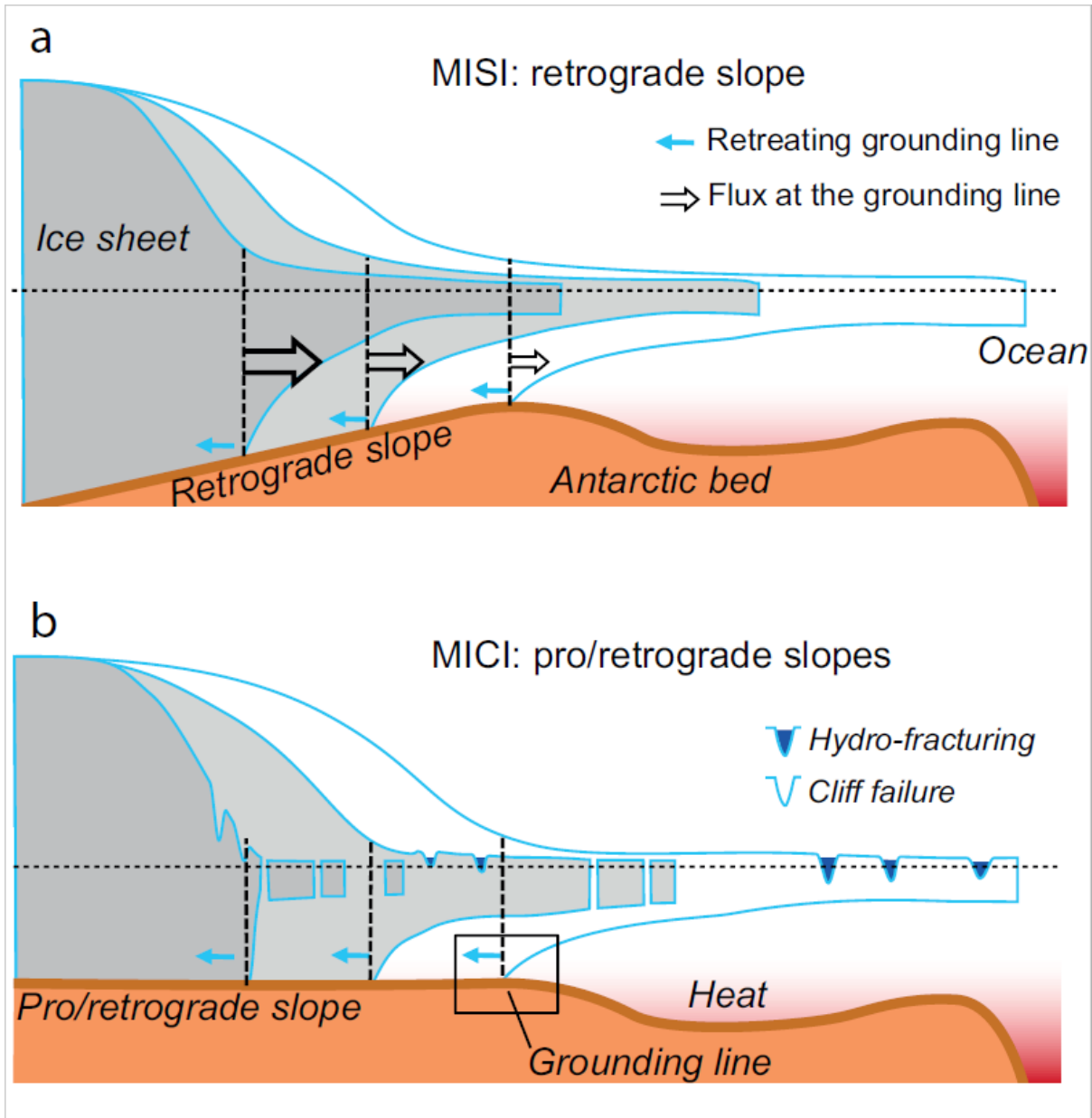
2369 **Figure 3.** Mass rates for the Antarctic (top) and Greenland (bottom) ice sheets derived from  
 2370 published studies. The horizontal extent of each rectangle indicates the period that each  
 2371 estimate spans, while the height indicates the error estimate. Studies published between 2011  
 2372 and 2017 are shown with thin lines, studies published in 2018 and early 2019 with heavier  
 2373 lines. The colour of the lines indicates the type of estimate used, and any estimate that is  
 2374 based explicitly on more than one technique is treated as a ‘combined’ estimate. The  
 2375 IMBIE (Shepherd et. al, 2012 for Greenland, Shepherd et al., 2018 for Antarctica) estimates  
 2376 are shown in black. Rectangles are overplotted with annual mass balance estimates from  
 2377 Rignot et al. (2019) for Antarctica and Mouginot et al. (2019) for Greenland, to indicate  
 2378 interannual variability. The studies cited in this plot are described in Supplemental Table I.  
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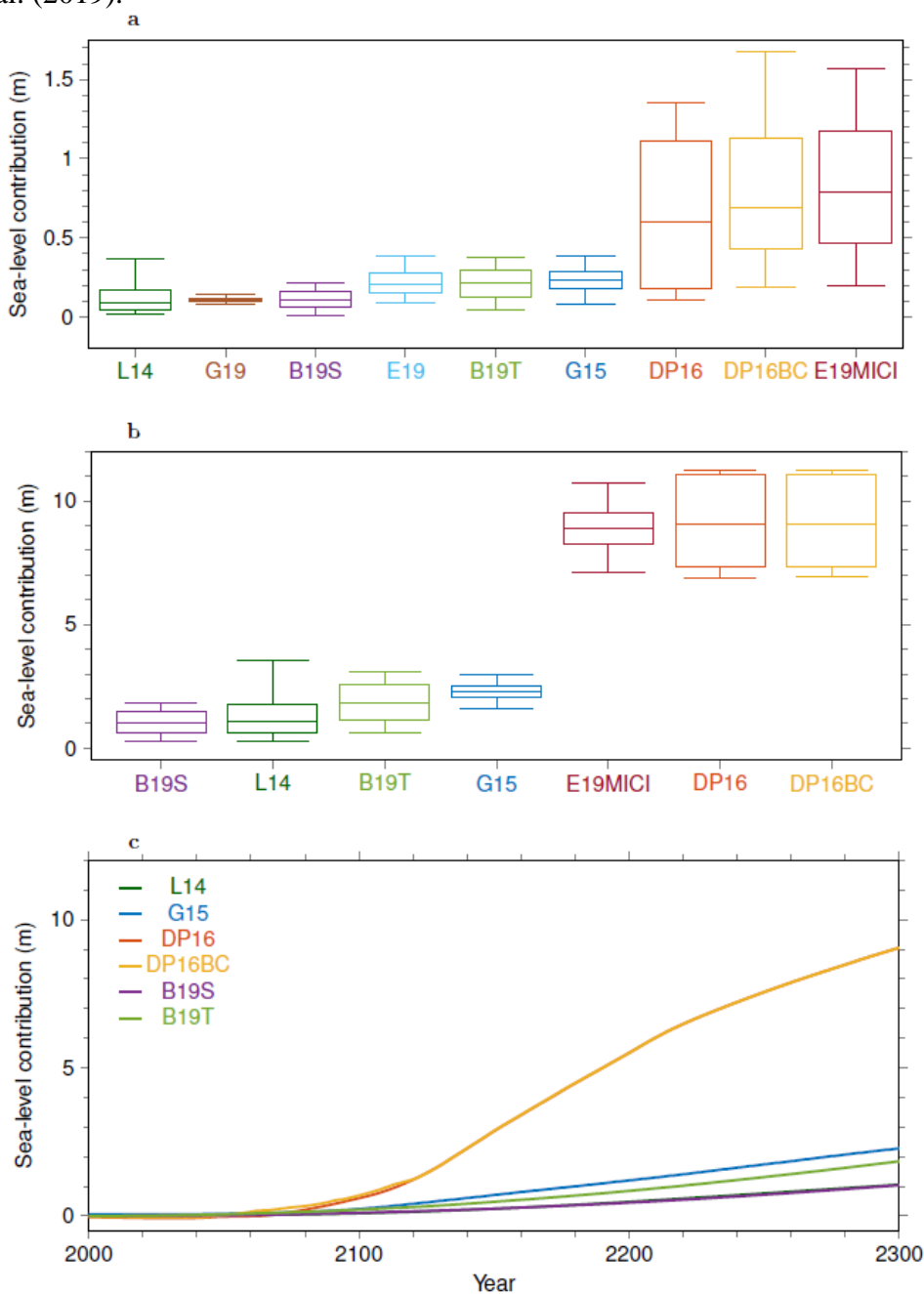
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**Figure 4.** Schematics of (a) Marine Ice Shelf Instability (MISI) and (b) Marine Ice Cliff Instability (MICI). The reader is referred to Section 3.1 for a discussion of MISI/MICI.



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2399 **Figure 5.** Projections of Antarctic sea-level contribution at (a) 2100 and (b) 2300 under  
 2400 RCP8.5. Boxes and whiskers show the 5th, 25th, 50th, 75th and 95th percentiles. The  
 2401 uncertainty range for Golledge et al. (2015) is based on a Gaussian interpretation for the  
 2402 projections with the 5th percentile given by the low scenario and the 95th percentile given by  
 2403 the high scenario. Idem for Golledge et al. (2019) with the 5th percentile given by the  
 2404 simulation without melt feedback and the 95th percentile given by the simulation with melt  
 2405 feedback. (c) Median projections of Antarctic sea-level contribution until 2300 (RCP8.5).  
 2406 Colour legend: **L14**: Simulations by Levermann et al. (2014), **G15**: Simulations by Golledge  
 2407 et al. (2015), **DP16**: Simulations by DeConto and Pollard (2016), **DP16BC**: Bias-corrected  
 2408 simulations by DeConto and Pollard (2016), **B19S**: Simulations with Schoof's  
 2409 parameterisation by Bulthuis et al. (2019), **B19T**: Simulations with Tsai's parameterisation  
 2410 by Bulthuis et al. (2019), **E19**: Simulations without MICI by Edwards et al. (2019),  
 2411 **E19MICI**: Simulations with MICI by Edwards et al. (2019), **G19**: Simulations by Golledge  
 2412 et al. (2019).



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## Supplementary Information

**Supplemental table I.** Details of mass-balance estimates used in Figure 4. Key for measurement type: G = gravimetry, L = laser altimetry, IOM = in/out (mass budget) method, A = airborne photogrammetry, RL and GLRIOM = combined.

### (a) Greenland Ice Sheet

Reference	Year	Type	Time 0	Time 1	Rate	Error
Zwally et al. 2011	2011	R	1992	2002	-7	3
Zwally et al. 2011	2011	L	2003.6	2007.8	-171	4
Shepherd et al. 2012	2012	GLRIOM	1992	2000	-51	65
“	2012	GLRIOM	1993	2003	-83	63
“	2012	GLRIOM	2000	2011	-211	37
“	2012	GLRIOM	2005	2010	-263	30
Wouters et al. 2013	2013	G	2003.1	2012.9	-249	20
Csatho et al. 2014	2014	L	2003.2	2010	-243	18
Enderlin et al. 2014	2014	IOM	2000	2005	-153	33
“	2014	IOM	2005	2009	-265	18
“	2014	IOM	2009	2012	-378	50
Groh et al. 2014	2014	L	2003	2009.9	-233	39
“	2014	G	2001.1	2013	-230	23.5
Hurkmans et al. 2014	2014	R	1996.1	2001.9	6	32.1
“	2014	RL	2003.1	2008.1	-235	47
Schrama et al. 2014	2014	G	2003.2	2013.6	-278	19
Velicogna et al. 2014	2014	G	2003.1	2013.9	-280	58
Andersen et al. 2015	2015	IOM	2007.1	2011.9	-262	21
Kjeldsen et al. 2015	2015	A	1983	2003	-74	41
“	2015	G	2003.3	2010.3	-186	18.9
McMillan et al. 2016	2016	R	2011.1	2014.9	-269	51
van den Broeke et al. 2016	2016	G	2003.1	2014	-270	4
“	2016	IOM	2003.1	2014	-294	5

Talpe et al. 2017	2017	G	2002.1	2013.9	-321	22
“	2017	IOM	1993.1	2000.9	-129	39
Mouginot et al. 2019	2019	IOM	1990	2000	-41.1	17
“	2019	IOM	2000	2010	-186.7	17
“	2019	IOM	2010	2018	-286.2	20
Zhang et al. 2019	2019	G	2007	2010	-267	47
“	2019	G	2010	2013	-476	63
“	2019	G	2013	2016	-187	51

**(b) Antarctic ice sheets**

<b>Reference</b>	<b>Year</b>	<b>Type</b>	<b>Time 0</b>	<b>Time 1</b>	<b>Rate</b>	<b>Error</b>
King et al. 2012	2012	G	2002.7	2010.9	-78	49
Bauer et al. 2013	2013	G	2002.5	2011.4	-104	48
Ivins et al. 2013	2013	G	2003	2012	-57	34
Sasgen et al. 2013	2013	G	2003	2012.7	-114	23
Groh et al. 2014b	2014	L	2003.1	2009.1	-126	39
Groh et al. 2014b	2014	G	2003.1	2009.1	-95	24
Gunter et al. 2014	2014	LG	2003.2	2009.1	-100	44
McMillan et al. 2014	2014	R	2010	2013	-159	48
Memin et al. 2014	2014	GR	2003.1	2010.8	-28	29
Schrama et al. 2014	2014	G	2003.1	2013.5	-171	22
Velicogna et al. 2014	2014	G	2003	2013	-180	10
Williams et al. 2014	2014	G	2003.3	2012.7	-62	7
Gao et al. 2015	2015	G	2003	2013.9	-120	80
Harig and Simons 2015	2015	G	2003.2	2013.6	-92	10
Li et al. 2016	2016	L	2003	2009	-44	21
Zamit-Magion et al. 2015	2015	LRG	2003	2009.9	-47	29
Zwally et al. 2015	2015	L	2003	2008	82	25
Zwally et al. 2015	2015	R	1992	2001	112	61
Jin et al. 2016	2016	G	1993	2002	-28	17

Jin et al. 2016	2016	G	2003	2011	-55	17
Martín-Español et al. 2016	2016	LRG	2003	2013.12	-84	22
Martín-Español et al. 2016	2016	LRG	2003	2006	9	22
Martín-Español et al. 2016	2016	LRG	2007	2009	-104	21
Martín-Español et al. 2016	2016	LRG	2010	2013	-159	22
Peng et al. 2016	2016	G	2002.5	2011.25	-65	7
Sasgen et al. 2017	2017	RG	2003	2013	-141	27
Shepherd et al. 2018	2018	LRG/IO	1992	1997	-48	45
Shepherd et al. 2018	2018	R/IO	1997	2002	-37	44
Shepherd et al. 2018	2018	LRG/IO	2002	2007	-63	56
Shepherd et al. 2018	2018	LRG/IO	2007	2012	-158	53
Shepherd et al. 2018	2018	RG/IO	2012	2017	-213	51
Talpe et al. 2017	2017	G/IO	1993	2000	-56	28
Talpe et al. 2017	2017	G/IO	2000	2005	20	41
Talpe et al. 2017	2017	G/IO	2005	2014	-103	20
Zhang et al. 2017	2017	LGG	2003.7	2009.7	-46	43
Gardner et al. 2018	2018	IO	2008	2015	-183	94
Gao et al. 2019a	2019	G	2002.25	2016.6	-119	16
Gao et al. 2019b	2019	LRG	2003.1	2009.7	-84	31
Rignot et al. 2019	2019	IOM	1979	1989	-40	9
Rignot et al. 2019	2019	IOM	1989	1999	-50	14
Rignot et al. 2019	2019	IOM	1999	2009	-166	18
Rignot et al. 2019	2019	IOM	2009	2017	-252	27
Sasgen et al. 2019	2019	RG	2011	2017.5	-178	23
Schroder et al. 2019	2019	R	1992	2017	-85	15
Schroder et al. 2019	2019	LR	1992	2010	-59	20
Schroder et al. 2019	2019	R	2010	2017	-137	25

