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THE ANTARCTIC ICE SHEET AND SEA LEVEL

Contemporary Changes and Future Projections

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7.1 Introduction

The Antarctic Ice Sheet (AIS) is the largest body of ice on Earth, located in the Southern Hemisphere over the geographic South Pole. Formed over hundreds of thousands of years through the gradual buildup of snow today the AIS has an average thickness of 2.2 km and covers an area of almost 14 million km^2, about 2.75% of the Earth's surface. In total, the AIS contains 30 million km^3 of ice and represents about 62% of the world's total freshwater (Shiklomanov, 1993). If the AIS were to completely melt, global sea levels would rise by about 58 m (Figure 7.1). The AIS is roughly separated by the trans-Antarctic Mountains into two regions, with distinct drainage basins that route grounded ice to the ocean, where each drainage basin has its own ice shelf or ice shelves that are fed by glaciers and ice streams (Figure 7.1). The largest region is the East Antarctic Ice Sheet (EAIS) which covers more than two-thirds of the continent area and contains 52 m of Sea Level Equivalent (SLE). The West Antarctic Ice Sheet (WAIS) has 5.3 m SLE, and the Antarctic Peninsula region has just 0.7 m SLE (Figure 7.1).

The AIS gains mass primarily by snow deposition, and currently loses mass primarily by basal melting and iceberg calving, and to a smaller extent surface melting and sublimation. The buildup of ice in the interior and loss of ice near the coasts causes the ice surface to slope towards its margins. This drives ice flow, which redistributes mass from higher elevation inland to lower elevation at its margins, and regulates how much the AIS contributes to the global sea level. The AIS is surrounded by ice shelves, which form where grounded ice flows into the ocean at the grounding zone and cover nearly 40% of the Antarctic continental shelf seas. The grounding zone is a transition zone between fully grounded and freely floating ice that is typically a few kilometres wide. Ice shelves can impede the flow of

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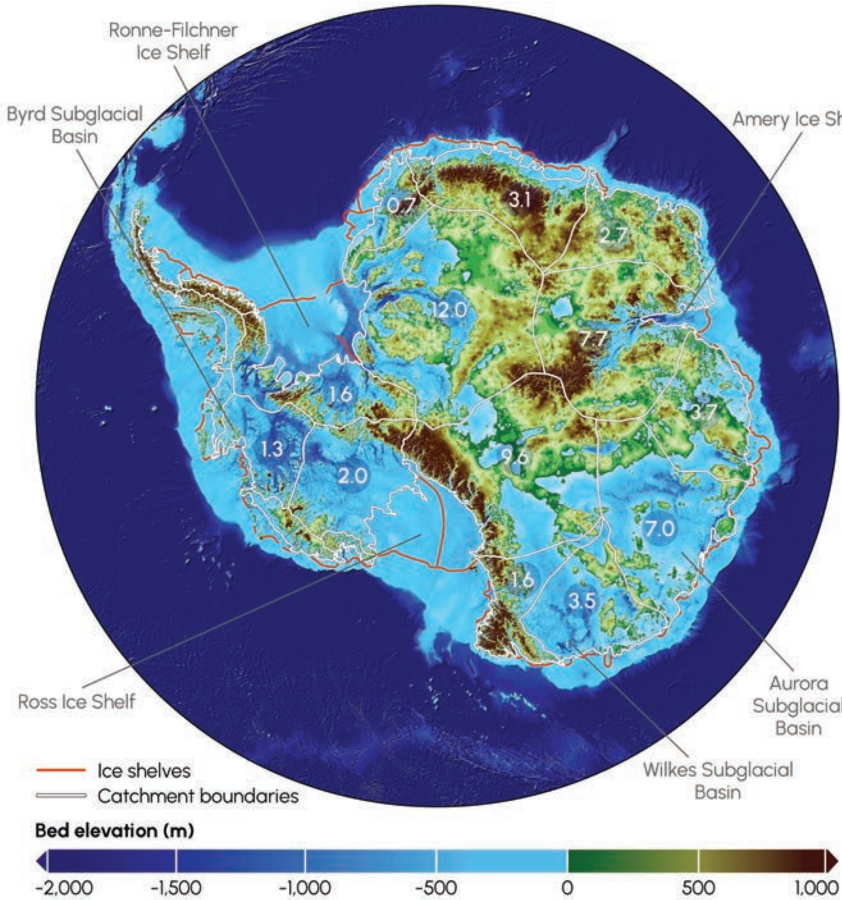

FIGURE 7.1 Antarctic bedrock topography from BedMachine (Morlighem et al., 2020), with drainage basins (Rignot et al., 2011), and estimated global mean sea level potential from each basin, in metres (Tinto et al., 2019).

ice discharge from the grounded ice upstream. When an ice shelf drags against bedrock walls or seafloor highs, resistive forces are transmitted upstream, which reduces driving stress and therefore the flow of ice across the grounding line. This effect is known as "buttressing" (Thomas and Bentley, 1978), and loss or reduction of buttressing can increase discharge of grounded ice to the ocean, causing Sea Level Rise (SLR; e.g. Scambos et al., 2004; Gudmundsson et al., 2019). Ice shelves also influence the surrounding ocean with the freshwater generated through their basal melting being a significant source of cold and freshwater into the Southern Ocean.

In this chapter, we explore several important questions, including:

- *How will Antarctica contribute to sea level in the coming decades to centuries?*
- *What processes and regions should be the focus of future scientific research?*
- *What key processes and regions control uncertainty in projections of future behaviour of the AIS?*

We begin with a review of the processes and feedback that control ice-sheet evolution before providing an overview of the present state and trends of the AIS. We then discuss the deep uncertainty in ice-sheet behaviour, and how this is implicated in future projections of change to the ice sheet. We conclude by examining how sea level will change at a regional scale due to mass loss from the AIS. The locations of the places referred to in the text are shown in Figure 0.1.

7.2 Factors Governing Ice-Sheet Evolution

Understanding the primary controls on the ice sheet and ice shelves and the processes that act to alter them is key to understanding their evolution (Figure 7.2). There are several key interrelated factors (geographical, internal and external) that determine the structure and rheology of ice shelves, and control their mass balance processes. Geographical factors, such as topography (Figure 7.1) and the

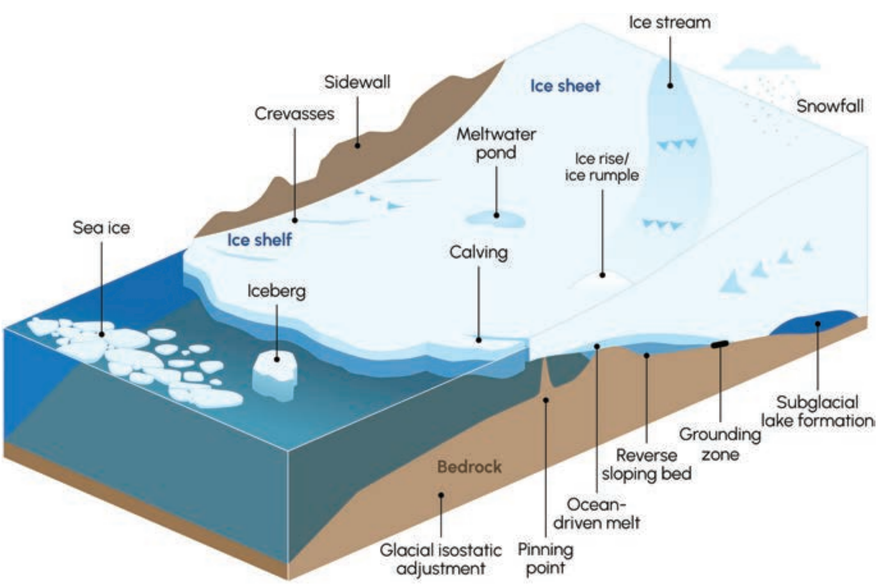

FIGURE 7.2 Schematic of processes affecting the evolution of the Antarctic Ice Sheet.

underlying conditions, set the ice thickness at the grounding line, the ice draft and the sub-ice cavity shape, including the basal and sidewall contact points. Internal factors, such as ice temperature, history and age of ice, ice type (firn vs meteoric vs marine ice), impurities and the degree of damage (crevasses and rifts), set the ice rheology. External factors are the atmospheric and oceanographic settings, which affect both surface and basal processes. These processes also interact with each other, and act on multiple spatial and temporal scales; the response of the ice sheet to future climate states will depend on these interactions and feedbacks. In this section, we describe the main controls on ice-sheet evolution.

7.2.1 Processes

Ice rheology: Ice flow is controlled by its rheology, i.e., how deformable the ice is, which depends on the ice temperature, the orientation of the ice crystals (its fabric) and the presence of impurities in the ice, and by how readily it can fracture. Several factors determine the relative magnitude of the stress applied during ice flow that can feedback on the ice rheology. At the low stresses generated by the small surface slopes in the interior of the ice sheet, ice flows as a polycrystalline solid (Cuffey and Paterson, 2010; Treverrow et al., 2012). Over shorter time scales (measurable in hours) ice can behave as an elastic solid, and the ability to release stored elastic energy can be important for the brittle fracture of ice. The deformation rate (or strain rate) of ice in tertiary creep is sensitive to deviatoric stress (roughly speaking, the shear force per unit area in the ice; Glen, 1955). Consequently, ice velocity more than quadruples if ice thickness or surface slope doubles, and bedrock troughs play a large role in channelising ice flow.

The relationship between strain rate and stress is also dependent on temperature, impurities (e.g., air, sediment or salt; however, their influences are poorly known) and the presence of any meltwater when the temperature is close to the melting point (Paterson and Budd, 1982). In general, warmer and wetter ice deforms more easily. In turn, the temperature of the ice is controlled in part by flow, as faster flow leads to more internal deformation that dissipates more heat. This leads to potentially self-sustaining feedback in which faster flow warms the ice, causing even faster flow. Feedback of this type allows ice to self-organise into patterns of alternating fast- and slow-flowing features (Hindmarsh, 2009), potentially leading to the formation of "ice streams" (Joughin et al., 1999).

Sliding: The high velocities of ice streams, around 100–1000 m per year (e.g., Joughin et al., 1999), are caused by rapid sliding at the interface between ice and bed, or rapid deformation within the underlying bed near that interface (Alley et al., 1986). Both of these processes require ice temperatures at the melting point to provide the highly-pressurised liquid water needed to permit rapid basal motion (see also Bentley et al., 1998, Tulaczyk et al., 2000). Basal ice is often at a temperature close to the pressure melting point (−0.87°C under 1 km of ice), or the ice may even be undergoing active melting. Melting is due to a combination of geothermal heating and frictional heat generated as the ice slides over bedrock and sediments (e.g., Tulaczyk et al., 2000). Subglacial meltwater lubricates the ice to enable sliding, which has resulted in the formation of ice streams (Hughes et al., 1977; Bentley, 1987). An additional self-sustaining feedback can contribute to ice stream formation: warming of the bed due to dissipation of heat by incipient sliding (Hindmarsh, 2009; Mantelli et al., 2019), and the production of meltwater by frictional dissipation once the melting point is reached (e.g. Kyrke-Smith et al., 2014, Schoof and Mantelli, 2021), although the details remain poorly understood (e.g. Mantelli and Schoof, 2019).

Ice shelf buttressing: The degree of buttressing of an ice shelf is set by (a) the lateral drag exerted on it by the sidewalls, where concentrated deformation leads to the formation of distinctive shear margins, and (b) the lateral and basal drag due to "pinning points" (localised topographic highs on the seafloor that come in contact with the base of the ice shelf (Figure 7.2; Goldberg et al., 2009). The amount of drag resulting from contact with sidewalls and pinning points or scales with contact area and friction. Long narrow ice shelves, such as Amery Ice Shelf, have negligible extensional stresses at the grounding line due to the large lateral shear from the margins (Pegler, 2016). However, for most ice shelves, stability is controlled primarily by compressive stresses between key pinning points (Still, 2018).

Although complete ice-shelf removal has the strongest effect on buttressing (Sun et al., 2020), it can also be reduced through ice-shelf thinning (Haseloff and Sergienko, 2018; Gudmundsson et al., 2019), with melting near the grounding zone and sidewalls having a large influence on buttressing and ice flow (Gagliardini et al, 2010; Reese et al, 2015). Buttressing can be lost gradually as the ice is thinned through ocean-driven melting (Gudmundsson et al., 2019). More rapid reductions in buttressing can also occur if overall thinning or retreat via calving is sufficient to weaken or lose the compressive arch, or to cause loss of contact with a pinning point (Still et al., 2019). Buttressing is also impacted by the weakening of shear margins through the formation of cracks (Macgregor et al., 2012) or melting (Alley et al., 2019).

Snowfall and surface melting: Snowfall is the only way the surface of the AIS gains mass. Surface mass budget (SMB) is the net result of mass gain, including precipitation (solid and liquid), and mass loss, including surface melting, evaporation, sublimation and runoff (Lenaerts et al., 2019), and is influenced by atmospheric interactions, such as drifting snow and katabatic winds (Mottram et al., 2021). Snowfall that falls on the AIS compacts under its own weight into firn, which can be O (10–100) m thick. Surface melting is a less significant mass loss process for the AIS than for Greenland, and while it has been linked to ice-shelf collapse on the Antarctic Peninsula (Scambos et al., 2004), it has not yet been detected to occur in large volumes upstream of the grounding zones anywhere in Antarctica. However, 65,000 surface lakes were tallied in 2017, of which 60% were on ice shelves, many located just downstream of grounding zones (Stokes et al., 2019).

Surface melting on Antarctic ice shelves has been projected to double by 2050 (Trusel et al., 2015), which has implications for ice-shelf stability (Warner et al., 2021; Kingslake et al., 2017; Bell et al., 2018). In the longer term, surface meltwater could spread upstream of the grounding lines and ultimately may reach the bed through moulins, leading to faster ice-sheet flow and influencing sub-ice-shelf ocean dynamics, a process that presently occurs in Greenland (Trusel et al., 2018). Recent studies suggest it is not just melt but also the ratio of melt over accumulation that matters to the SMB (Donat-Magnin et al., 2021).

Basal melting and refreezing: Conditions that drive the ocean circulation in the ice-shelf cavity and govern basal melt rates are a complex interplay between the shape of the cavity geometry, basal roughness and many types of processes in the ocean and external forcing (Dinniman et al., 2016; Adusumilli et al., 2020; Rosevear et al., 2024). Deeper ice that is in contact with seawater melts faster than shallow ice due to the pressure dependence on the freezing temperature of seawater (McDougall et al., 2014). Typically the deepest parts of an ice shelf are immediately adjacent to the grounded ice. Melting here steepens the surface of the ice flowing into the ice shelf and leads to a stronger reduction in buttressing than melting elsewhere (Gagliardini et al., 2010) and can drive feedback with ice flow leading to the evolution of basal channels (Dow et al., 2018; Section 2.2).

Generally, more melting occurs at depth which both cools and critically also freshens the ocean, forming ice shelf water that is therefore more buoyant and will ascend along the underside of an ice shelf. Basal melting of ice shelves may also be influenced by subglacial meltwater and associated sedimentation processes (Gwyther et al., 2023), and other oceanographic processes (e.g., tides, eddies, open ocean, sea ice and atmosphere processes). For some ice shelves, typically those with the deepest drafts, the rising meltwater can become supercooled – cooler than the local freezing point temperature of seawater – leading to the formation of marine ice. Marine ice is created by the accretion of frazil – small ice crystals that grow in seawater and can accrete to the ice base and directly refreeze to the ice shelf base (Lewis and Perkin, 1986; Galton-Fenzi, 2012). Some ice shelves that fringe EAIS have significant marine ice that is thought to contribute to ice shelf stability (e.g., 9% by volume for Amery Ice Shelf; Fricker et al., 2001) that can potentially arrest the development of rifts (Khazendar et al., 2009).

Rifting and calving: Although ice flows like a viscous fluid over long-time scales, on shorter timescales, it can also act like a brittle solid and fracture to form crevasses, which can extend vertically and horizontally. Once a crevasse has fully extended through the thickness of the ice, a rift forms, and the horizontal growth of such rifts can lead to large iceberg calving events (Benn et al., 2007). The initiation and propagation of crevasses and rifts thus control the calving behaviour of ice shelves; though this is a normal process in ice-shelf mass loss, it also can influence ice-shelf stability (e.g., Hulbe et al., 1998; Walker and Gardner, 2019; Bassis et al., 2024).

Around half of the overall AIS mass loss occurs through iceberg calving (Greene et al., 2022), which is represented by a spectrum of sizes and time scales from the formation of large, tabular regular icebergs that occurs infrequently, to the production of many small and irregular icebergs that occurs more frequently (Bassis et al., 2024). Most iceberg calving is part of a natural cycle that balances episodic retreat of ice-shelf fronts with gradual advance through ice flow. Rifts and crevasses frequently initiate where stresses in the ice are concentrated: along the margins of ice shelves and near pinning points. Although pinning points generally increase buttressing and decrease the flux of grounded ice, interaction with pinning points has been associated with ice-shelf fracture following ice-shelf thinning. Rifts that propagate from these locations can contribute to the loss of buttressing by weakening the ice shelf even before an iceberg fully detaches (De Rydt et al., 2018). Rifts in Antarctic ice shelves can be filled with marine ice – ice that forms from the ocean beneath ice shelves - or a mixture of snow, sea ice and blocks of ice (mélange) that has been hypothesised to provide different amounts of structural integrity (Hulbe et al., 1998; Khazandar et al., 2009). On some ice shelves, refrozen marine ice can heal rifts from below (Holland et al., 2009) leading to reduced fracturing and increased stability (Craw et al., 2023).

7.2.2 Feedbacks and Instabilities

The processes that affect mass balance are interconnected and can involve mutually-reinforcing feedbacks (Figure 7.2). The possibility of switching from mutually-reinforcing to mutually-suppressing feedbacks leads to the presence of "tipping points" in the ice sheet, where instability and dramatic change can be triggered at a critical threshold. For a cold, ocean-terminating ice sheet like the AIS, retreat could be significantly influenced by the self-reinforcing feedback processes known as the marine ice sheet instability (MISI) (Weertman, 1974; Thomas and Bentley, 1978) and the marine ice cliff instability (MICI) (Bassis and Walker, 2012; Crawford et al., 2021; DeConto and Pollard, 2016; Pollard et al., 2015) (Figure 7.3). Both are characterised by the rate of mass loss increasing as the depth to the sea floor at the grounding line increases. This depth will progressively increase if the grounding line retreats across a reverse-sloping bed – also known as a retrograde bed – (Figure 7.3), leading to accelerating mass loss. The creep-flow-related mechanisms underlying MISI are relatively well understood. By contrast, the MICI conjecture relies on thresholds and rates that depend on processes controlling the fracture of ice, which are less well understood.

In ice-sheet models, the flow-geometry coupling can reverse MISI and stabilise the grounding line on a retrograde slope (Gudmundsson et al., 2012). This stabilising effect hinges on there being a sufficiently long and narrow ice shelf with limited mass loss from basal melting. In the absence of calving, a retreat of the grounding line can then occur without a comparable retreat in the position of the calving front (the ice face at a glacier's terminus), leading to a longer shelf with a stronger buttressing effect (Schoof et al., 2017; Haseloff and Sergienko, 2018). Gomez et al. (2010a) suggested isostatic adjustment and changes in the geoid associated with grounding line retreat can further mitigate the onset of MISI, even on retrograde slopes. These nuances show that MISI – when it occurs – is not controlled solely by the local geometry of the ice-sheet bed but involves multiple feedbacks and processes that can mitigate runaway retreat.

MISI: Although initially controversial, it has since been shown that the simplified case of ice flow on a retrograde bed is always unstable (Weertman, 1974; Thomas and Bentley, 1978; Schoof, 2007a) until other factors, such as buttressing and changes in thermal properties, can complicate the potential feedback. This is because flux through the grounding line increases with the deviatoric stress and ice thickness (Schoof, 2007b; Figure 7.3a). In the absence of buttressing, this results in MISI, as stress at the grounding line, then also increases with ice thickness, and therefore (the ice being just afloat) with depth to the sea floor. In practice, ice shelves do exert buttressing and the geometry of the bed varies laterally and

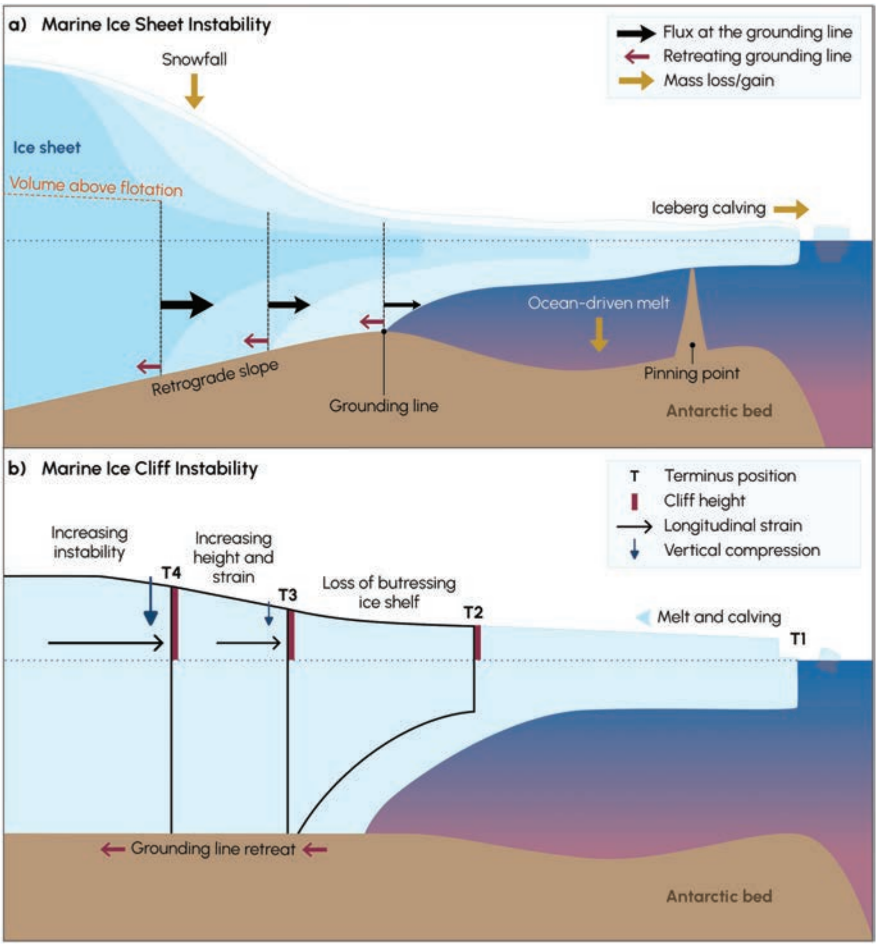

FIGURE 7.3 Feedbacks driving ice sheet evolution: (a) Marine Ice-Sheet Instability (MISI). (b) Marine Ice Cliff Instability (MICI).

temporally, and the strength of buttressing is not simply dictated by external forcing but is coupled to the migration of the grounding zone.

MICI: If a grounded calving face is exposed after the loss of an ice shelf, ice-cliff failure could initiate if the calving face extends past a threshold height above the sea surface (Figure 7.3b). That threshold corresponds to a point at which the calving face is not able to withstand stresses generated by the weight of the ice (Bassis and Walker, 2012). At its simplest, a stress threshold might be expected to cause an irreversible cascade of calving events once initiated: as soon one piece of ice has been removed by calving, thicker ice upstream is exposed producing a sequence of ever taller cliffs that results in a runaway collapse, known as MICI. Calving does not need to be instantaneous and a self-sustaining retreat will only occur if newly-exposed calving faces are not drawn below the threshold height by dynamic thinning caused by horizontal stretching (Bassis et al., 2021; Crawford et al., 2021).

There are no direct observations of MICI; therefore, there is large uncertainty in the threshold cliff height after which ice-cliff failure will initiate, as well as the ensuing retreat rates. Existing attempts to quantify ice-cliff retreat rates for the purpose of predicting future ice-sheet evolution have either used parameterisations based on limited data (DeConto and Pollard, 2016) or calibration against synthetic results (Crawford et al., 2021). Recent studies are beginning to compare rates of retreat of calving cliffs that are close to the theoretical limit (e.g., Needell and Holschuh, 2023) to provide more empirical constraints on cliff failure.

7.3 Present State and Trends

Satellite estimates show that the AIS has been losing mass since the 1990s, and this loss is accelerating with time (Otosaka et al., 2023). Mass loss has been dominated by changes in WAIS (see Figure 7.4) (Smith et al., 2020). The latest study (Otosaka et al., 2023) has shown that for the period 1992–2020, ice loss from WAIS and the Antarctic Peninsula were 82±9 Giga-tonnes per year (Gt yr^{-1}) and 13±5 Gt yr^{-1}, while EAIS had a small gain of 3±15 Gt yr^{-1}. All recent studies agree on the trend of ice loss from WAIS and the Antarctic Peninsula, and on the rate of loss having increased since around 2006 (Rignot et al., 2019; Otosaka et al., 2023).

West Antarctica: Most research on processes that contribute to AIS mass loss has focused on WAIS (e.g., Rignot et al., 2014; Joughin et al., 2014). Satellite laser altimetry over grounded and floating ice (Smith et al., 2020) has shown that between 2003 and 2019, WAIS lost 76 ±49 Gt y^{-1} of floating ice and −169 ± 10 Gt^{-1} of grounded ice (7.5 mm SLE). Basins that drain into the Amundsen Sea Embayment (Figure 7.1) have experienced dramatic changes (Figure 7.4a); here, the ice shelves are in contact with warm ocean waters and have experienced enhanced melting and thinning (Adusumilli et al., 2020; Smith et al., 2020) and grounding line retreat (Khazendar et al., 2015). Ice-shelf thinning and grounding-line retreat have led to reduced buttressing on the upstream grounded ice, and the dynamic thinning has spread inland (Gudmundsson et al., 2019; Section 2). Accelerated ice flows of glaciers, such as Thwaites and Pine Island into the ocean (Smith et al., 2020; Otasaka et al., 2022), are contributing to SLR. Because this drainage basin is vulnerable to MISI (Figures 7.1 and 7.3a), the onset of instability remains difficult to confidently determine at this time. Changes in processes have been triggered by this thinning: e.g., the floating portion of Pine Island Glacier has transitioned from a previously quasi-stable cycle of advance and retreat at its calving front to a calving regime characterised by more frequent detachment of tabular icebergs and calving front retreat (Jeong et al., 2016). Similarly, portions of the Thwaites Glacier ice tongue underwent an abrupt change from a previously intact ice shelf into fragmented remnants (Miles et al., 2020). Although ocean forcing has been implicated in these transitions, the changes to the Pine Island and Thwaites ice shelves were not predicted by models used to simulate ice-sheet and ice-shelf processes. This highlights that the processes leading to shelf fracture and disintegration remain poorly understood and must be a priority for future research.

East Antarctica: EAIS is estimated to be close to balance or even slightly positive in its mass change (Martín-Español et al., 2016; Smith et al., 2020; Otosaka et al., 2023). This is because mass losses from the ocean-driven melt are compensated by increased snowfall over its large area (Smith et al., 2020; Otosaka et al., 2023; Figure 7.4a). Rignot et al. (2019) estimated an EAIS SLR contribution of 4.4 ± 0.9 mm from 1979–2009. Satellite laser altimetry for 2003–2019 (Smith et al., 2020) has shown EAIS gained a total of 90 ± 21 Gt y^{-1} of floating ice and 106 ±29 of grounded ice (−4.0 mm SLE). There are regional differences in behaviour within EAIS. Increases in snowfall are concentrated in Dronning Maud Land (Velicogna et al., 2014; Smith et al., 2020), while multiple studies measure ongoing mass loss from Wilkes Land since the mid-2000s (Velicogna et al., 2014; Smith et al., 2020). Wilkes Land is susceptible to MISI and glaciers there are showing signs of change (Li and Dawson et al., 2023).

Since satellite observations began there have been major calving events from several ice shelves, but the record is short compared to the calving cycles. However, for ice shelves with a more rapid calving cycle, recent observations point to the influence of ocean and atmospheric forcing on increasing calving rates, as is evident from retreat and terminus change at Pine Island (Bradley et al., 2022) and Thwaites Glacier (Seroussi et al., 2017). The influence of increasing atmospheric temperatures on calving, ice-shelf stability and retreat has been particularly dramatic in the Antarctic Peninsula, e.g., the 2002 collapse of the Larsen B Ice Shelf. Historic radar altimetry (which extends only to 72°S) suggests that changes in the Antarctic Peninsula began before 1992 (Fricker and Padman, 2012).

Over the past five years (2019–2023), ICESat-2 observations have detected a change in the pattern and rate of snow accumulation (Fricker et al., 2020; Adusumilli et al., 2023). WAIS gained accumulation via several short-period extreme events, driven by moisture-laden atmospheric rivers, contributing 41% of the increases in height in 2019 (Adusumilli et al., 2023). The subsequent years, especially 2020 and

2021, displayed ongoing ice losses around the margins and gains in the interior. In 2022, there was a large amount of snow accumulation and the surface mass balance anomaly reached +325 Gt (net mass gain of 290 Gt; Adusumilli et al., 2023). Spatial patterns indicated increased elevations, particularly over EAIS in Wilkes Land.

7.4 Future Projections and Deep Uncertainty

Projections of the future SLR contribution from the AIS are highly uncertain (Figure 7.5). Projections up to 2100 show that the AIS represents a relatively small fraction of SLR contribution, in the absence of rapid ice-sheet collapse, but from 2100 to 2300 has the potential to be the dominant source of SLR. Recent projections for ice-sheet evolution indicate a potential sharp acceleration in SLR towards the end of this century (Edwards et al., 2021), with high-end scenarios indicating the small possibility that global SLR could exceed 1.5 m by 2100 (Figure 7.5). This led to a new high-end risk scenario being introduced with IPCC (2021) stating that a global rise "approaching 2 m by 2100 and 5 m by 2150 under a high greenhouse gas emissions scenario cannot be ruled out due to deep uncertainty in ice-sheet processes". Deep uncertainty occurs when experts and/or decision makers do not know or cannot agree upon the system model relating actions to consequences or the prior probabilities on key parameters of the system model (Lempert and Collins, 2007). The large range of SLR projections confuses many efforts for coastal adaptation on century time scales (Dietz et al., 2022; Hirschfeld et al., 2023).

Large uncertainties in projections come from deficiencies in: (i) understanding of processes and feedbacks that influence ice-sheet evolution (Seroussi et al., 2024); (ii) knowledge of ice-sheet and near-ocean geometry and conditions (e.g., how the shape and evolution of the bed and presence of subglacial water and ocean melting can influence flow); and (iii) future climate model projections used as input to these simulations of ice-sheet evolution (Li et al., 2023).

Future ice-sheet behaviour subject to tipping points depends sensitively on initial conditions and climate forcing. This means that, even when the processes that control thresholds and tipping points are well understood, small changes in climate forcing or initial conditions can lead to large differences in projected SLR. This was demonstrated by Robel et al. (2019) for MISI, where simulations using the same ice-sheet model subject to intrinsic variability in climate forcing resulted in a large distribution of grounding line retreat rates and associated SLR. For processes like ice shelf collapse and MICI, which depend on poorly-quantified thresholds, the uncertainty in outcomes is even larger as small changes in under-observed parameters can lead to diverging results. Therefore, SLR projections are often treated probabilistically with the various thresholds and instabilities leading to substantial probabilities for more extreme outcomes in a "negatively skewed" distribution (Robel et al., 2019), providing what are actually unknown-likelihood, but potentially high-impact events (Figure 7.5).

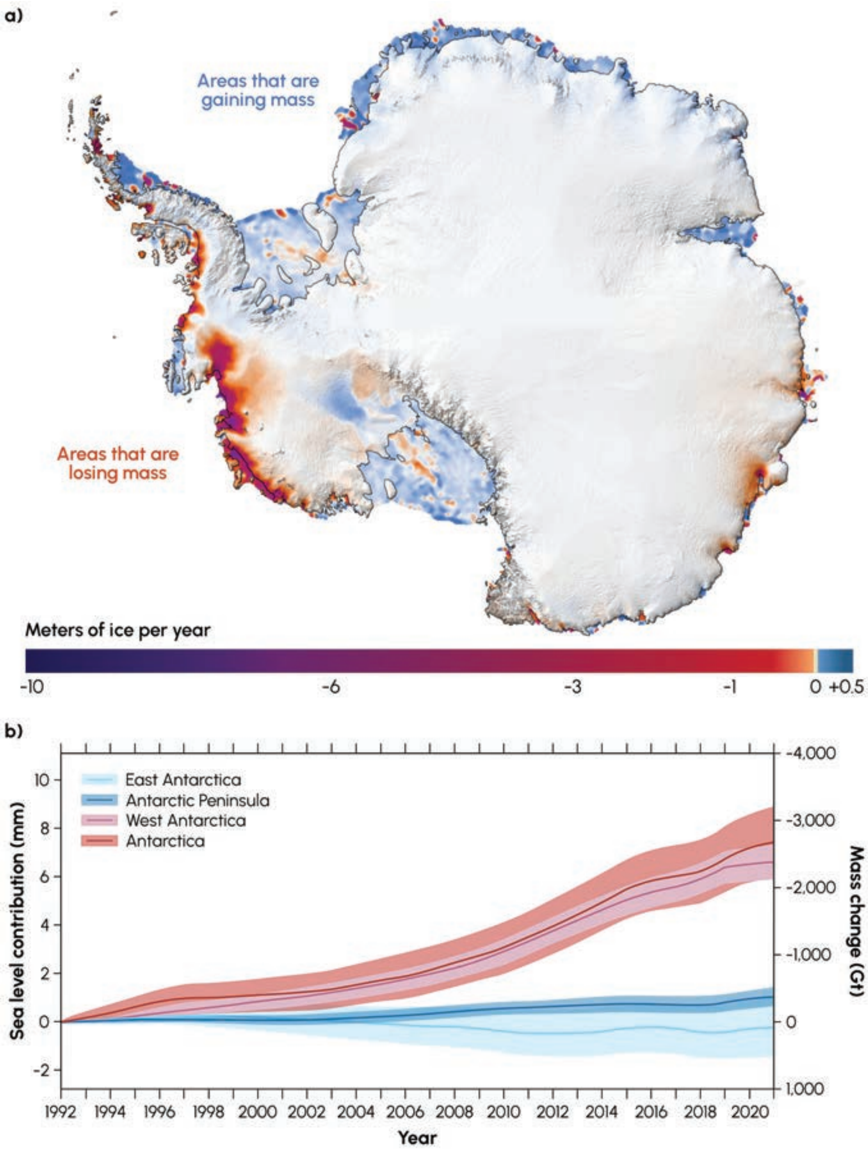

FIGURE 7.4 Antarctic mass loss: (a) Ice-sheet thinning for grounded and floating ice between 2003 and 2019 estimated by differencing elevations from NASA's Ice, Cloud and land Elevation Satellite (ICESat) and ICESat-2 laser altimeters (Smith et al., 2020). (b) Cumulative Antarctic mass loss compiled from 24 separate studies by the IMBIE team (adapted from Otosaka et al., 2023); shading represents the associated uncertainties.

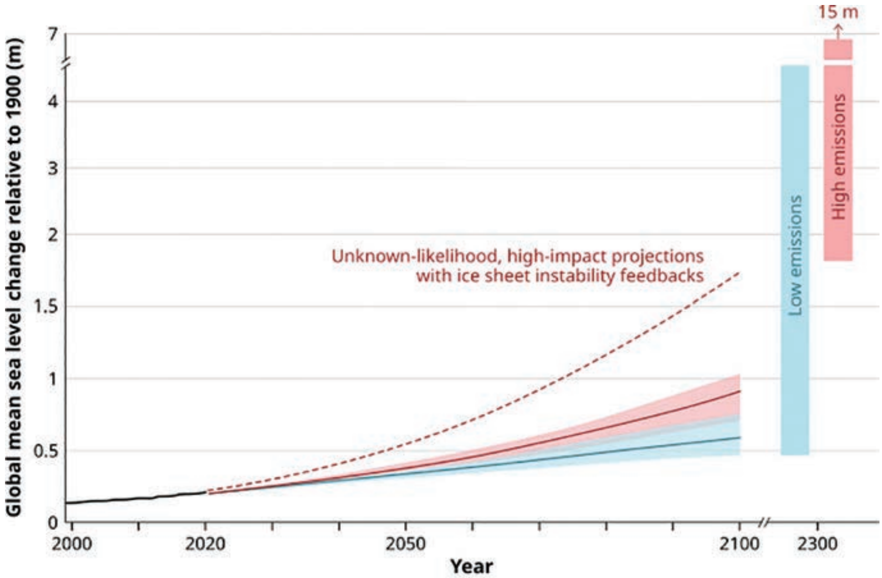

FIGURE 7.5 Global SLR IPCC AR6 climate projections from 1990 to 2100 and 2300 (metres relative to 1950; adapted from IPCC, 2021).

Many processes occur in complex areas on spatial scales that are too small to be resolved by current satellite data; for example, observing ice-shelf rifts (Walker et al., 2021), grounding zones (Freer et al., 2023) and melt rates in basal channels require high-resolution observations of ice-shelf height (Alley et al., 2016) and flexure (Rignot et al., 2024). One of the largest uncertainties remains the open question of how to appropriately represent the various processes through which calving occurs above and below the height threshold at which ice-cliff failure might initiate (MICI).

There is significant variability in atmospheric and oceanic climate forcing used to drive models, and potential feedbacks remain poorly quantified (e.g., Hanna et al., 2024). For instance, since ice shelves play a buttressing role, ice-sheet behaviour is prone to intrinsic variability driven by ocean melting of the ice shelves that can produce a pseudo-steady state, advancing or retreating behaviour (e.g., Gwyther et al., 2018; McCormack et al, 2021). The ISMIP6 project came to little consensus on the role of emission scenarios in driving ice-sheet change because increased snowfall, particularly over EAIS, can significantly offset some of the expected increase in discharge. There is similar significant uncertainty in ocean forcing (Seroussi et al., 2020). The influence of these external processes on the projected uncertainties seems to be as important as the influences of approximations in model physics and simulated ice evolution (Seroussi et al., 2024).

7.5 Regional Sea Level Changes

AIS melting is expected to dominate global mean SLR in the coming centuries, but site-specific sea level changes are very heterogeneous, and can be dominated by regional and local processes such as subsidence and coastal erosion in some areas. Local SLR (in addition to other factors such as changing storm frequency and strength) in turn can intensify the impact of extreme sea level and flooding events (IPCC, 2021). Here, we focus on characterising the spatially variable sea level changes associated with AIS evolution. Changes in the distribution of ice cover of the AIS cause spatially variable sea level changes that are often much higher or lower than global average sea level. The global pattern of sea level changes associated with AIS loss is primarily due to the effects of gravity, changes in Earth rotation and viscoelastic deformation of the solid Earth, often referred to as "gravitational, rotational and deformational (GRD) effects" (Farrell et al., 1976; Mitrovica et al., 2011; see Figure 7.6). The addition of meltwater to the oceans can also contribute to more localised spatial variability in sea level changes (Golledge et al., 2019).

Ice mass loss on timescales of years to centuries leads to a local drawdown of the geoid – the gravitational equipotential corresponding to the sea surface – and uplift of the solid Earth, which, combined, lead to a relative sea level fall within roughly 2000 km of the region of ice loss (Woodward, 1888; Mitrovica et al., 2011). This sea level fall can be an order of magnitude or more larger than the global average SLR. Gravitational effects also lead to a greater than average rise at greater distances from the melting ice sheet. Crustal uplift beneath oceanic areas freed of marine-based ice expels water out of the vicinity of the melting ice sheet further amplifying far-field SLR, an effect termed "water expulsion" (Gomez et al., 2010b). Water loading of the global oceans also deforms the solid Earth in the far-field, adding to the spatial variability of the sea level change associated with the ice loss. Finally, the ice mass loss drives a shift in the Earth's rotation axis towards the region of ice loss, and this, in turn, redistributes water in the oceans, acting to decrease relative sea level in quadrants of the Earth's surface that approach the rotation axis and increase relative sea level in the other quadrants (Milne and Mitrovica, 1998).

Solid Earth deformation due to surface ice and ocean loading changes is in general viscoelastic, with a largely elastic response on short, years to centuries, timescales and viscous flow of the Earth mantle towards isostatic equilibrium on longer timescales (Peltier, 1974). Viscous deformation depends on the thickness of the lithosphere and viscosity of the Earth's mantle. Viscosities are variable across relatively short spatial scales and much lower in some areas (e.g., Barletta et al., 2018; Lloyd et al., 2020), leading to viscous uplift in response to ice unloading occurring on timescales of years to decades in some areas rather than the more typical millennial and longer timescales. Substantial uncertainty remains in the structure and associated response of the solid Earth in Antarctica, with implications for ice mass loss estimates (IPCC, 2021).

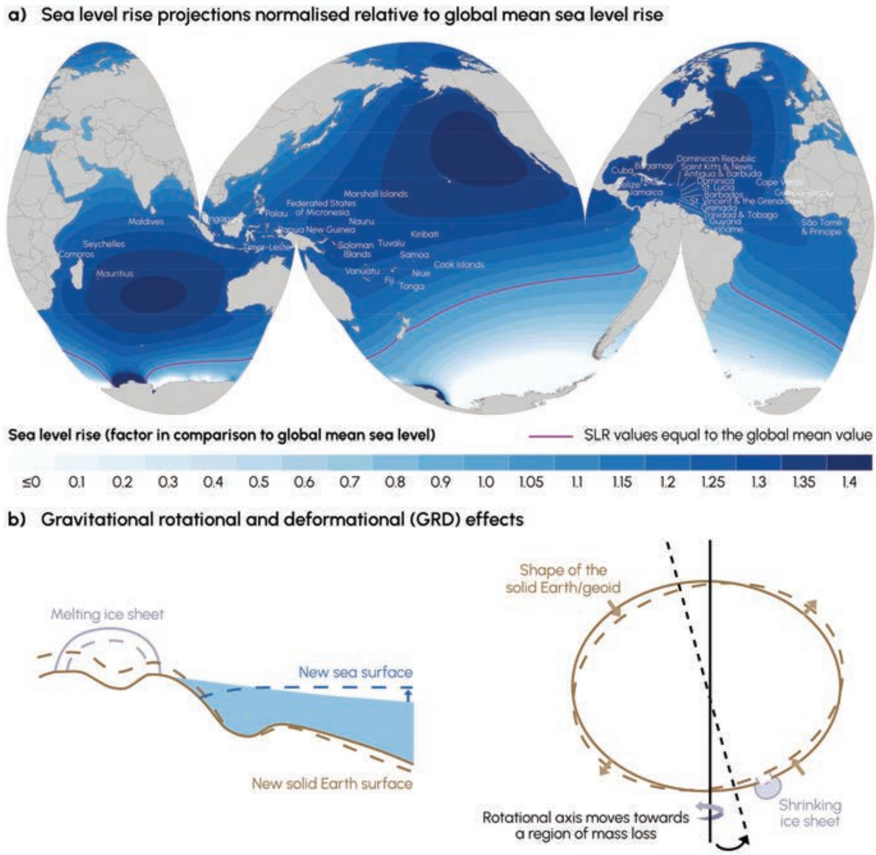

FIGURE 7.6 Antarctic Ice Sheet deglaciation impacts on regional sea level changes: (a) Pattern of sea level factor at 2100 under RCP4.5 – the mid-level future emissions estimates (adapted from Sadai et al., 2022): e.g., a factor of 1.2 indicates SLR 1.2 times the global mean. (b) Schematics of the physical effects causing the spatial variability in sea level changes in (a): left – gravitational effects on the sea surface and deformation of the solid Earth; right – the effects of Earth rotational changes (adapted from Whitehouse et al., 2018).

Global sea level patterns due to recent and future ice loss: Sea level changes due to ice loss from the AIS combine with the effects of Greenland ice loss, resulting in up to about 30% greater than average SLR in mid- to low-latitude regions by the end of the century (Golledge et al., 2019; Figure 7.5). However, much greater local SLR amplification can occur in some areas on decadal timescales when more localised patterns of ice cover changes drive constructive interference between GRD effects (Roffman et al., 2023). The global pattern of SLR away from the ice sheets is significantly more sensitive to the geometry of Antarctic ice loss than to that of Greenland ice, loss due to its location on the rotation axis. Uncertainty in projections of AIS ice loss (Seroussiet al., 2020) in turn leads to uncertainty about the projected regional patterns of sea level changes (Roffman et al., 2023).

WAIS ice loss produces SLR peaks greater than 30% higher than the global average along North American coastlines and in the Indian Ocean (e.g., Gomez et al., 2010b; see Figure 7.6a), while SLR peaks due to EAIS melting are shifted relative to the WAIS case to the North Pacific and South Atlantic Oceans, mainly due to rotational effects. Some coastal areas will experience larger impacts of SLR at lower levels of warming than other areas, highlighting the climate injustice implications of global mean temperature targets (Sadai et al., 2022). For example, Small Island Nations, which are already experiencing SLR impacts, are expected to see a greater than average rise associated with AIS loss in the future regardless of the ice-sheet model projection and level of future warming (Roffman et al., 2023; see Figure 7.6). Multiple decades of far-field sea level measurements are required to detect the contribution of recent ice loss to global sea level above natural variability and ocean dynamic effects (Kopp et al., 2010), and have only recently begun to be detected in sea surface altimetry and tide gauge records (e.g., Moreira et al., 2021). Earlier detection is possible in the near-field of the ice sheets where the signal is larger and detection will improve with longer and more near-field records as ice mass loss accelerates.

7.6 Summary and Future Research Directions

The AIS is the largest body of ice on Earth, and it will continue to have a profound influence on sea level and Earth's climate. We have provided a review of how we expect Antarctica to contribute to sea level in the coming decades and beyond, what we understand from trends in behaviour since the 1990s and what we currently think to be the most important processes and behaviours that should continue to be the focus of future scientific research to best constrain uncertainty in projections of the future behaviour of the AIS.

We set out to explore the questions posed in the introduction and here, in summary, we present the following approaches that we have identified as necessary to ensure progress:

Processes: Integration between new observations, simulations and laboratory experiments is needed to evaluate and constrain models of key processes, including buttressing, sliding dynamics, subglacial water behaviour, basal melting and freezing of ice shelves, ice rheology and the influences of impurities, the formation and evolution of features such as ice shelf rifts and basal channels, surface mass budget processes, solid Earth processes and processes that govern the far-field impacts of SLR. Work is also needed to understand processes driving both temporal variability (Hanna et al., 2024), and AIS tipping-points, including those that cause shelf fracture and disintegration. (Winkelmann et al., 2023)

Modelling: Planning requires credible projections based on physical ice-sheet and climate models. The most significant uncertainties in projections of ice-sheet evolution are associated with climate forcing and the fidelity of present ice-sheet models to represent key physical processes. Uncertainty estimates can only be reasonably achieved by examining large suites of possible future outcomes. Continued development of state-of-the-art models is therefore critical to combine advanced process understanding with observations. This will lead to useful simulations that can guide and aid interpretation of measurements and provide future projections with substantially constrained uncertainties.

Observations: Targeted high-resolution observations of the AIS and surrounding oceans, sea ice, atmosphere (climate) and solid Earth are critical to refine the understanding of processes for modelling, fill in gaps in maps of bedrock and ice-sheet shape and assess the state of the ice sheet and cavities beneath the ice shelves. Long-term monitoring of key locations of the AIS and bedrock and sea level in the Southern Ocean are critical to provide enhanced confidence in present-day trends and associated contributing processes. The EAIS, specifically the Wilkes and Aurora subglacial basins, must be a focus of future activity.

Coordinated science: Concerted system-scale approaches are needed to link together the different components that are often studied individually, given the critical feedbacks and interconnected processes we have identified. Sustained and coordinated effort is key to make the best use of resourcing with multi-national collaborations to pool expensive logistics resources. The pathway to improvements needs complementary use of field, satellite, laboratory and simulations, which is presently underutilised (Cook et al., 2022, Gwyther, 2018). The approach can be used to prioritise research focus needed to make progress on constraining future contributions from the AIS to SLR within the next decade.

The substantial uncertainty in the AIS contribution to future SLR, especially the deep uncertainty under high-end warming, presents a delicate challenge for coastal planning efforts (Kopp et al., 2023; van de Wal, 2022), which use widely varying SLR projections in adaptation decisions (Hirschfeld et al., 2023). For many applications, the deep uncertainty is dominated by the social environment when decision makers and stakeholders do not agree on the likelihood and magnitude of future scenarios. Scientific uncertainties are typically much smaller than the uncertainties associated with socio-economics and appropriate decision making frameworks. Constraining the deep uncertainty to enable adaptive planning and decision making must therefore involve all stakeholders – funding bodies, scientists, planners and policy-makers – to ensure ongoing advances in ice sheet and sea level science are made for society to be able to adapt and react to future change.

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References

Adusumilli, S., et al. (2023). Ice-sheet mass balance in state of the climate in 2022. *Bulletin of the American Meteorological Society*, 104(9): S339–S341.

Adusumilli, S., et al. (2020). Interannual variations in meltwater input to the Southern Ocean from Antarctic ice shelves. *Nature Geoscience*, 13(9): 616–620.

Alley, K.E., et al. (2016). Impacts of warm water on Antarctic ice shelf stability through basal channel formation. *Nature Geoscience*, 9(4): 290–293.

Alley, K.E., et al. (2019). Troughs developed in ice-stream shear margins precondition ice shelves for ocean-driven breakup. *Science Advances*, 5(10): 2215.

Alley, R.B., et al. (1986). Deformation of till beneath ice stream B, West Antarctica. *Nature*, 322(6074): 57–59.

Barletta, V.R. et al. (2018). Observed rapid bedrock uplift in Amundsen Sea embayment promotes ice-sheet stability. *Science*, 360(6395): 1335–1339.

Bassis, J.N., et al. (2021). Transition to marine ice cliff instability controlled by ice thickness gradients and velocity. *Science*, 372(6548): 1342–1344.

Bassis, J.N., et al. (2024). Stability of ice shelves and ice cliffs in a changing climate. *Annual Review of Earth and Planetary Sciences*, 52: 1.

Bassis, J.N. and Walker, C.C. (2012). Upper and lower limits on the stability of calving glaciers from the yield strength envelope of ice. *Proceedings of the Royal Society A*, 468(2140): 913–931.

Bell, R.E., et al. (2018). Antarctic surface hydrology and impacts on ice-sheet mass balance. *Nature Climate Change*, 8(12): 1044–1052.

Benn, D.I., et al. (2007). Calving processes and the dynamics of calving glaciers. *Earth-Science Reviews*, 82(3): 143–179.

Bentley, C.R. (1987). Antarctic ice streams: A review. *Journal of Geophysical Research: Solid Earth*, 92(B9): 8843–8858.

Bentley, C.R., et al. (1998). Radar reflections reveal a wet bed beneath stagnant Ice Stream C and a frozen bed beneath ridge BC, West Antarctica. *Journal of Glaciology*, 44(146): 149–156.

Bradley, A.T., et al. (2022). The influence of Pine Island Ice Shelf calving on basal melting. *Journal of Geophysical Research: Oceans*, 127(9): p.e2022JC018621. 10.1029/2022JC018621.

Cook, S., et al. (2022). Data initiatives for ocean-driven melt of Antarctic ice shelves. *Annals of Glaciology*, 63(87–89): 27–32.

Craw, L., et al. (2023). Modelling the influence of marine ice on the dynamics of an idealised ice shelf. *Journal of Glaciology*, 69(274): 342–352.

Crawford, A.J., et al. (2021). Marine ice-cliff instability modeling shows mixed-mode ice-cliff failure and yields calving rate parameterization. *Nature Communications*, 12(1): 2701.

Cuffey, K.M. and Paterson, WSB. (2010). *The Physics of Glaciers*. Academic Press.

DeConto, R.M, and Pollard, D. (2016). Contribution of Antarctica to past and future sea-level rise. *Nature*, 531(7596): 591–597.

De Rydt, J., et al. (2018). Recent rift formation and impact on the structural integrity of the Brunt Ice Shelf, East Antarctica. *The Cryosphere*, 12(2): 505–520.

Dietz, S., and Koninx, F. (2022). Economic impacts of melting of the Antarctic Ice Sheet. *Nature Communications,* 13: 5819.

Dinniman, M.S., et al. (2016). Modeling ice shelf/ocean interaction in Antarctica: A review. *Oceanography,* 29(4): 144–153.

Donat-Magnin, M., et al. (2021). Future surface mass balance and surface melt in the Amundsen sector of the West Antarctic Ice Sheet. *The Cryosphere*, 15: 571–593.

Dow, C.F., et al. (2018). Basal channels drive active surface hydrology and transverse ice shelf fracture. *Science Advances*, 4(6).

Edwards, T.L., et al. (2021). Projected land ice contributions to twenty-first-century sea level rise. *Nature*, 593 (7857): 74–82.

Farrell, W.E. and Clark, J.A. On postglacial sea level. *Geophysical Journal of the Royal Astronomical Society*. 46(3): 647–667.

Freer, B.I.D., et al. (2023). Modes of Antarctic tidal grounding line migration revealed by Ice, Cloud, and land Elevation Satellite-2 (ICESat-2) laser altimetry. *The Cryosphere*, 17: 4079–4101.

Fricker, H.A., and Padman, L. (2012). Thirty years of elevation change on Antarctic Peninsula ice shelves from multimission satellite radar altimetry. *Journal of Geophysical Research: Oceans*, 117(C2). 10.1126/sciadv.aao721.

Fricker, H.A, et al. (2001). Distribution of marine ice beneath the Amery Ice Shelf. *Geophysical Research Letters*, 28(11): 2241–2244.

Gagliardini, O. et al. (2010). Coupling of ice-shelf melting and buttressing is a key process in ice-sheets dynamics. *Geophysical Research Letters*, 37(14). 10.1029/2010GL043334.

Galton-Fenzi, B.K., et al. (2012). Modeling the basal melting and marine ice accretion of the Amery Ice Shelf. *Journal of Geophysical Research: Oceans*, 117(C9). 10.1029/2012JC008214.

Glen, J.W. (1955). The creep of polycrystalline ice. *Proceedings of the Royal Society of London A*, 228(1175): 519–538.

Goldberg, D.N., et al. (2012). Investigation of land ice-ocean interaction with a fully coupled ice-ocean model: 2. Sensitivity to external forcings. *Journal of Geophysical Research: Earth Surface*, 117(F2). 10.1029/2011JF002247.

Golledge, N. et al. (2019). Global environmental consequences of twenty-first century ice-sheet melt. *Nature,* 566: 65–72

Gomez, N., et al. (2010a). Sea level as a stabilizing factor for marine-ice-sheet grounding lines. *Nature Geoscience*, 3(12): 850–853.

Gomez, N. et al. (2010b). A new projection of sea level change in response to collapse of marine sectors of the Antarctic Ice Sheet. *Geophysical Journal International*, 180: 623–634.

Greene, C.A., et al. (2022). Antarctic calving loss rivals ice-shelf thinning. *Nature*, 609(7929): 948–953.

Gudmundsson, G.H., et al. (2012). The stability of grounding lines on retrograde slopes. *The Cryosphere*, 6(6): 1497–1505.

Gudmundsson, G.H., et al. (2019). Instantaneous Antarctic ice sheet mass loss driven by thinning ice shelves. *Geophysical Research Letters*, 46(23): 13903–13909.

Gwyther, D. (2018). Towards truly integrated modeling and observing of marine ice sheets. *Advances in Polar Science*, 29(4): 231–232.

Gwyther, D.E., et al. (2018). Intrinsic processes drive variability in basal melting of the Totten Glacier Ice Shelf. *Nature Communications*, 9(1): 3141.

Gwyther, D.E., et al. (2023). Subglacial freshwater drainage increases simulated basal melt of the Totten Ice Shelf. *Geophysical Research Letters*, 50(12). 10.1029/2023GL103765.

Hanna, E., et al. (2024). Short- and long-term variability of the Antarctic and Greenland ice sheets. *Nature Reviews: Earth & Environment*, 5: 193–210.

Haseloff, M. and Sergienko, O. (2018). The effect of buttressing on grounding line dynamics. *Journal of Glaciology*, 64(245): 417–431.

Hindmarsh, R.C. (2009). Consistent generation of ice-streams via thermo-viscous instabilities modulated by membrane stresses. *Geophysical Research Letters,* 36(6). 10.1029/2008GL036877

Hirschfeld, D., et al. (2023). Global survey shows planners use widely varying sea-level rise projections for coastal adaptation. *Communications Earth & Environment,* 4: 102.

Holland, P.R., et al. (2009). Marine ice in Larsen Ice Shelf. *Geophysical Research Letters*, 36: L11604.

Hughes, T. (1977). West Antarctic ice streams. *Reviews of Geophysics*, 15(1): 1–46.

Hulbe, C.L. et al. (1998). Comparison of ice-shelf creep flow simulations with ice-front motion of Filchner-Ronne Ice Shelf, Antarctica, detected by SAR interferometry. *Annals of Glaciology,* 27: 182–186.

IPCC (2021). Climate change 2021: The physical science basis. In *Contribution of Working Group I to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change* [Masson-Delmotte, VP. et al (eds.)]. Cambridge University Press, Cambridge and New York, pp. 2391.

Jeong, S., et al. (2016). Accelerated ice shelf rifting and retreat at Pine Island Glacier, West Antarctica. *Geophysical Research Letters*, 43(22): 11–720.

Joughin, I., et al. (1999). Tributaries of West Antarctic ice streams revealed by RADARSAT interferometry. *Science*, 286(5438): 283–286.

Joughin, I. (2014). Marine ice sheet collapse potentially under way for the Thwaites Glacier Basin, West Antarctica. *Science,* 344: 735–738.

Khazendar, A., et al. (2015). The evolving instability of the remnant Larsen B Ice Shelf and its tributary glaciers. *Earth and Planetary Science Letters*, 419: 199–210.

Khazendar, A., et al. (2009). Roles of marine ice, rheology, and fracture in the flow and stability of the Brunt/Stancomb-Wills Ice Shelf. *Journal of Geophysical Research: Earth Surface*, 114(F4).

Kingslake J., et al. (2017). Widespread movement of meltwater onto and across Antarctic ice shelves. *Nature*, 544(7650): 349–352.

Kopp, R.E., et al. (2010). The impact of Greenland melt on local sea levels: A partially coupled analysis of dynamic and static equilibrium effects in idealized water-hosing experiments: A letter. *Climatic Change*, 103: 619–625.

Kopp, R., et al. (2023). Communicating future sea-level rise uncertainty and ambiguity to assessment users. *Nature Climate Change*, 13(7): 648–660.

Kyrke-Smith, T.M., et al. (2014). Subglacial hydrology and the formation of ice streams. *Proceedings of the Royal Society A*, 470(2161).

Lenaerts, J.T.M. et al. (2019). Observing and modeling ice sheet surface mass balance. *Reviews of Geophysics*, 57(2): 376–420.

Lempert, R.J. and Collins, M.T. (2007). Managing the risk of uncertain threshold responses: Comparison of robust, optimum, and precautionary approaches. *Risk Analysis: An International Journal*, 27(4): 1009–1026.

Lewis, E.L. and Perkin, R.G. (1986). Ice pumps and their rates. *Journal of Geophysical Research: Oceans*, 91(C10): 11756–11762.

Li, T., et al. (2023). Grounding line retreat and tide-modulated ocean channels at Moscow University and Totten Glacier ice shelves. *East Antarctica, the Cryosphere*, 17, 1003–1022.

Li, D., et al. (2023). Climate model differences contribute deep uncertainty in future Antarctic ice loss. *Science Advances,* 9(7).

Lloyd, A.J., et al. (2020). Seismic structure of the Antarctic upper mantle imaged with adjoint tomography. *Journal of Geophysical Research: Solid Earth*, 125(3).

MacGregor, J.A., et al. (2012). Widespread rifting and retreat of ice-shelf margins in the eastern Amundsen Sea Embayment between 1972 and 2011. *Journal of Glaciology*, 58(209): 458–466.

Mantelli, E., et al. (2019). Ice sheet flow with thermally activated sliding. Part 1: The role of advection. *Proceedings of the Royal Society A*, 475(2230): 20190410.

Mantelli, E. and Schoof, C. (2019). Ice sheet flow with thermally activated sliding. Part 2: The stability of subtemperate regions. *Proceedings of the Royal Society A*, 475(2231): 20190411.

Martin-Español, A. et al. (2017). Constraining the mass balance of East Antarctica. *Geophysical Research Letters,* 44(9): 4168–4175.

McDougall, T.J., et al. (2014). Melting of ice and sea ice into seawater and frazil ice formation. *Journal of Physical Oceanography*, 44(7): 1751–1775.

Miles, B.W.J., et al. (2020). Intermittent structural weakening and acceleration of the Thwaites Glacier Tongue between 2000 and 2018. *Journal of Glaciology*, 66: 485–495.

Milne, G.A. and Mitrovica, J.X. (1998). Postglacial sea-level change on a rotating Earth. *Geophysical Journal International*, 133(1): 1–19

Mitrovica, J.X., et al. (2011). On the robustness of predictions of sea level fingerprints. *Geophysical Journal International*, 187(2): 729–742.

Moreira, L., et al. (2021). Sea-level fingerprints due to present-day water mass redistribution in observed sea-level data. *Remote Sensing*, 13(4667).

Morlighem, M., et al. (2020). Deep glacial troughs and stabilizing ridges unveiled beneath the margins of the Antarctic ice sheet. *Nature Geoscience,* 13(2): 132–137.

Mottram, R., et al. (2021). What is the surface mass balance of Antarctica? An intercomparison of regional climate model estimates. *The Cryosphere*, 15: 3751–3784.

Needell, C. and Holschuh, N. (2023). Evaluating the retreat, arrest, and regrowth of Crane Glacier against marine ice cliff process models. *Geophysical Research Letters*, 50(4): e2022GL102400.

Otosaka, I.N., et al. (2023). Mass balance of the Greenland and Antarctic ice sheets from 1992 to 2020. *Earth System Science Data*, 15: 1597–1616.

Paterson, W.S.B. and Budd, W.F. (1982). Flow parameters for ice sheet modeling. *Cold Regions Science and Technology,* 6(2): 175–177.

Pegler, S.S. (2016). The dynamics of confined extensional flows. *Journal of Fluid Mechanics*, 804: 24–57.

Peltier, W.R. (1974). The impulse response of a Maxwell Earth. *Reviews of Geophysics*, 12(4): 649–669

Pollard, D. et al. (2015). Potential Antarctic Ice Sheet retreat driven by hydrofracturing and ice cliff failure. *Earth and Planetary Science Letters,* 412: 112–121.

Reese, R., et al. (2018). Grounding-line flux formula applied as a flux condition in numerical simulations fails for buttressed Antarctic ice streams. *The Cryosphere*, 12(10): 3229–3242.

Rignot, E., et al. (2011). Antarctic grounding line mapping from differential Satellite Radar interferometry. *Geophysical Research Letters*, 38. 10.1029/2011GL047109.

Rignot, E., et al. (2014). Widespread, rapid grounding line retreat of Pine Island, Thwaites, Smith, and Kohler glaciers, West Antarctica, from 1992 to 2011. *Geophysical Research Letters,* 41(10): 3502–3509.

Rignot, E., et al. (2019). Four decades of Antarctic Ice Sheet mass balance from 1979–2017. *Proceedings of the National Academy of Sciences*, 116(4): 1095–1103.

Rignot, E., et al. (2024). Widespread seawater intrusions beneath the grounded ice of Thwaites Glacier, West Antarctica. *Proceedings of the National Academy of Sciences*, 121(22): e2404766121

Robel, A., et al. (2019). Marine ice sheet instability amplifies and skews uncertainty in projections of future sea-level rise. *Proceedings of the National Academy of Sciences*, 116: 2019 04822.

Roffman, J., et al. (2023). Spatial and temporal variability of 21st century sea level changes, *Geophysical Journal International*, 235(1): 342–352

Rosevear, M.G., et al. (2025). How does the ocean melt Antarctic Ice Shelves? *Annual Review of Marine Science*, 17: 325–353.

Sadai, S., et al. (2022). The Paris Agreement and climate justice: Inequitable impacts of sea level rise associated with temperature targets. *Earth's Future*, 10(12). 10.1029/2022EF002940.

Scambos, T.A., et al. (2004). Glacier acceleration and thinning after ice shelf collapse in the Larsen B embayment, Antarctica. *Geophysical Research Letters*, 31(18). 10.1029/2004GL020670.

Schoof, C. (2007a). Ice sheet grounding line dynamics: Steady states, stability, and hysteresis. *Journal of Geophysical Research - Earth Surface*, 112 (F3). 10.1029/2006JF000664.

Schoof, C. (2007b). Marine ice-sheet dynamics. Part 1. The case of rapid sliding. *Journal of Operations Management*, 573: 27–55.

Schoof, C., et al. (2017). Boundary layer models for calving marine outlet glaciers. *The Cryosphere*, 11: 2283–2303.

Schoof, C. and Mantelli, E. (2021). The role of sliding in ice stream formation. *Proceedings of the Royal Society A*, 477(2248): 20200870.

Seroussi, H., et al. (2017). Continued retreat of Thwaites Glacier, West Antarctica, controlled by bed topography and ocean circulation. *Geophysical Research Letters*, 44(12): 6191–6199.

Seroussi, H., et al. (2020). ISMIP6 Antarctica: A multi-model ensemble of the Antarctic ice sheet evolution over the 21st century. *The Cryosphere*, 14(9): 3033–3070.

Seroussi, H., et al. (2024). Evolution of the Antarctic Ice Sheet over the next three centuries from an ISMIP6 model ensemble. *Earth's Future*. 10.1029/2024EF004561.

Shiklomanov, I. (1993). In World fresh water resources. In *Water in Crisis: A Guide to the World's Fresh Water Resources*. Ed: P.H. Gleick. Oxford University Press, New York, 13–24.

Smith, B., et al. (2020). Pervasive ice sheet mass loss reflects competing ocean and atmosphere processes. *Science*, 368(6496): 1239–1242.

Still, H. et al. (2019). Mechanical analysis of pinning points in the Ross Ice Shelf, Antarctica. *Annals of Glaciology*, 60(78): 32–41.

Stokes, C.R., et al. (2019). Widespread distribution of supraglacial lakes around the margin of the East Antarctic Ice Sheet. *Scientific Reports*, 9(1): 13823.

Sun, S., et al. (2020). Antarctic ice sheet response to sudden and sustained ice-shelf collapse (ABUMIP). *Journal of Glaciology*, 66(260): 891–904.

Thomas, R.H. and Bentley, C.R. (1978). A model for Holocene retreat of the West Antarctic ice sheet. *Quaternary Research*, 10(2): 150–170.

Tinto, K.J., et al. (2019). Ross Ice Shelf response to climate driven by the tectonic imprint on seafloor bathymetry. *Nature Geoscience*, 12: 441–449

Treverrow, A., et al. (2012). The tertiary creep of polycrystalline ice: Experimental evidence for stress-dependent levels of strain-rate enhancement. *Journal of Glaciology*, 58(208): 301–314.

Trusel, L.D., et al. (2015). Divergent trajectories of Antarctic surface melt under two twenty-first-century climate scenarios. *Nature Geoscience*, 8(12): 927–932.

Tulaczyk, S., et al. (2000). Basal mechanics of ice stream B, West Antarctica: 1. Till mechanics. *Journal of Geophysical Research: Solid Earth*, 105(B1): 463–481.

van de Wal, R.S.W., et al. (2022). A high-end estimate of sea level rise for practitioners. *Earth's Future*, 10(11), e2022EF002751.

Velicogna, I., et al. (2014). Regional acceleration in ice mass loss from Greenland and Antarctica using GRACE time-variable gravity data. *Geophysical Research Letters*, 41(22): 8130–8137.

Walker, C.C, and Gardner, A.S. (2019). Evolution of ice shelf rifts: Implications for formation mechanics and morphological controls. *Earth and Planetary Science Letters*, 526: 115764.

Walker, C.C., et al. (2021). A high resolution, three-dimensional view of the D-28 Calving event from Amery Ice Shelf with ICESat-2 and Satellite Imagery. *Geophysical Research Letters*, 48(3), e2020GL091200.

Warner, R.C., et al. (2021). Rapid formation of an ice doline on Amery Ice Shelf, East Antarctica. *Geophysical Research Letters*, 48(14), e2020GL091095.

Weertman, J. (1974). Stability of the junction of an ice sheet and an ice shelf. *Journal of Glaciology*, 13(67): 3–11.

Whitehouse, P.L. (2018). Glacial isostatic adjustment modelling: Historical perspectives, recent advances, and future directions. *Earth Surface Dynamics*, 6: 401–429.

Winkelmann, R. et al. (2023). Tipping points in the cryosphere. *The Global Tipping Points Report*, 2023.

Woodward, R.S. (1888). On the form and position of mean sea level. *United States Geological Survey Bulletin*, 48: 87–170.