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

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Abstract. Recent research shows increasing decadal ice mass losses from the Greenland and Antarctic Ice Sheets and more generally from glaciers worldwide in the light of continued global warming. Here, in an update of our previous ISMASS paper (Hanna et al., 2013), we review recent observational estimates of ice sheet and glacier mass balance, and their related uncertainties, first briefly considering relevant monitoring methods. Focusing on the response to climate change during 1992-2018, and especially the post-IPCC AR5 period, we discuss recent changes in the relative contributions of ice sheets and glaciers to sea-level change. We assess recent advances in understanding of the relative importance of surface mass balance and ice dynamics in overall ice-sheet mass change. We also consider recent improvements in ice-sheet modelling, highlighting data-model linkages and the use of updated observational datasets in ice-sheet models. Finally, by identifying key deficiencies in the observations and models that hamper current understanding and limit reliability of future ice-sheet projections, we make recommendations to the research community for reducing these knowledge gaps. Our synthesis aims to provide a critical and timely review of the current state of the science in advance of the next Intergovernmental Panel on Climate Change Assessment Report that is 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 (ISMASS; 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 elevation anomaly corrected for the accumulation-driven firn compaction, to derive the total mass change and its accumulation- and dynamic-driven components Secondly, satellite gravimetry effectively weighs the ice sheets through their gravitational pull on a pair of orbiting satellites called GRACE (or, since May 2018, the subsequent GRACE Follow On mission). Thirdly, the mass budget or component method compares SMB model output with multi-sensor satellite radar observations of ice velocity across a position on or close to the grounding line, from which ice discharge can be inferred if the thickness and vertical velocity profile of ice at that point are also assumed/known.All three methods have their strengths and weaknesses (e.g. Hanna et al., 2013; Bamber et al., 2018). Altimetry and, especially, gravimetry, require accurate quantification of Glacial Isostatic Adjustment (GIA; Section 2.3) which contaminates the ice-sheet mass loss signals. Gravimetry is limited by a relatively short time series (since 2002) and low spatial resolution (~300 km) compared with the other methods but is the method that most directly measures mass change.

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

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

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

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

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

2.1 Antarctic ice sheets

Recent work agrees on significant and steadily growing mass losses from the West Antarctic Ice Sheet (WAIS) and the Antarctic Peninsula but highlights considerable residual uncertainty regarding the recent contribution of the East Antarctic Ice Sheet (EAIS) to global sea-level rise (SLR) (Shepherd et al., 2018; Rignot et al., 2019). For Antarctica there is relatively little surface melt and subsequent runoff, and surface accumulation has been relatively stable, although recent reports show an increase in AIS snowfall (Medley and Thomas, 2019). In Antarctica, the main sustained mass losses are through ice dynamics, expressed as increased ice discharge across the grounding line. Mass loss through this mechanism occurs primarily through increased flow speeds of marine terminating glaciers in the Amundsen and Bellingshausen Sea sectors, which are sensitive to ocean warming, although superimposed on these relatively gradual changes there are significant short-term, i.e. interannual to decadal, SMB variations (Rignot et al., 2019). As a key output of the IMBIE2 project, Shepherd et al. (2018) built on Shepherd et al. (2012) by significantly extending the study period and reconciling the results of 24 independent estimates of Antarctic ice-sheet mass balance using satellite altimetry, gravimetry and the mass budget methods encompassing thirteen satellite missions and approximately double the number of studies previously considered. They found that between 1992-2017 the Antarctic ice sheets lost 2725±1400 Gt of ice, therefore contributing 7.6±3.9 mm to SLR, principally due to increased mass loss from the WAIS and the Antarctic Peninsula. However, they also found that EAIS was close to balance, i.e. 5±46 Gt yr^{-1} averaged over the 25 years, although this was the least certain region, attributed to its enormous area and relatively poorly constrained GIA (Section 2.3) compared with other regions. Shepherd et al. (2018) found that WAIS mass loss steadily increased from 53±29 Gt yr^{-1} for 1992-1996 to 159±26 Gt yr^{-1} during 2013-2017, and that Antarctic Peninsula mass losses increased by 15 Gt yr^{-1} since 2000, while the EAIS had little overall trend in mass balance during the period of study. The overall reconciled sea-level contribution from Antarctica rose correspondingly from 0.2 to 0.6 mm yr^{-1}. These authors also reported no systematic Antarctic SMB trend, and they therefore attributed WAIS mass loss to increased ice discharge. Of particular concern is the case of ongoing grounding line retreat in the Amundsen Sea in West Antarctica, as well as basal melt of ice shelves through polynya-related feedbacks, e.g. in the Ross Sea (Stewart et al., 2019).

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

2.2 Greenland Ice Sheet

According to several recent estimates, the GrIS lost 257±15 Gt yr^{-1} of mass during 2003-2015 (Box et al., 2018), 262±21 Gt yr^{-1} during 2007-2011 (Andersen et al., 2015), 269±51 Gt yr^{-1} during 2011-2014 (McMillan et al., 2016), 247 Gt yr^{-1} of mass – representing 37% of the overall land ice contribution to global sea-level rise – during 2012-2016 (Bamber et al. 2018), and 286±20 Gt yr^{-1} during 2010-2018 (Mouginot et al., 2019). A slightly greater mass loss of 308±12 Gt yr^{-1} based on GRACE gravimetric satellite data for 2007-2016 was given by Zhang et al. (2019). Some of the difference between these numbers can be attributed to different methods considering either just the contiguous ice sheet or also including disconnected peripheral glaciers and ice caps, the latter being the case for GRACE-based estimates. However, GrIS mass loss approximately quadrupled during 2002/3 to 2012/13 (Bevis et al., 2019). The GrIS sea-level contribution over 1992-2017 was approximately one and a half times the sea-level contribution of Antarctica (Box et al., 2018). However this kind of average value masks very significant interannual variability of ±228 Gt yr^{-1}, and even 5-year mean values can vary by ±102 Gt yr^{-1}, based on 2003-2016 data; for example recent annual mass losses ranged from >400 Gt in 2012 (a record melt year caused by jet-stream changes, e.g. Hanna et al., 2014) to <100 Gt just one year later (Bamber et al., 2018).

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

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

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

2.3 Glacial Isostatic Adjustment

Processes associated with GIA must be accounted for when quantifying contemporary ice-sheet change (Shepherd et al., 2018) and also when predicting the dynamics of future change (Adhikari et al., 2014; Gomez et al., 2015; Konrad et al., 2015). Specifically, ongoing changes to the height of the land surface and the shape of Earth's gravitational field, in response to past ice-mass change, will bias gravimetry- and altimeter-based measurements of contemporary ice mass balance and alter the boundary conditions for ice sheet dynamics. Due to density differences between the ice sheet and the solid Earth, the impact of GIA on gravimetry measurements will be 4-5 times greater than the impact on altimetry measurements (Wahr et al., 2000).

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

3.0 Recent advances and challenges in modelling including links with observational needs

3.1 Modelling ice-sheet instabilities

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

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

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

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

3.2 Model initialisation, uncertainty and inter-comparison

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

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

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

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

3.3 Ice sheet model-climate model coupling

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

Coupling approaches between atmosphere/ice/ocean/sea ice for the Antarctic ice sheet have been considerably developed since the AR5 (Asay-Davis et al., 2017; Pattyn et al., 2017; Favier et al., 2017; Donat-Magnin et al., 2017) but there is still an important need to document the processes occurring at the interface between ocean and ice. Due to the computational cost, these are limited to a single basin (Seroussi et al., 2017) or intermediate coupling for the whole ice sheet (Golledge et al., 2019). Observations are currently being developed to study the ocean characteristics below the ice shelves using autonomous underwater vehicle (AUVs) or remotely operated vehicle (ROVs) (Jenkins et al., 2010; Kimura et al., 2016; Nicholls et al., 2006) and should offer critical information for modellers.

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

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

3.4 Earth system/regional climate modelling and surface mass balance modelling: advances and challenges

3.4.1 General

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

Although their performance in simulating ice-sheet SMB is continually improving (Cullather et al., 2014; Vizcaino et al., 2014; Lenaerts et al., 2016; van Kampenhout et al., 2017), Earth System Models (ESMs) currently have insufficient (50-100 km) horizontal resolution in the atmosphere to properly resolve marginal SMB gradients, although downscaling via elevation classes (Lipscomb et al., 2013; Alexander et al., 2019; Sellevold et al., submitted), and upcoming variable-resolution ESMs may alleviate this. Moreover, as they do not assimilate observations, ESMs do not simulate realistic weather. Atmospheric reanalyses have similar low resolution, although this is improved in the recently released ERA5 reanalysis, but do assimilate meteorological observations, and hence can be used to force regional climate models (RCMs) at their boundaries. As a result, RCMs provide reasonably realistic ice-sheet weather at acceptable resolutions: typically 25 km for the full AIS (van Wessem et al., 2018; Agosta et al., 2019) and 5 km for AIS sub-regions (van Wessem et al., 2015; Lenaerts et al., 2012; Lenaerts et al., 2018; Datta et al., 2019) and the GrIS (Lucas-Picher et al., 2012; Fettweis et al., 2017; van den Broeke et al., 2016). Further statistical downscaling to 1 km resolution is required to resolve SMB over narrow GrIS outlet glaciers (Noël et al., 2018a). The resulting gridded SMB products cover multiple decades (1979/1958-present for AIS/GrIS, respectively) at (sub-)daily timescales, allowing synoptic case studies at the SMB component level but also multidecadal trend analysis. RCM products also helped to extend ice-sheet SMB time series further back in time by guiding the interpolation between firn cores (Thomas et al., 2017; Box, 2013).

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

3.4.2 Greenland

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

The elevation classes downscaling method has been applied to 1850-2100 GrIS SMB simulations in several studies with the Community Earth System Model (CESM): these encompass regional climate and SMB projections (Vizcaino et al., 2014), a freshwater forcing reconstruction and effect on ocean circulation (Lenaerts et al., 2015), the relationship between SMB variability and future climate change (Fyke et al., 2014b), and the time of emergence of an anthropogenic SMB signal from background SMB variability (Fyke et al., 2014c). The latter study assesses the point in time when the anthropogenic trend in the SMB becomes larger than the “noise”, and addresses an observational gap given the short records and/or limited density of remote-sensing/in-situ observations and high GrIS SMB variability (Wouters et al., 2013). Fyke et al. (2014c) identified a bimodal emergence pattern, with upward emergence (positive SMB trend) in the interior due to increased accumulation, downward emergence (negative SMB trend) in the margins due to increased ablation, and an intermediate area of no emergence due to compensating elevated ablation and accumulation. This study suggests the Greenland summit as an interesting area to monitor emergence, due to its high signal-to-noise ratio and resulting early emergence. This high ratio is due to low SMB variability from drier and colder conditions relative to the margins. These results should be revisited with further simulations, e.g., from an ensemble and/or multiple models. Additionally, they should be confronted with available observations of the recent strong SMB decline to identify whether the models adequately represent the causes of this trend (e.g., Greenland Blocking, Hanna et al., 2018).

3.4.3 Antarctica

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

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

Recent progress has shown that an improved description of the atmospheric structure is needed during precipitation events; several studies present site-specific results on precipitation origins [precipitation from synoptic scale systems, hoar frost, diamond dust (Dittmann et al., 2016; Stenni et al., 2016; Schlosser et al., 2016)] and their impact on the local SMB. Synoptic-scale precipitation is known to control the inter-annual variability of accumulation in Dronning Maud Land (Gorodetskaya et al., 2014), Dome C, and Dome F (Schlosser et al., 2016) through high-intensity precipitation events, but continental-scale studies for Antarctica are still rare (Turner et al., 2019). High precipitation events are related to warm and moist air mass intrusions linked to mid-tropospheric planetary waves (Turner et al. 2016) that are connected with the main modes of atmospheric circulation variability at southern high-latitudes (Thompson et al., 2011; Turner et al., 2016; Nicolas et al., 2017; Bromwich et al., 2012). Low-elevation surface melt in West Antarctica (Nicolas et al., 2017; Scott et al., 2019) and on the Larsen ice shelves (Kuipers Munneke et al., 2018; Bozkurt et al., 2018) occurs during increased foehn events (Cape et al., 2015) and moisture intrusions favoured by large synoptic blockings (Scott et al., 2019). These melt-related moisture intrusions generally occur in the form of atmospheric rivers (Wille et al., 2019). However, the synoptic causes of these events are still poorly known. Moreover, the feedbacks between melting and albedo, which may be critical for processes prior to ice shelf collapse (Kingslake et al., 2017; Bell et al;, 2018), are poorly observed in the field. Currently, there is a major gap between the large scale on which models and remote sensing typically operate (Lenaerts et al., 2016; Kuipers Munneke et al., 2018) and the local scale, especially regarding snow erosion and redistribution (Amory et al., 2017). These latter processes typically occur at a decametre scale (Libois et al., 2014; Souverijns et al., 2018), which is not matched by space- and airborne microwave radar (e.g., between 4 and 6 GHz) or ground penetrating radar (GPR) (Fujita et al., 2011; Verfaillie et al., 2012; Medley et al., 2013, 2015; Frezzotti et al., 2007) observations on the kilometre scale that are used to evaluate regional climate models (Agosta et al., 2019; van Wessem et al., 2018).

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

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

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

4.1 Methods used to estimate the global glacier mass balance

For estimating the global mass balance of glaciers, in addition to the techniques already discussed for ice sheets, such as repeated altimetry (e.g. Moholdt et al., 2010), gravity observations (e.g. Luthcke et al., 2008), or the mass budget method (e.g. Deschamps-Berger et al., 2019), other methods are commonly used, which are sometimes variations of those mentioned above. Purely observation-based techniques include the extrapolation of both in-situ direct observations by the glaciological method and geodetic mass balance estimates (Cogley, 2009), as well as reconstructions based on glacier length changes (Leclercq et al., 2011, 2012, 2014). The glaciological method relies on point measurements of surface mass balance, which are then integrated to the entire glacier surface (Cogley et al., 2011). Such measurements are available for a reduced sample of <300 glaciers (Zemp et al., 2015) out of more than 200,000 glaciers inventoried worldwide (Pfeffer et al., 2014), which introduces a bias when extrapolating to the whole glacierized area of undersampled regions (Gardner et al, 2013). The geodetic mass balance, in turn, is determined using volume changes from DEM differencing and then converting to mass changes using an appropriate assumption for the density (Huss, 2013). The reconstructions based on observed glacier length changes convert these, upon normalization and averaging to a global mean, to normalized global volume change. The latter is converted into global glacier mass change using a calibration against global glacier mass change over a certain period (Leclercq et al., 2011).

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

4.2 20th century and current estimates

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

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

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

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

4.3 Projected estimates to the end of the 21st century

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

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

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

The contribution from glaciers to SLR is expected to continue to increase during most of the 21st century. Note e.g. that the projections by Huss and Hock (2015) give average rates, over their 90-yr modelled period, between 0.88±0.27 and 1.74±0.34 mm SLE yr^{-1}, depending on the emission scenario, which are larger than the current rates. However, this contribution is expected to decay as the total ice volume stored in glaciers becomes smaller as the low-latitude and low-altitude glaciers disappear and those remaining become confined to the higher latitudes and altitudes. The projections by Huss and Hock (2015) yield a global glacier volume loss of 25–48% between 2010 and 2100, depending on the scenario. In parallel, the contribution from the ice sheets is increasing (e.g. Shepherd et al., 2013, 2018; this paper), and thus the sea-level rise caused by mass losses from land ice masses will more and more be dominated by losses from the ice sheets (**Table 3**).

5.0 Summary and outlook

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

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

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

Recent advances in ice-sheet models show major improvements in terms of understanding of physics and rheology and model initialization, especially thanks to the wealth of satellite data that has recently become available. However, recent model intercomparisons (Goelzer et al., 2018a; Seroussi et al., 2019) still point to large process and parameter uncertainties. Nevertheless, new techniques need to be further explored to improve initialization methods using both surface elevation and ice velocity changes, allowing for improved understanding of underlying friction laws and rheological conditions of marine-terminating glaciers (e.g. Gillet-Chaulet et al., 2016; Gillet-Chaulet, 2019). Given that marine outlet glaciers are especially sensitive to small-change topographic variations, multi-parameter ensemble modelling and the use of novel emulation methods to evaluate uncertainty will become an essential tool in ice-sheet modelling. There is a corresponding need to acquire additional high resolution subglacial topography data to help with predictions. Several paleo-studies have also emphasized the importance of subglacial topography in controlling grounding zone location. Jamieson et al. (2012), Batchelor and Dowdeswell (2015), and Danielson and Bart (2019) all demonstrate that the post-LGM Antarctic grounding line preferentially stabilized in regions where there are vertical or lateral topographic restrictions. Meanwhile, in recognition of the remaining limitations of ice-sheet models, despite significant recent progress, alternative novel approaches including structured expert judgment are useful to assess the likely impact of ongoing ice-sheet melt on SLR. For example, Bamber et al. (2019) indicate that a high-emissions greenhouse warming scenario gives a not insignificant chance of a total >2 m SLR by 2100.

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

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Table 1. Probabilistic projections (5th, 25th, 50th, 75th and 95th percentiles) of Antarctic sea-level contribution at 2300 (in metres) under RCP8.5. Colour legend: L14: Simulations by Levermann et al. (2014), G15: Simulations by Golledge et al. (2015), DP16: Simulations by DeConto and Pollard (2016), DP16BC: Bias-corrected simulations by DeConto and Pollard (2016), B19S: Simulations with Schoof's parameterisation by Bulthuis et al. (2019), B19T: Simulations with Tsai's parameterisation by Bulthuis et al. (2019), E19MICI: Simulations with MICI by Edwards et al. (2019).

	5%	25%	50%	75%	95%
L14	0.30	0.64	1.06	1.75	3.54
G15	1.61	2.07	2.28	2.50	2.96
DP16	6.86	7.35	9.05	11.09	11.25
DP16BC	6.94	7.37	9.05	11.08	11.27
B19S	0.27	0.61	1.04	1.47	1.81
B19T	0.59	1.16	1.85	2.55	3.12
E19MICI	7.08	8.28	8.90	9.51	10.71

	5%	25%	50%	75%	95%
L14	0.30	0.64	1.06	1.75	3.54
G15	1.61	2.07	2.28	2.50	2.96
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E19MICI	7.08	8.28	8.90	9.51	10.71

Table 2. Pentad mass balance rates for all glaciers and ice caps, excluding the peripheral glaciers of Greenland and Antarctica. Modified from Bamber et al. (2018). The contributions from the peripheral glaciers are here excluded because in Bamber et al. (2018) the peripheral glacier contributions are included in those of the corresponding ice sheet because most data sources (many of them from GRACE) do not separate the peripheral glacier contributions. For reference, the mass-change rates during 2003-2009, according to Gardner et al. (2013), were of −38±7 Gt yr^{-1} (0.10±0.02 mm SLE yr^{-1}) for the Greenland peripheral glaciers, and of −6±10 Gt yr^{-1} (0.02±0.03 mm SLE yr^{-1}) for the Antarctic peripheral glaciers. According to Zemp et al. (2019), the contributions during 2002-2016 were of −51±17 Gt yr^{-1} (0.14±0.05 mm SLE yr^{-1}) for Greenland periphery and −14±108 Gt yr^{-1} (0.00±0.30 mm SLE yr^{-1}) for the Antarctic periphery.

Pentad	1992-1996	1997-2001	2002-2006	2007-2011	2012-2016
Gt yr^{-1}	−117 ± 44	−149 ± 44	−173 ± 33	−197 ± 30	−227 ± 31
mm SLE yr^{-1}	0.32 ± 0.12	0.42 ± 0.12	0.48 ± 0.09	0.55 ± 0.08	0.63 ± 0.08

Table 3. Estimated contributions to sea-level rise by glaciers and by ice sheets over different recent periods. The data sources are indicated. The percentages indicate the relative contributions of the glaciers and of the ice sheets with respect to the total contribution from the landed ice masses.

	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^{-1}	%	mm SLE yr^{-1}	%	mm SLE yr^{-1}	%
Glaciers	0.86	59	0.72	43	0.73 [a]	40 [a,b]
Ice sheets	0.60	41	0.95	57	1.10 [a,b]	60 [a,b]

[a] Including the contributions from the peripheral glaciers of Greenland and Antarctica.
[b] If the more recent estimate for the Antarctic Ice Sheet by Shepherd et al. (2018) for 2012-2017 were taken instead of that by Bamber et al. (2018) for 2012-2016, the contribution from the ice sheets would increase to 1.29 mm SLE yr^{-1} and the relative contributions would be of 36% for glaciers and 64% for ice sheets.

Figures

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

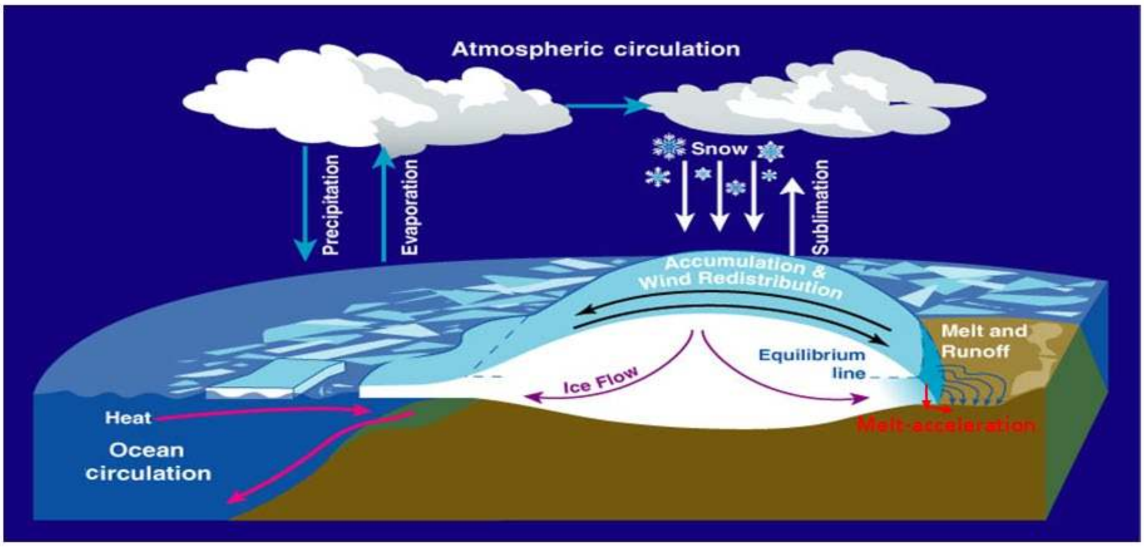


Figure 2. Surface mass balance (averaged over the period 1989-2009) of the Antarctic ice sheets (left) and the Greenland Ice Sheet (right) from the regional climate model RACMO2.3p2 in kg m^{-2} yr^{-1} (van Wessem et al., 2018; Noël et al., 2018a). Elevation contour levels (dashed) are shown every 500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

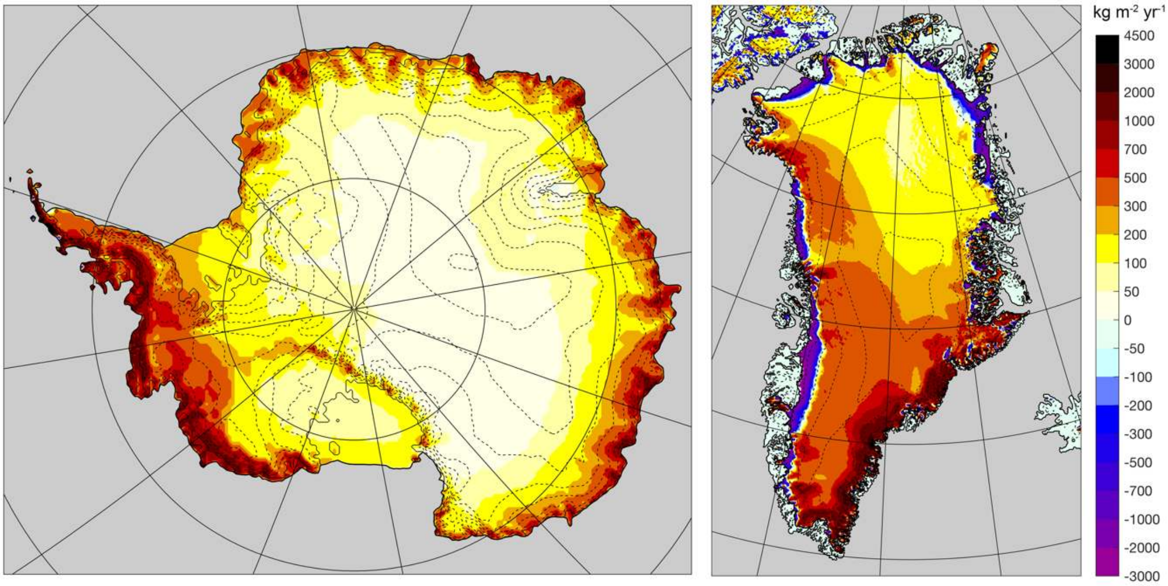


Figure 3. Mass rates for the Antarctic (top) and Greenland (bottom) ice sheets derived from published studies. The horizontal extent of each rectangle indicates the period that each estimate spans, while the height indicates the error estimate. Studies published between 2011 and 2017 are shown with thin lines, studies published in 2018 and early 2019 with heavier lines. The colour of the lines indicates the type of estimate used, and any estimate that is based explicitly on more than one technique is treated as a 'combined' estimate. The IMBIE (Shepherd et. al, 2012 for Greenland, Shepherd et al., 2018 for Antarctica) estimates are shown in black. Rectangles are overplotted with annual mass balance estimates from Rignot et al. (2019) for Antarctica and Mouginot et al. (2019) for Greenland, to indicate interannual variability. The studies cited in this plot are described in Supplemental Table I.

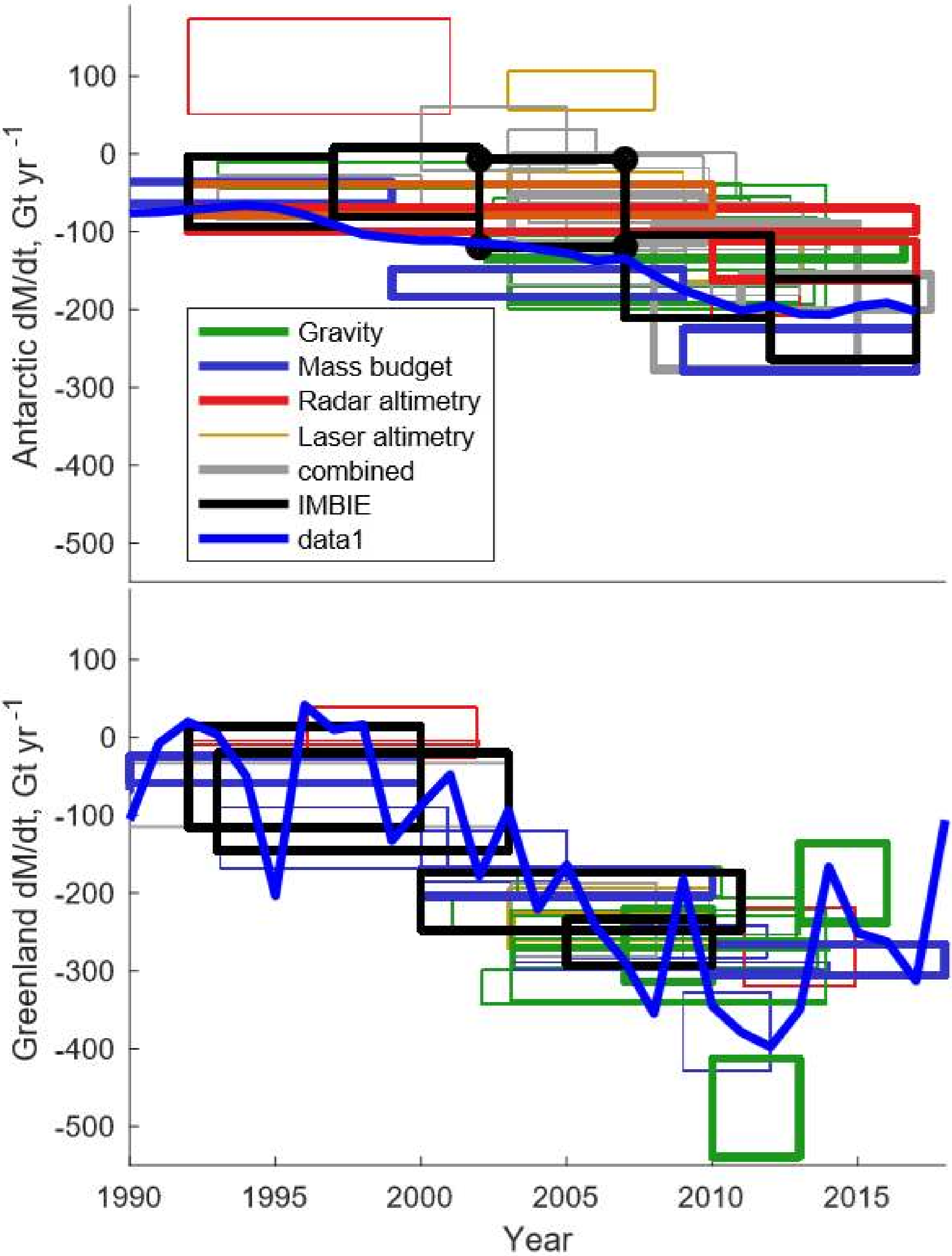


Figure 4. Schematics of (a) Marine Ice Shelf Instability (MISI) and (b) Marine Ice Cliff Instability (MICI). The reader is referred to Section 3.1 for a discussion of MISI/MICI.

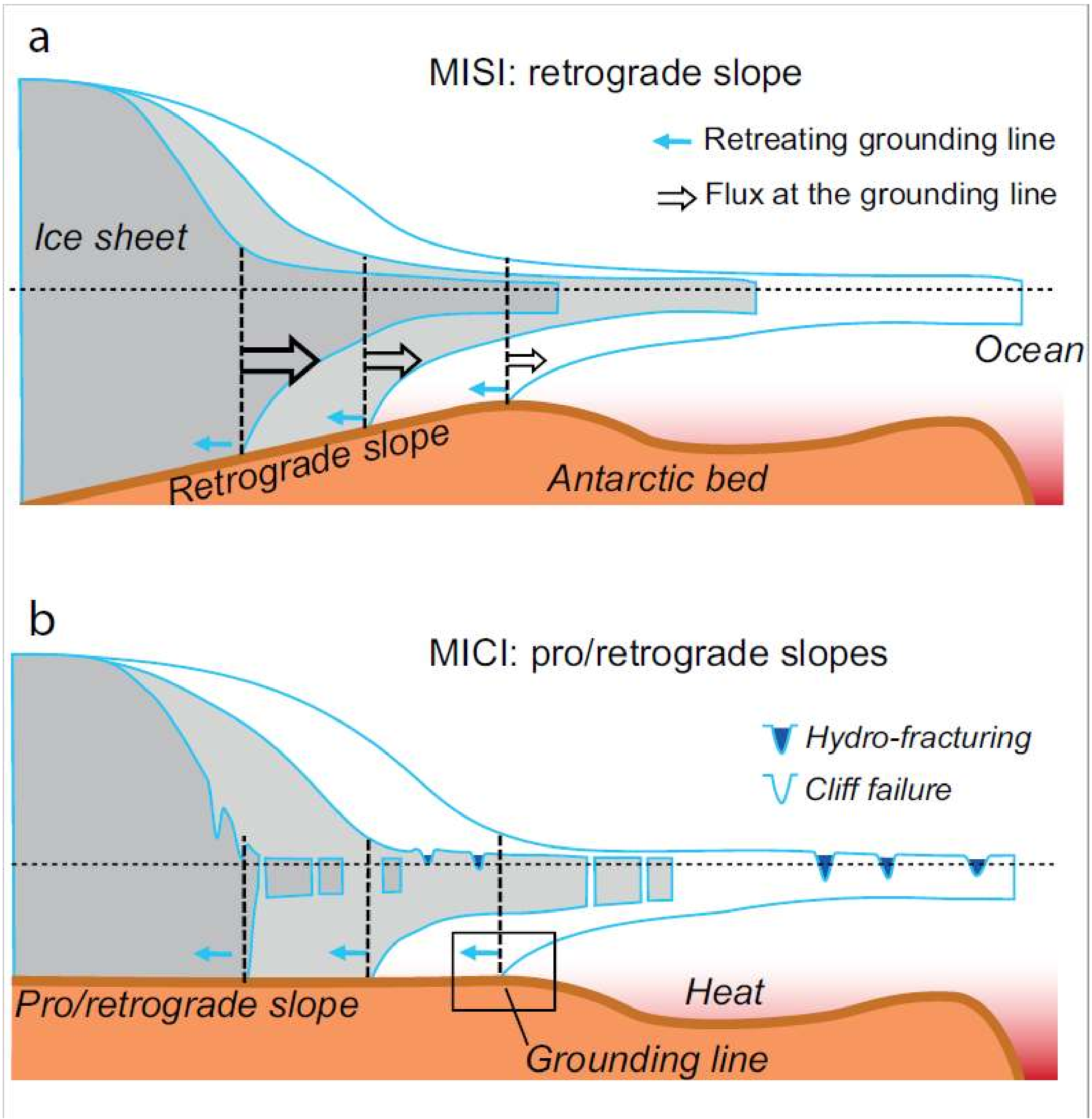
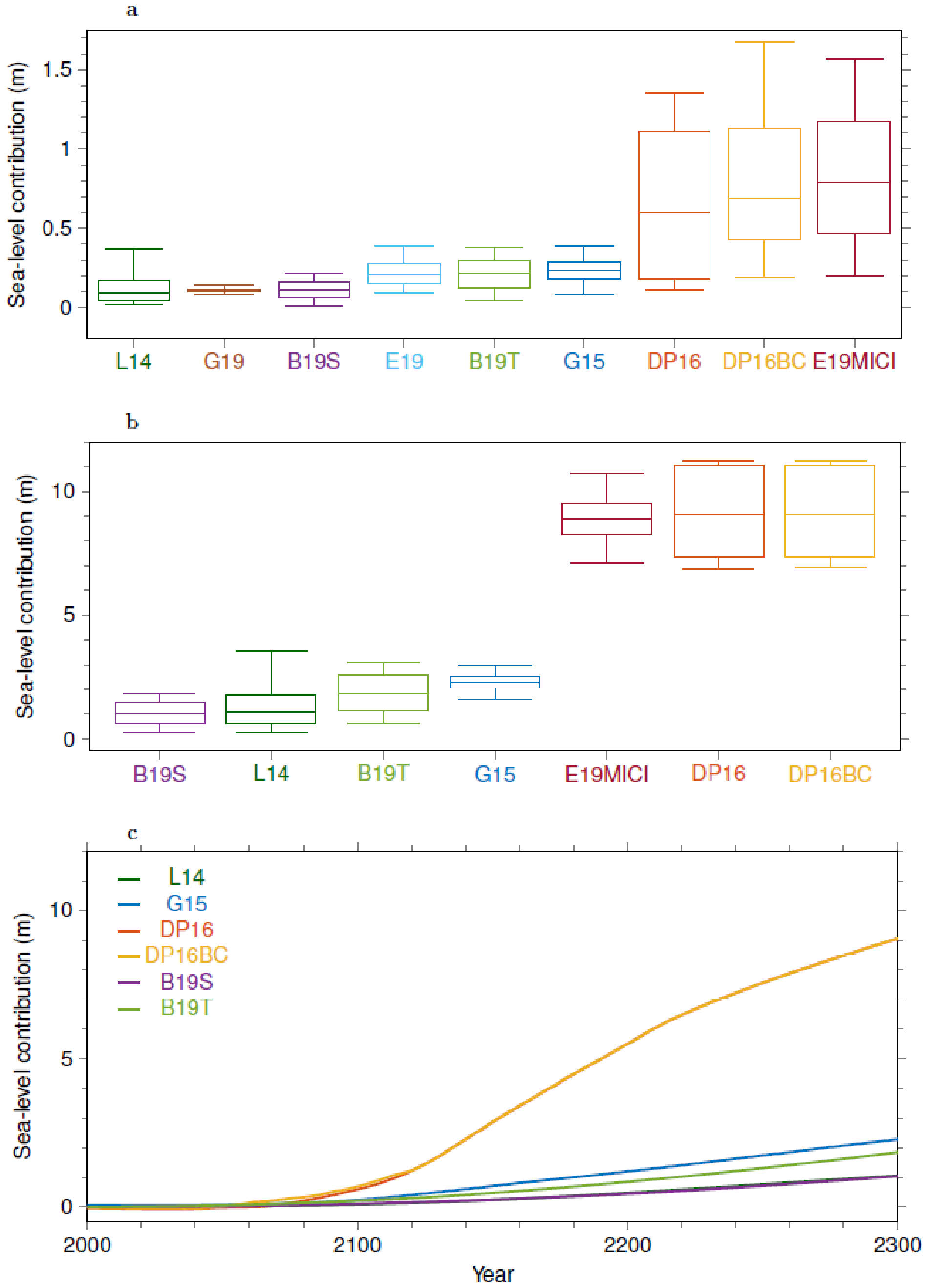


Figure 5. Projections of Antarctic sea-level contribution at (a) 2100 and (b) 2300 under RCP8.5. Boxes and whiskers show the 5th, 25th, 50th, 75th and 95th percentiles. The uncertainty range for Golledge et al. (2015) is based on a Gaussian interpretation for the projections with the 5th percentile given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



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

Reference	**Year**	**Type**	**Time 0**	**Time 1**	**Rate**	**Error**
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