FACULTÉ
DES SCIENCES

UNIVERSITÉ LIBRE DE BRUXELLES

THE LONG-TERM FUTURE OF THE ANTARCTIC ICE SHEET

Uncertainties in ice sheet–Earth system interactions

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in fulfilment of the requirements for the academic degree of

Docteur en Sciences

Academic year 2022–2023

Under the supervision of Professor Frank PATTYN

Laboratoire de Glaciologie

CRYOSPHERE & SEA LEVEL

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This work was financially supported by the Fonds de la Recherche Scientifique de Belgique (F.R.S.-FNRS) through an F.R.S.-FNRS Research Fellowship and by the European Union's Horizon 2020 research and innovation programme under grant agreement No 869304.

Computational resources have been provided by the Shared ICT Services Centre, Université Libre de Bruxelles and by the Consortium des Équipements de Calcul Intensif (CÉCI), funded by the Fonds de la Recherche Scientifique de Belgique (F.R.S.-FNRS) under Grant No 2.5020.11 and by the Walloon Region.

ACKNOWLEDGEMENTS

First of all, I would like to thank Frank for being such a great supervisor. You have provided me with an amazing balance between guidance and support whenever I needed it, and also trust and independence. You have been (and still are) a real mentor. Thank you for helping me grow as a scientist.

I would also like to thank the members of my thesis committee for their presence and guidance all along the way.

I would like to acknowledge the precious help, advice and feedbacks provided by Frank, Lars, and all of my co-authors, which have significantly contributed to the improvement of this thesis.

Enormous thanks to all of the team from the Labo de Glacio for the amazing daily atmosphere, making it a genuine pleasure to go to work. I feel very lucky. Special thanks to the GEMICE group (including the remote ones) for your presence, help and constant support. Thanks also to the old gang, Sophie, Brice, Marie, Fanny, Caro, Mana, ... the beginning of my PhD would not have been the same without you.

Special thanks to Kevin, for his countless time and more than precious help. Working with you has been a real turning point in my PhD.

Thanks also to Heiko, for sharing these mind-breaking discussions which have taught me a lot, for your time, your guidance, for the trust and credit that you have given me, and for being an inspiration on how to think and behave as a scientist.

Thanks to Elise, for being my binôme and friend and for your support since day 1.

Thanks to Ann Kristin, my remote working partner for the past two years. Working with you has been an amazing source of motivation. More generally, I would like to thank the PROTECT members thanks to whom I have discovered the pleasure of working together on a scientific project.

Special thanks again to Lars, who has been my wingman in tackling fetish, leading to numerous 'VL' versions of the model which drove Frank (and ourselves) crazy, but also had, I believe, a significant impact on the results presented in this thesis.

Thanks to Konz, for your friendship and constant presence and support, especially in challenging moments, and for what you brought to the lab and the GEMICE group.

Thanks to Sainan, for your advice and your presence during so much time. I still miss hearing your 'Viu' from behind my screen.

Thanks to Christoph, for being so friendly and available since our very first exchange, and for your precious help and advice, which have provided me with lots of support and motivation.

This PhD has been a truly enriching experience, which made me gain true friendships along the way. Special thoughts to the EGU Cryo ECS team and to Rebecca, Liz, Julius and many others.

Pour finir, merci merci merci à mes amis et à ma famille pour votre soutien sans faille, chacun à votre façon. Mentions spéciales à John pour son soutien quotidien (et pour m'avoir nourrie et supportée ces derniers mois), à Jobi pour ses playlists, et à Jude pour avoir suivi chaque étape de cette thèse; tes questions et ton intérêt ont été une incroyable source de motivation.

SUMMARY

The Antarctic ice sheet (AIS) is the largest and yet the most uncertain potential contributor to future sea-level rise. While recent satellite observations have highlighted that the AIS is currently losing mass at an accelerating rate, projections of the future evolution of the ice sheet in a warming climate remain highly uncertain. As future sea-level rise is probably one of the biggest threats imposed on us by climate change, predicting it with the lowest possible uncertainty is of capital societal importance. Uncertainties in the future evolution of the AIS can be explained, notably, by the fact that the ice sheet is capable of abrupt and self-sustained changes associated with several positive feedback mechanisms, especially in its marine areas, i.e., where the ice lies on bedrock below sea level. This is the case for most of the West Antarctic ice sheet (WAIS) as well as for some basins of the East Antarctic ice sheet (EAIS). The interactions between the ice sheet and its surrounding environment (namely the ocean, the atmosphere, and the solid Earth) have been shown to strongly influence its stability, more particularly by triggering or dampening the instabilities threatening the ice sheet. Despite the uncertainties, recent studies suggest that the WAIS will lose mass in the future and eventually (partially) collapse. The uncertainties pertain to when, and to whether the weak Earth structure beneath that area of the ice sheet may be a stabilising factor, as a rapid bedrock uplift in response to ice mass loss has been shown to delay or even limit mass loss. The fate of the EAIS is less clear. A pending question is: will the EAIS lose or gain mass in the future? More specifically, will the grounding line retreat in its marine basins, and if so, can the associated mass loss be compensated by sufficient mass gain due to increased snow accumulation in the interior of the ice sheet?

In this thesis, we contribute to clarifying and providing new insights to these questions, and therefore on the long-term future of the AIS. Using a numerical ice-sheet model, we investigate the influence of uncertainties in ice sheet–Earth system interactions on its future stability. In addition, we produce observationally-calibrated projections and associated quantified uncertainties of the evolution of the AIS over the millennium. Our results show that the ocean will be the main driver of Antarctic short-term mass loss, leading to significant retreat in the WAIS (especially in the Amundsen Sea Embayment), even under limited warming. Under sustained warming, however, this may lead to a complete WAIS collapse over the course of the millennium, despite a stabilising weak solid Earth structure beneath West Antarctica. In addition, our results suggest that a sustained warming will likely turn the EAIS into a positive contributor to sea-level rise over the course of the next century. Indeed, we project that the ocean-driven grounding-line retreat in its marine basins, which cannot be efficiently stabilised by bedrock uplift given the rigid structure of the solid Earth in that area, will progressively outweigh the surface mass balance (SMB, the balance between accumulation, sublimation and runoff at its surface). Finally, we show that the mitigating role of the SMB may strongly be reduced under sustained warming, due to a significant increase in surface runoff with increasing temperatures, hence further increasing the net AIS contribution to sea-level rise.

RÉSUMÉ

La calotte glaciaire Antarctique est le plus gros contributeur potentiel à l'élévation future du niveau marin global, mais aussi le plus incertain. De plus en plus d'observations satellitaires mettent en évidence une actuelle perte de masse de l'Antarctique, et ce à un rythme accéléré. Malgré cela, les projections de l'évolution future de la calotte Antarcique en réponse aux changements climatiques actuels et à venir restent très incertaines. Pourtant, être en mesure de prédire avec la plus faible incertitude possible l'élévation future du niveau de la mer est d'une importance sociétale capitale. Les incertitudes sur l'évolution future de l'Antarctique s'expliquent notamment par le fait que la calotte glaciaire est capable de changements brusques associés à plusieurs mécanismes de rétroactions positives, et ce particulièrement dans les régions où la glace repose sur un socle rocheux situé sous le niveau de la mer. C'est le cas pour la majorité de l'Antarctique de l'Ouest ainsi que dans certains bassins de l'Antarctique de l'Est. Il a été démontré que les interactions entre la calotte glaciaire et son environnement (à savoir l'océan, l'atmosphère et le lit rocheux sous-jacent) influencent fortement sa stabilité, notamment en déclenchant ou en atténuant les instabilités qui la menacent. Malgré les incertitudes, des études récentes suggèrent que, à l'avenir, l'Antarctique de l'Ouest perdra de la masse et finira par (au moins partiellement) s'effondrer. Les incertitudes concernent dès lors le moment de cet effondrement, mais également une potentielle stabilisation de la calotte par le lit rocheux (qui repose sur une région du manteau terrestre particulièrement peu visqueuse) situé sous cette région occidentale. En effet, il a été démontré qu'un rebond rapide du lit rocheux pouvait ralentir voire limiter la perte de masse de la calotte. Le sort de l'Antarctique de l'Est est moins clair. Une des questions en suspens est la suivante : l'Antarctique de l'Est perdra-t-elle ou gagnera-t-elle de la masse ? Plus précisément, perdra-t-elle de la masse dans ses régions potentiellement instables, et, si oui, cette perte de masse sera-t-elle compensée par un gain de masse associé à une augmentation des précipitations neigeuses à l'intérieur du continent ?

Cette thèse contribue à clarifier et à apporter de nouveaux éléments de réponses à ces questions, permettant ainsi d'en savoir plus sur l'évolution future de la calotte Antarctique. Pour ce faire, nous étudions, à l'aide d'un modèle numérique de calotte glaciaire, l'influence des incertitudes concernant les interactions de la calotte avec son environnement sur sa stabilité future. De plus, nous produisons, tout en quantifiant les incertitudes, des projections de l'évolution de l'Antarctique au cours du prochain millénaire. Nos résultats montrent que l'océan sera le principal moteur de la perte de masse à court terme de l'Antarctique, engendrant un recul important en Antarctique de l'Ouest (et plus particulièrement au niveau de la mer d'Amundsen), et ce même en cas de réchauffement climatique limité. Dans le cas d'un réchauffement soutenu, cependant, cela pourrait conduire, malgré le rebond du lit rocheux, à un effondrement complet de l'Antarctique de l'Ouest au cours du millénaire. De plus, nos résultats suggèrent qu'un réchauffement soutenu transformera probablement l'Antarctique de l'Est en un contributeur positif à l'élévation du niveau marin global au cours du siècle prochain. En effet, nous prévoyons que le recul de la calotte dans ses bassins instables, non efficacement stabilisé par le rebond du lit rocheux étant donné le manteau terrestre très visqueux dans cette région, l'emportera progressivement sur son bilan de masse de surface (c'est-à-dire l'équilibre entre accumulation, sublimation et ruissellement à la surface de la calotte). Enfin, nous démontrons que le bilan de masse de surface peut être fortement réduit en cas de réchauffement soutenu, en raison d'une augmentation significative du ruissellement de surface avec l'augmentation des températures, augmentant ainsi la contribution nette de l'Antarctique à l'élévation du niveau marin.

CONTENTS

List of acronyms

AIS	Antarctic ice sheet
AR5	IPCC fifth Assessment Report
AR6	IPCC sixth Assessment Report
ASE	Amundsen Sea Embayment
CDW	Circumpolar Deep Water
CMIP5	Coupled Model Intercomparison Project Phase 5
CMIP6	Coupled Model Intercomparison Project Phase 6
EAIS	East Antarctic ice sheet
ESM	Earth System Model
f.ETISh	fast Elementary Thermomechanical Ice Sheet
GHG	Greenhouse Gas
GIA	Glacial Isostatic Adjustment
GMSL	Global mean sea level
GRD	Earth Gravity, Earth Rotation and viscoelastic solid Earth Deformation
IPCC	Intergovernmental Panel on Climate Change
ISMIP6	Ice Sheet Model Intercomparison Project for CMIP6
LARMIP	Linear Antarctic Response Model Intercomparison Project
PICO	Postdam Ice-shelf Cavity mOdel
PDD	Positive Degree-Day
RCM	Regional Climate Model
RCP	Representative Concentration Pathway
RSL	Relative Sea Level

SGVEM	Self-Gravitating Earth Model
SIA	Shallow-ice approximation
SLR	Sea Level Equivalent
SLR	Sea Level Rise
SMB	Surface mass balance
SROCC	Special Report on the Ocean and Cryosphere in a Changing Climate
SSA	Shallow-shelf approximation
SSP	Shared Socio-economic Pathway
UQ	Uncertainty quantification
VAF	Volume Above Flotation
WAIS	West Antarctic ice sheet

LIST OF PUBLICATIONS

Goelzer, H., **Coulon, V.**, Pattyn, F., de Boer, B., and van de Wal, R.: Brief communication: On calculating the sea-level contribution in marine ice-sheet models, *The Cryosphere*, 14, 833–840, https://doi.org/10.5194/tc-14-833-2020, 2020.

Coulon, V., Bulthuis, K., Whitehouse, P. L., Sun, S., Haubner, K., Zipf, L., and Pattyn, F.: Contrasting Response of the West and East Antarctic ice sheets to glacial isostatic adjustment, *Journal of Geophysical Research: Earth Surface*, 126, doi:10.1029/2020JF006003, 2021

Kazmierczak, E., Sun, S., **Coulon, V.**, and Pattyn, F.: Subglacial hydrology modulates basal sliding response of the Antarctic ice sheet to climate forcing, The Cryosphere Discuss., https://doi.org/10.5194/tc-2022-53, in review, 2022.

Coulon, V., Klose, AK., Kittel, C., Winkelmann, R., and Pattyn, F. Projections of long-term future Antarctic ice loss with a historically-calibrated ice-sheet model, in preparation.

Klose, AK., **Coulon, V.**, Winkelmann, R., and Pattyn, F. The long-term sea-level commitment from Antarctica, in preparation.

CHAPTER 1

Antarctica and sea level

1.1 Introduction

Since the preindustrial era, the global surface temperature has increased by 1.07 (0.8 to 1.3)[1] °C from 1850–1900 to 2010–2019, with each of the last four decades being successively warmer than any decade that preceded (IPCC, 2021). As a consequence of this continued warming, global sea level rose during the 20th century more than three times faster than at any time during the last 2000 years (Kopp et al., 2016). The global mean sea level has increased by about 0.20 m between 1901 and 2018, and the average rate of sea level rise was 1.3 [0.6 to 2.1] $\mathrm{mm\,yr^{-1}}$ between 1901 and 1971, increasing to 1.9 [0.8 to 2.9] $\mathrm{mm\,yr^{-1}}$ between 1971 and 2006, and further increasing to 3.7 [3.2 to 4.2] $\mathrm{mm\,yr^{-1}}$ between 2006 and 2018 (IPCC, 2021). Current estimates predict that if the Paris Agreement targets are surpassed, sea-level rise could exceed one meter by the beginning of the 22nd century. According to the latest IPCC report, global mean sea level (GMSL) is projected to rise (relative to the 1995-2014 average) by a likely range of 0.28–0.55 m under a low GHG emissions scenario (SSP1-1.9) and by 0.63–1.01 m under the very high-emissions scenario (SSP5-8.5) over the course of the twenty-first century (IPCC, 2021, see Figure 1.1). Even under the more optimistic climate scenarios, sea-level rise (SLR) projections are sufficient to more than double the frequency of extreme coastal flooding events and impair the habitability of low-lying Pacific Islands (Vitousek et al., 2017). Therefore, with about 30% of the world population living on coastal areas, sea-level rise is probably one of the biggest threats imposed on us by climate change, with a very high social risk to human coastal settlements. According to a study by the world bank, the one meter of sea-level rise expected during the course of the next century under unabated warming pathways could cost up to 18 trillion USD just to protect populated coastal areas worldwide (Nicholls et al., 2019).

A key question remains 'how much, how fast': what rise of sea level is expected and on what time scales ranging from decades, centuries to millennia? This requires a comprehensive understanding of all processes interacting within the Earth System, of which the large ice sheets on Earth are an important component. Not only do ice sheets respond to changes in the climate through waxing and waning, they also influence and modify the climate itself through many feedback mechanisms.

Understanding the precise response of ice sheets to expected changes in the climate is still in its infancy. Prior to 2014, sea-level projections were based on thermosteric and mass balance estimates using atmo-

[1]In IPCC reports, square brackets [x to y] are typically used to provide the assessed very likely range, or 90% interval, while simple brackets (x to y) assess the likely range, i.e., the 66% interval.

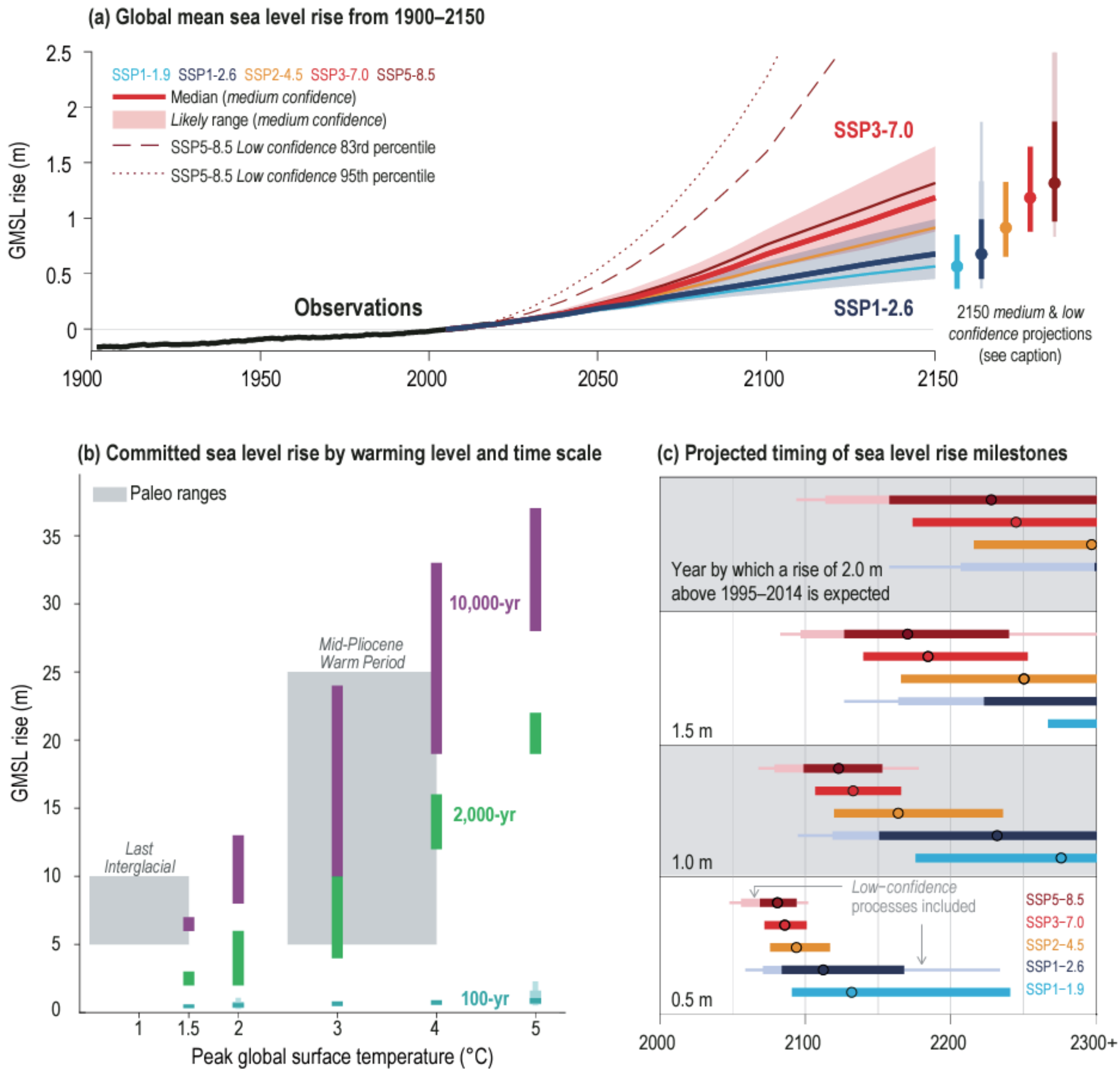

Figure 1.1: Global mean sea level (GMSL) change on different time scales and under different scenarios [Credit: IPCC (2021)]. The lightly shaded thick/thin bars show 17th–83rd/5th–95th percentile low-confidence ranges in 2150 for SSP1-2.6 and SSP5-8.5, based upon projection methods incorporating structured expert judgement and marine ice cliff instability (see section 1.2.3.1.1).

sphere and ocean models. The 5th assessment report of IPCC was the first step to include ice-dynamical changes, albeit in a statistical assessment. While the 6th assessment report uses physically-based ice-sheet model projections to gauge the dynamical contribution of the Greenland and Antarctic ice sheets to sea level rise, it notes that *'under the higher* $\mathrm{CO_2}$ *emissions scenarios, there is deep uncertainty in sea-level projections for 2100 and beyond associated with the ice-sheet responses to warming'* (IPCC, 2021). In particular, the Antarctic Ice Sheet (AIS) remains a climate component still largely misunderstood, especially in predicting global sea-level changes. Of all the components of the global sea-level budget, the contribution of the AIS is the most uncertain.

At present, mountain glaciers and the Greenland and Antarctic ice sheets are all contributing roughly equally to sea-level rise through melting (IPCC, 2021), but this might change dramatically in the near future. With 90% of all ice on Earth (representing about 60 meters of equivalent SLR – enough to wipe out all coastal cities of the world), major efforts are needed to improve our understanding of the AIS dynamics on different time scales in a reaction to global climate change.

This can be done using numerical process-based ice-sheet models, which have the advantage of being useful to make predictions of future ice-sheet changes. Moreover, ice-sheet models are ideal tools to investigate the sensitivity of an ice sheet to changes in the environment, enabling us identify processes that control its sensitivity. However, many uncertainties remain, both in the magnitude and exact nature of future climate changes, as well as in the physical processes controlling the ice dynamics (and hence the response of the ice sheet).

The more we know and understand, the more we will be able to reduce the uncertainty in AIS projections, and the more efficient we will be able to adapt and help stakeholders define priorities in mitigation and adaptation strategies to reduce the socio-economic and ecological impacts of future global mean sea-level rise and sea-level extremes. Moreover, the better our predictions and understanding of the ice sheet behaviour and sensitivity, the more efficient we will be able to communicate and convince in the hope of avoiding the crossing of irreversible tipping points[1]. We have probably already induced ice-sheet dynamical processes that will last for centuries, but we still have ways to reduce the damage significantly (Edwards et al., 2021). What will happen in the next decades to the Antarctic ice sheet has the potential to be a complete game changer for humanity. Decisions on mitigation of emissions in the near future (decades to centuries) may trigger ice-sheet mass changes that continue for centuries to millennia, due to the long-term and potentially irreversible nature of their responses (Fox-Kemper et al., 2021).

With this thesis, I wish and hope to contribute to a better understanding of the behaviour and dynamics of the Antarctic ice sheet under a warming climate using a numerical ice-sheet model, both by estimating its contribution to future global mean sea-level rise, and by investigating the interaction of the ice sheet with the Earth system.

[1] A tipping point may be defined as the critical point in a situation, process, or system beyond which a significant and often unstoppable effect or change takes place.

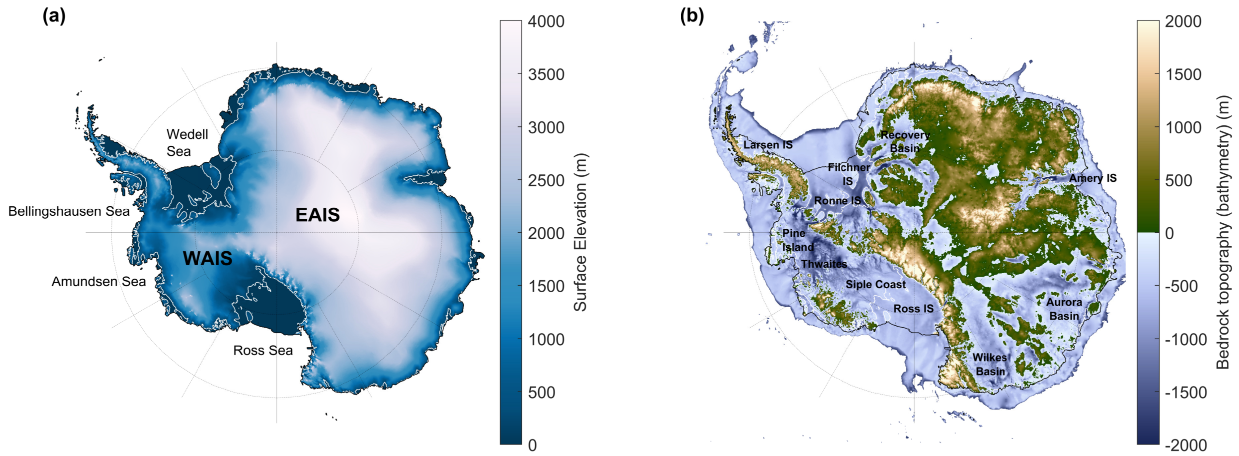

Figure 1.2: Present-day Antarctic ice sheet surface elevation (a) and bedrock topography (b) (Morlighem et al., 2020) with characteristic regions labelled in black; WAIS stand for West Antarctic ice sheet, EAIS for East Antarctic ice sheet, and IS for ice shelf. The grounding line is shown in white and the ice front is shown in black.

1.2 The Antarctic ice sheet and its contribution to sea-level changes

1.2.1 How does the Antarctic ice sheet contribute to sea-level changes?

1.2.1.1 The ice-sheet mass balance

The Antarctic ice sheet (Figure 1.2) is a perennial mass of ice centered around the South Pole that covers an area of about 14 million km^2[1], representing ~8% of the global land surface (Fretwell et al., 2013). With a mean ice thickness of ~2 km, the AIS contains enough ice to raise the global mean sea level by 58 meters (Fretwell et al., 2013).

The AIS, like all ice sheets, grows by successive accumulation of snow at the surface of a landmass. Once accumulated, snow progressively becomes more dense, turns into ice, and flows (driven by gravity) towards the ocean. As the ice thins towards the edge of the ice sheet, its thickness becomes equal to the buoyant thickness of ice and it starts to float on the ocean, forming (sometimes very large – see Figure 1.2) ice shelves. The location where the ice becomes afloat is called the grounding line. These ice shelves, in direct contact with the ocean, progressively melt or calve icebergs. About 10% of the AIS area is constituted by ice shelves, whom are largely located in the West part of the ice sheet (called the West Antarctic ice sheet – WAIS), with the largest ice shelves being the Ross and the Ronne-Filchner ice shelves (see Figure 1.2).

By definition, the **mass balance** of an ice sheet is the trade-off between total ice mass gains and total ice mass losses. The dominant source of mass gain is precipitation, i.e., the transfer of ice particles (snow) or water droplets (rain) from the atmosphere to the ice-sheet surface. Sources of mass loss are more diverse. At the ice-sheet surface, mass is lost when melt water is not retained in the firn (an intermediate stage between snow and glacial ice, representing a layer of old – i.e., multiyear – compressed snow) by refreezing and/or capillary forces and leaves the ice sheet as runoff. Snow can also be redistributed by the

[1]Twice as large as Australia

wind (erosion/deposition) and/or sublimate. At the ice–bedrock interface, mass may also be lost by basal melting, though it generally represents a small term. Finally, additional sources of mass loss appear at the ice–ocean interface, which, in the case of the AIS, occurs essentially in the ice shelves (Bindschadler et al., 2011). Mass loss in the ice shelves appears in the form of (1) melting of the ice-shelf base that is in contact with the ocean (basal melting), and (2) iceberg calving at the ice shelf front that drift away and melt in the ocean (Depoorter et al., 2013). Under 'cold-water' ice shelves (such as Ross, Ronne-Filchner and Amery) rising plumes of buoyant meltwater can also lead to refreezing downstream of the grounding line, creating a layer of so-called 'marine ice' on the ice-shelf base (Lewis and Perkin, 1986). Under the Ronne and Amery ice shelves, such refreezing rates can locally be high, leading to marine ice fields hundreds of meters thick (Adusumilli et al., 2020).

It is important to note that, apart from a negligible halosteric effect due to the difference between freshwater (melted ice) and saline ocean water densities, the melting of any ice located below buoyancy level does not contribute to global mean sea-level changes as it is already displacing sea water (Goelzer et al., 2020; Gregory et al., 2019). Therefore, melting of the floating ice shelves (whether by calving, sub-shelf or surface melt) does not directly lead to sea-level rise. The ice-sheet **mass balance** and its **contribution to sea-level rise** are thus, in fact, two different quantities. In principle, to be representative of the ice-sheet **contribution to sea level**, mass balance can alternatively be (and often is) expressed as the balance between the surface mass balance (SMB – the net difference between precipitation, sublimation, evaporation, wind erosion and melt water runoff) *over the grounded parts of the ice sheet* and the ice discharge, i.e., the amount of ice that flows across the grounding line (driven by basal sliding and internal deformation) and starts to float, forming ice shelves. Nevertheless, in the case of a marine-terminating ice mass lying on bedrock below sea level (as may for example be observed in the WAIS – see Figure 1.2b), corrections for the ice that crosses the grounding-line while being already below buoyancy still need to be applied in order to distinguish the ice-sheet contribution to sea-level from its mass imbalance (Bamber et al., 2018). Indeed, land ice contribution to global mean sea-level changes must exclude the mass whose liquid-water equivalent volume equals the volume of sea water already displaced (Gregory et al., 2019). The remainder, which is not currently displacing sea water, is referred to as the ice mass or volume above flotation.

Overall, the SMB over grounded ice is (essentially via snow accumulation) the only potential negative contributor to sea-level rise (Medley and Thomas, 2019). Therefore, the case of a negative SMB is sometimes referred to as a tipping point for ice sheet mass loss (Robinson et al., 2012).

1.2.1.2 The Antarctic ice shelves – sources of feedbacks

Despite their negligible direct contribution to sea-level changes, the ice shelves – the floating extensions of ice sheets, connected to and nourished by land-based ice – are key features of ice sheets because they control the flow of ice as it drains into the ocean, thereby moderating the pace of the mass loss. Being several hundreds of metres thick and interacting with land or shallow rocks, they act as natural barriers that restrain the grounded ice-sheet flow into the ocean. More specifically, in regions where ice shelves are laterally confined or locally run aground (such as at ice rises and ice rumples; Matsuoka et al., 2015), they generate resistive stresses that are transmitted upstream and slow down the flow of ice into the ocean (Fürst et al., 2016; Reese et al., 2018b). This restraining potential is called **buttressing**. When buttressing ice shelves thin, their front gradually retreat or, more drastically, when they disintegrate, the restraining

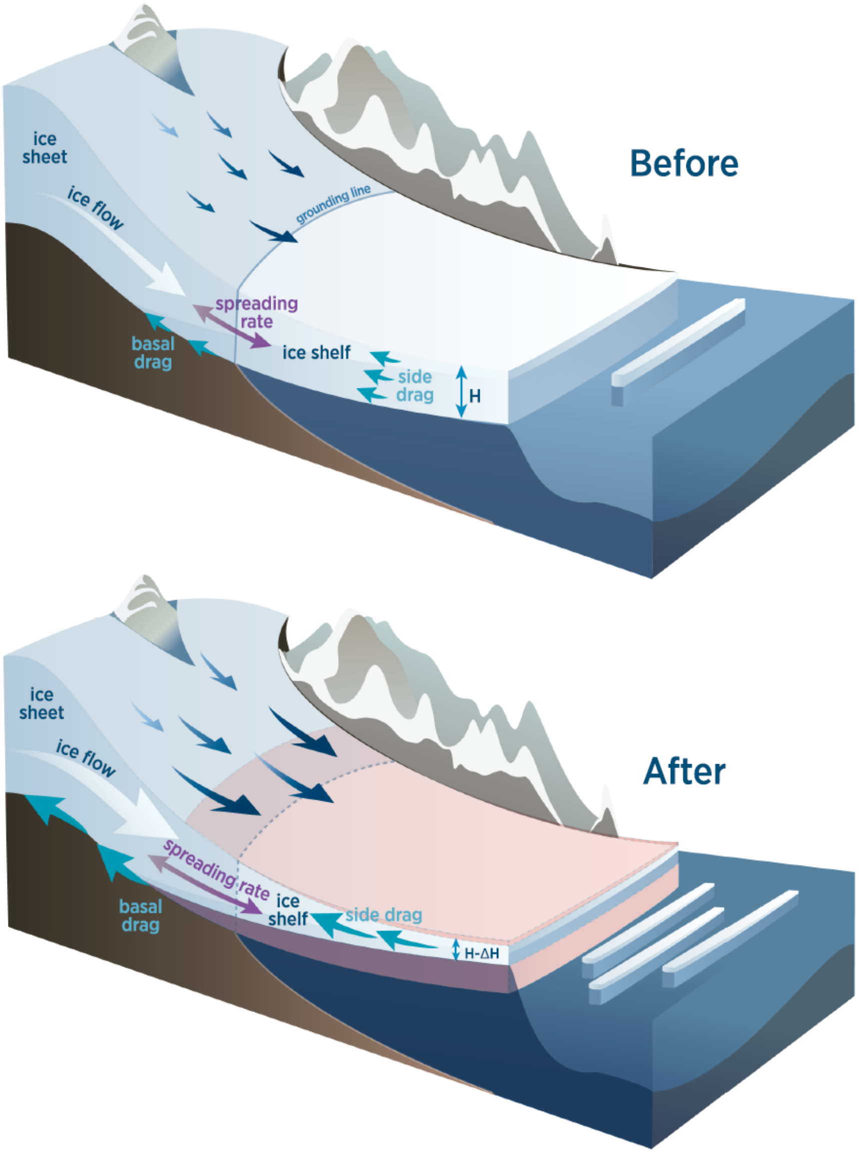

Figure 1.3: **Ice shelf buttressing and grounding line flux.** Confined ice shelves restrain the ice flow from upstream glaciers. Thinning of such a buttressing ice shelf increases the longitudinal stress and with it the spreading rate at the grounding line. This eventually results in an increase in the ice flux across the grounding line. Figure (a) shows the ice-shelf thickness (H) and ice flow prior to the onset of ice shelf thinning. Figure (b) illustrates the onset of ice-shelf thinning (thickness loss is shown in red), which reduces buttressing and increases longitudinal spreading and the ice flux across the grounding line, increase the ice discharge into the ocean. [Figure Credit: Gudmundsson et al. (2019)]

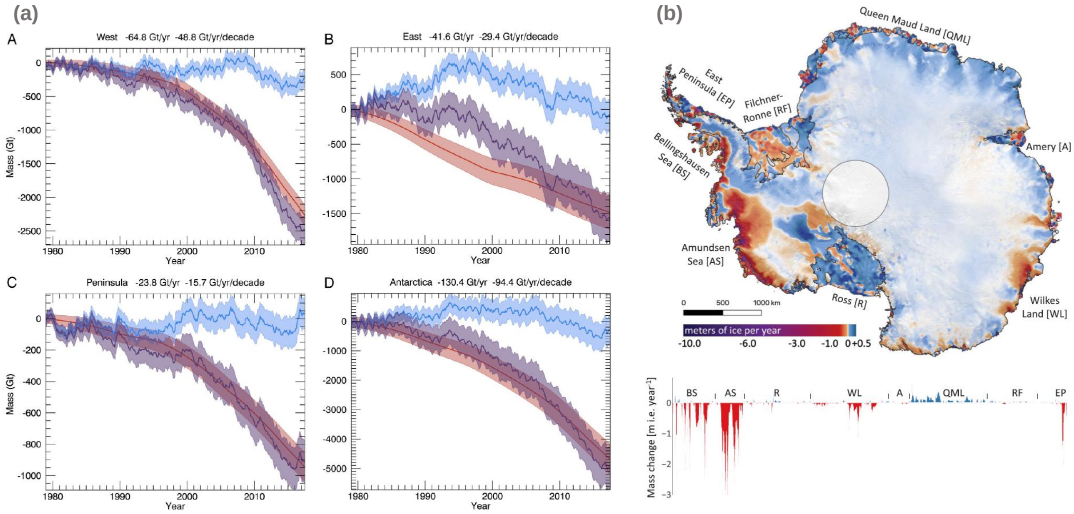

Figure 1.4: **Current trends in Antarctic ice mass changes.** Figure (a) shows time series of cumulative anomalies in SMB (blue), ice discharge (D, red), and total mass (M, purple) with error bars in billions of tons for (A) West Antarctica, (B) East Antarctica; (C) Antarctic Peninsula), and (D) Antarctica, with mean mass loss in billions of tons per year and an acceleration in billions of tons per year per decade for the time period 1979 to 2017 [Credit: Rignot et al. (2019)]. Figure (b) shows the mass change rates for Antarctica averaged over the period 2003 to 2019, with average mass change of the entire ice sheet shown in the top, and mass change rates at the grounding line shown in the bottom. Blue colors indicate mass gain while red colors indicate mass loss [Credit: Smith et al. (2020)].

force that they exert on the ice outflow at the grounding line is lowered and more ice is discharged into the ocean – upstream glaciers accelerate. Reese et al. (2018b) have shown that the buttressing provided by ice shelves can affect inland ice hundreds of kilometres away. Such loss of ice-shelf buttressing has effectively been witnessed in Antarctica at the beginning of the 21st century, where ice-shelf collapse in the Antarctic Peninsula led to increased glacier discharge (Scambos et al., 2004).

In addition, the buttressing potential of the ice shelves can lead to a stabilisation of the ice sheet or a slowdown of grounding-line retreat in areas where the bedrock configuration makes the ice sheet intrinsically unstable and therefore subject to self-sustained grounding line retreat (see coming section 1.2.3.1.1). For these reasons, ice shelves exert a crucial control on the dynamics of the grounded ice-sheet sectors (Gudmundsson, 2013), and ice shelf thinning and/or removal holds an indirect, albeit significant potential for a contribution to sea-level variations. Note that some ice shelves, or parts of them, are considered as ‘passive’, meaning that ice can be removed without major effects on the ice sheet dynamics (Fürst et al., 2016; Reese et al., 2018b). Notably, Fürst et al. (2016) and Reese et al. (2018b) suggest that the ice shelves in the Amundsen and (to a lesser degree) Bellingshausen seas have limited or almost no ‘passive’ portion, which implies that further retreat of current ice-shelf fronts will yield important dynamic consequences in these regions.

1.2.2 The present: current trends in Antarctic ice mass changes

During the last few decades, the AIS has contributed to about 10% of the observed SLR (Fox-Kemper et al., 2021). However, current observations agree in showing that the rate of AIS contribution to SLR is increasing (from 0.14 ± 0.02 mm yr^{-1} between 1992 and 2001 to 0.55 ± 0.07 mm yr^{-1} between 2012 and 2016; Oppenheimer et al., 2019, see Figure 1.4a). More specifically, these observations (e.g., Rignot et al.,

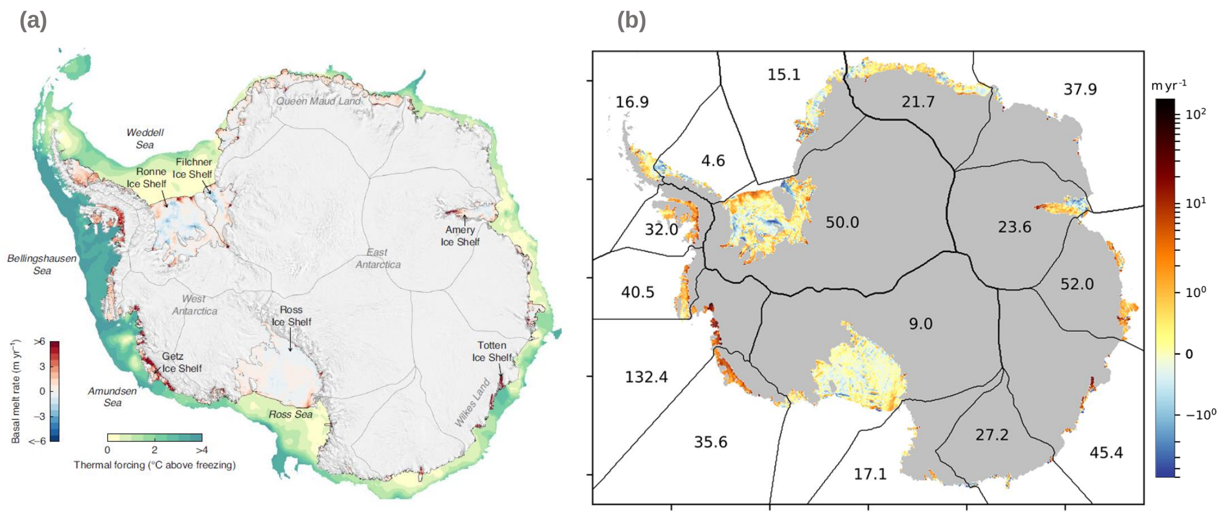

Figure 1.5: Observational estimates of ice-shelf basal melt rates for present-day from (a) Rignot et al. (2013) [Credit: Figure from Jourdain et al. (2020)] and (b) Adusumilli et al. (2020) [Credit: Figure from Adusumilli et al. (2020)].

2019; Shepherd et al., 2018) suggest that the AIS is currently thinning and losing mass at an accelerating rate along its periphery due to enhanced flow of its outlet glaciers (Figure 1.4b).

The observed acceleration of Antarctic glaciers occurs essentially in the WAIS, where the Amundsen Sea Embayment (ASE, and especially Thwaites glacier) is the region experiencing the greatest mass loss of the entire ice sheet, but also in some sectors of the East Antarctic ice sheet (EAIS) such as Wilkes Land (Rignot et al., 2019, see Figure 1.4). More specifically, mass loss has been shown to be concentrated in areas close to warm and salty ocean currents – Circumpolar Deep Water (CDW), which induce strong sub-shelf melting rates by intruding onto the continental shelf and into the cavities below the ice shelves (Paolo et al., 2015; Adusumilli et al., 2020; Rignot et al., 2019, see Figure 1.5). The resulting thinning of the floating ice shelves reduces their buttressing effect, therefore destabilising the glaciers and raising sea level by increased ice discharge (cf. section 1.2.1.2 above). While in West Antarctica, the increase in the transfer of warm CDW towards the ice shelves is likely due to localised wind-forcing perturbations modulated by tropical Pacific climate variability (such as the El Niño–Soutern Oscillation; Dutrieux et al., 2014; Jenkins et al., 2018), in East Antarctica, CDW warming at the continental slope may rather be triggered by a large scale poleward translation of mid-latitudes westerlies (Herraiz-Borreguero and Garabato, 2022).

In contrast, there is no clear continent-wide long-term trend in SMB (Shepherd et al., 2018; Rignot et al., 2019, see Figure 1.4), largely dominated by snowfall accumulation (Lenaerts et al., 2019; Mottram et al., 2021) while blowing snow sublimation is the dominant surface mass loss term, with a negative contribution of around 10 to 20% of the total SMB (Mottram et al., 2021). Ice cores suggest that on centennial timescales, SMB has increased, especially in the Antarctic Peninsula, representing a net reduction in sea level of ∼0.04 mm per decade since 1900 (Thomas et al., 2017), highlighting the fact that surface mass balance has mitigated sea-level rise over the twentieth century. Nonetheless, Kim et al. (2020) have shown that the recent ice mass loss acceleration observed around 2007 (Figure 1.4a) is associated to a decrease in precipitation, thereby underlining the importance of surface mass balance to explain Antarctic mass changes.

Finally, while all Antarctic ice shelves have been and are currently experiencing surface melting (Trusel et al., 2015), runoff has remained a relatively minor contributor to mass loss in Antarctica (Lenaerts et al., 2019), with stable surface melt rates since 1979 (Munneke et al., 2012). Ocean-induced basal melting and iceberg calving thus currently dominates ice-shelf mass losses (with about half of the ice-sheet surface mass gain lost through oceanic melting before reaching the calving front; Depoorter et al., 2013), particularly outside of the Antarctic Peninsula, where the most intense surface melt is currently observed (Trusel et al., 2015).

Overall, ocean-driven ice loss from Antarctica has thus exceeded mass gains over recent decades, and its contribution to SLR has accelerated. The largest mass imbalance is found in the WAIS, which lost more than 2000 Gt between 1992 and 2017 (Figure 1.4, corresponding to a global mean SLR of about 6 mm; Shepherd et al., 2018). In the EAIS, despite the fact that some regions have lost mass over recent decades (Figure 1.4b), recent mass balance estimates (e.g., Rignot et al., 2019; Shepherd et al., 2018; Schröder et al., 2019; Nilsson et al., 2022) tend towards equilibrium or show modest mass gains, although they are still subject to large uncertainties which obscure even in the sign of change (Stokes et al., 2022). Notably, Shepherd et al. (2018) found that the EAIS was close to balance over the last 25 years, with mass gain over the period 1992–2012 and mass loss over the period 2012–2017. On the other hand, Rignot et al. (2019) found a significant mass loss of $57 \pm 2\,\mathrm{Gt\,yr^{-1}}$ in East Antarctica over the period 1992-2017 (Figure 1.4), while Nilsson et al. (2022) found a mass gain of $73 \pm 5\,\mathrm{Gt\,yr^{-1}}$ in East Antarctica over the period 1992–2020, with nonetheless a significant decrease in the observed mass gain since the beginning of the last decade ($7 \pm 6\,\mathrm{Gt\,yr^{-1}}$ over the period 2011-2020). Regardless, according to Smith et al. (2020), the mass gains in East Antarctica are not sufficient to offset the rapid mass losses from West Antarctica and the Antarctic Peninsula, so that Antarctica's contribution to sea level change is unambiguously positive. Moreover, two of the latest efforts to reconcile EAIS mass balance estimates from different methods (Shepherd et al., 2018; Bamber et al., 2018) raise the possibility of an overall EAIS mass loss since around 2014 (Stokes et al., 2022).

1.2.3 An uncertain future: the Antarctic ice sheet in a changing climate

The IPCC 5th Assessment Report (AR5) assessed that *'warming of the climate system is unequivocal'*, and that since the 1950s, many of the observed changes, which are evident in all components of the climate system, are unprecedented over decades to millennia (Church et al., 2013; Clark et al., 2020; Golledge et al., 2019). Since then, the AR6 reported that changes to the state of the climate system have continued, with increasingly apparent impacts on natural and human systems. Many of these changes can be attributed to anthropogenic influences, and this despite efforts in international climate governance such as the Paris Agreement, which sets a long-term goal to hold the increase in global average temperature to *'well below 2°C above pre-industrial levels, and to pursue efforts to limit the temperature increase to 1.5°C above pre-industrial levels'* (IPCC, 2021). Indeed, according to Raftery et al. (2017), less than a 2°C warming by the end of the century is unlikely. To approximate the future climate of the Earth, the IPCC has defined a series of scenarios, namely the Representative Concentration Pathways (RCPs, that describe different levels of greenhouse gases emissions and other radiative forcings that might occur in the future) used in AR5 and the Shared Socio-economic Pathways (SSPs, that define a range of plausible trends in the evolution of the society over the 21st century) used in AR6.

As explained in the previous section, the AIS is currently losing mass, potentially already as a consequence of this continued and ongoing warming and climate change (Holland et al., 2019). In addition, due to the high inertia and the long response time of ice sheet systems, it is likely that the AIS still reacts to current climate changes on longer timescales and that the bulk of SLR from the AIS occurs after 2100

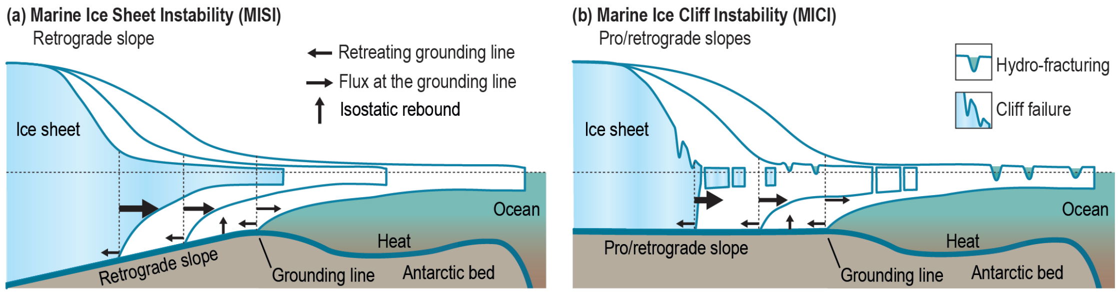

Figure 1.6: Schematic representation of Marine Ice Sheet Instability (MISI, a) and Marine Ice Cliff Instability (MICI, b) from Pattyn (2018). [Credit: SROCC, Oppenheimer et al. (2019)]

(Church et al., 2013). Therefore, numerous studies try to investigate how the AIS will behave under different climate scenarios. Unfortunately, despite considerable advances in physically-based models of ice dynamics over the last decade (Pattyn et al., 2017, cf. section 1.3), there are still large uncertainties in the projections of future sea-level rise from the Antarctic ice sheet, so that of all the components of the global sea-level budget, the contribution of Antarctica is the most uncertain (Kopp et al., 2017; Edwards et al., 2021). Indeed, ice-sheet mass changes are associated with several positive feedback mechanisms. Moreover, the Antarctic ice sheet, and more particularly its marine glaciers, is capable of abrupt and nonlinear changes, often referred to as tipping points. The interactions between the ice sheet and its surrounding environment (namely the ocean, the atmosphere and the solid Earth) have been shown to strongly influence the stability of the ice sheet, more particularly by triggering or dampening the instabilities threatening the ice sheet. In the coming section, we provide an overview of the key processes that will drive the response of the Antarctic ice sheet under climate change and thus contribute to the uncertainties in future Antarctic mass changes.

1.2.3.1 The contributors to the uncertainty

1.2.3.1.1 The instability mechanisms in marine ice sheets

Over 90% of the AIS coastline is characterised by marine margins, which essentially consist of fast-flowing outlet glaciers or ice shelves (Bindschadler et al., 2011, Figure 1.2). In such marine-terminating portions of the ice sheet, the ice rests on land that is below sea level and most of the mass loss occurs following the flow of grounded ice across the grounding line into the ocean. This is the case in the majority of the WAIS (with maximum depths of more than 2000 m below present sea level, see Figure 1.2 – defining it as a ‘marine ice sheet’), and would remain so even if the ice was removed and the underlying lithosphere was fully rebounded (Paxman et al., 2022). On the other hand, much of the (larger) EAIS lies on bedrock above sea level, with the exceptions of some large marine basins: the Wilkes, Aurora and Recovery basins (Morlighem et al., 2020). As the ice flux across the grounding line increases sharply with the thickness of ice there (Weertman, 1974; Schoof, 2007; Thomas and Bentley, 1978), such marine portions of ice sheets are susceptible, under some conditions, to dynamical instabilities, proposed via the mechanisms of Marine Ice Sheet Instability (MISI – Figure 1.6a) and Marine Ice Cliff Instability (MICI – Figure 1.6b). Therefore, a major uncertainty in future Antarctic mass losses is the possibility of rapid and/or irreversible ice losses through instability of its marine portions. Since they represent extensive regions of the ice sheet (Figure 1.2), the AIS has the potential to make a significant contribution to future sea-level rise (Golledge et al., 2015; Ritz et al., 2015; Bulthuis et al., 2019; Edwards et al., 2019).

The marine ice sheet instability

A first condition under which a marine ice sheet is at risk of dynamical instabilities is if it lies on a bedrock that deepens from the periphery of the ice sheet towards the interior of the continent (a so-called *retrograde bed slope*). Indeed, in such a configuration, a grounding-line retreat (from thinning and eventually flotation of the ice near the grounding line) into deeper water (associated to thicker ice), leads to increased ice flux into the ocean and therefore further thinning, inducing a positive feedback that leads to runaway ice loss (Figure 1.6a). The position of the grounding line is thus inherently unstable, in the absence of ice-shelf buttressing[1], until reaching an upward sloping bed or, in the worst case scenario, until its complete disappearance. This self-reinforcing mechanism is the so-called marine ice sheet instability (MISI), which was first presented by Mercer (1978). MISI is supported by theoretical (Weertman, 1974; Thomas and Bentley, 1978; Mercer, 1978; Schoof, 2007; Thomas, 1979) and numerical (Pattyn et al., 2013; Cornford et al., 2020; Pollard and DeConto, 2009; Clark et al., 2020; Sun et al., 2020) results, and also by some paleo-records which support past high retreat rates (Dowdeswell et al., 2020; Wilson et al., 2018; Blackburn et al., 2020; Graham et al., 2022), though they cannot unambiguously distinguish progressive from unstable retreat.

Marine portions of the AIS grounded on an inward sloping bed may be observed in the vast majority of the WAIS (and especially in the ASE) as well as in some basins of the EAIS, such as the Wilkes and Aurora basins (Fretwell et al., 2013; Morlighem et al., 2020, see Figure 1.2). More specifically, a recent update of bed topography has revealed large and overdeepened subglacial troughs in East Antarctica (Morlighem et al., 2020). These basins are therefore particularly vulnerable to rapid grounding-line retreat that may lead to MISI in case of weak or absence of buttressing. Indeed, it is important to underline that the MISI theory remains only valid for unconfined ice shelves, i.e., ice shelves that do not exert a force on the inland ice sheet other than the (ocean) water pressure (Schoof, 2007). Stable grounding line positions can be reached on retrograde sloping beds if ice shelves provide enough buttressing (Sergienko and Wingham, 2019; Cornford et al., 2020).

The possibility that some glaciers, such as Pine Island Glacier and Thwaites Glacier (which drain a large portion of West Antarctica and where the majority of current mass loss occurs – Figure 1.4) in the ASE are already undergoing MISI has been suggested by both satellite observations and numerical simulations using state-of-the-art ice-sheet models (Joughin et al., 2014; Rignot et al., 2014). More specifically, Pine Island Glacier underwent a sustained retreat up to 2010, triggered by oceanic forcing (warm circumpolar deep water breaching the continental shelf). After that time, it started to slow down, probably due to a combination of reduced forcing (Dutrieux et al., 2014) and a concomitant increase in ice-shelf buttressing (Favier et al., 2014). Unlike Pine Island Glacier, Thwaites Glacier is currently in a less-buttressed state because its ice shelf is mostly unconfined. Observations have shown that its grounding line retreats along clear retrograde pathways (Rignot et al., 2014), with little opportunity for stabilisation from ice-shelf buttressing (Parizek et al., 2013). Moreover, several simulations using state-of-the-art ice-sheet models indicate continued mass loss and possibly MISI even under present climatic conditions (Joughin et al., 2014; Seroussi et al., 2017). More recently, Milillo et al. (2022) have highlighted high retreat rates (faster than anticipated by numerical models) for the Pope, Smith, and Kohler glaciers, also located in the ASE. It is important to note, however, that whether these behaviours may really be attributed to MISI mechanisms has been questioned by others (Gudmundsson et al., 2019; Haseloff and Sergienko, 2018). These marine

[1] It is to be noted, however, that variations in basal properties have been shown to create, under certain circumstances, stable unbuttressed ice-sheet grounding lines on retrograde sloping beds (Brondex et al., 2017).

glaciers in West Antarctica, if subject to MISI, have the potential to destabilise adjacent basins and ultimately trigger a dramatic collapse of the West Antarctic ice sheet (Sun et al., 2020). In East Antarctica, most of the retrograde marine basins (except for some glaciers in the Aurora Basin) show limited evidence of change at present (Gardner et al., 2018; Rignot et al., 2019; Stokes et al., 2022, Figure 1.4b) and seem more stable than marine basins in West Antarctica due, in particular, to present-day grounding lines lying on shallower and narrower sills, providing greater ice-shelf buttressing (Morlighem et al., 2020). Current understanding is thus insufficient to determine whether and when thresholds of instability may be reached in East Antarctica's marine based sectors (Stokes et al., 2022).

Overall, evidence to identify the operation of instability mechanisms such as MISI in paleo-ice sheet retreat are limited, and the conditions necessary to initiate such a collapse (such as the speed and intensity of these changes) remain poorly known and understood.

The marine ice cliff instability

The Antarctic ice shelves terminate (at their calving front) in near vertical ice cliffs submerged in the ocean. For the calving fronts of thick glaciers to remain stable, they need to be grounded in deep water (Bassis and Walker, 2012). If not, they offer a second condition under which a marine ice sheet is at risk of dynamical instability. Indeed, thicker ice or shallower water increases the height of the ice cliff exposed above water, and where this exceeds the threshold for structural stability ($\sim$100 m), the cliff may collapse. Based on this principle, Pollard et al. (2015) proposed the 'Marine Ice Cliff Instability' (MICI) hypothesis, which describes rapid, unmitigated calving triggered by the collapse of tall ice cliffs under their own weight (a process called cliff collapse). In regions of ice sheets grounded deep below sea level, where the ice is 100 m or more above the ocean surface, cliff collapse may lead to a self-sustaining retreat of the ice front. Indeed, these tall ice cliffs at the ice-ocean boundary are structurally unstable, and their collapse could leave behind further tall cliffs, resulting in sustained ice losses (Figure 1.6b). MICI requires an a priori collapse of ice shelves, and is thus favored by hydrofracturing through the increase of water pressure in surface crevasses (Pollard et al., 2015; Bassis and Walker, 2012, see section 1.2.3.1.2).

Contrary to MISI, MICI could also occur on prograde bedrock slopes, but is more pronounced on retrograde slopes, as retreat (by calving) exposes thicker ice upstream. Therefore, when glaciers retreat into a marine basin with a retrograde slope more than 1,000 m deep (as may be observed in the West Antarctic ice sheet), both MISI and MICI may be superimposed (Pattyn et al., 2017; DeConto and Pollard, 2016). As a consequence, MICI could lead to sea-level contributions beyond 2100 considerably higher than the likely range projected by other models, due to catastrophic retreat of sections of West Antarctica on decadal-to-century time scales (DeConto and Pollard, 2016; DeConto et al., 2021).

Nevertheless, the current application of MICI in ice-sheet models (using quasi-empirical parameterisations of retreat rates as a function of cliff height; Pollard et al., 2015; DeConto and Pollard, 2016; DeConto et al., 2021; Crawford et al., 2021) is still debated in the scientific community. Indeed, observational evidence for MICI is indirect: although there is evidence supporting MICI in the paleo-record (Wise et al., 2017), it has yet to be observed in modern day glaciers. Only Crane Glacier on the Antarctic Peninsula has shown retreat consistent with MICI, after the Larsen B ice shelf collapsed (Oppenheimer et al., 2019), and MICI-style behaviour at Jakobshavn and Helheim glaciers in Greenland (Joughin et al., 2014) might not be representative of wider Antarctic glaciers. Observations from Greenland have also shown that steep cliffs commonly evolve into short floating extensions, rather than collapsing catastrophically (Joughin et al., 2020). In addition, new theoretical evidence nuance the MICI theory by suggesting that ice cliff collapse may only occur after very rapid ice shelf disintegration (Clerc et al., 2019), and that the subsequent rate of retreat depends on the terminus geometry (Bassis and Ultee, 2019). Bassis et al. (2021) also demonstrated that this kind of collapse can be slowed or even completely prevented either by upstream thinning of the ice sheet or by the resistive forces provided by sea ice and calved debris. Therefore, marine-terminating parts of Antarctica may be less vulnerable to rapid and irreversible collapse than suggested by DeConto and Pollard (2016); DeConto et al. (2021). Finally, while MICI was initially proposed by Pollard et al. (2015) as an additional physical process required to trigger a significant retreat of marine basins in East Antarctica in order to reproduce the high eustatic sea levels (5 to more than 20m above present) during the Pliocene epoch, Edwards et al. (2019) revealed that MICI was not required to reproduce high eustatic sea levels in the past.

Given the low agreement on the exact MICI mechanism and limited evidence of its occurrence in the present or the past, its potential to affect future sea level rise is very uncertain, and widespread, sustained and very rapid ice loss from Antarctica this century through MICI seems unlikely. Nonetheless, this poorly understood process has the potential to strongly increase Antarctic mass loss under high-emissions scenarios on century to multi century timescales (DeConto and Pollard, 2016; DeConto et al., 2021; Bassis et al., 2021; Fox-Kemper et al., 2021).

The risk of crossing irreversible tipping points

Therefore, the future SLR contribution from the AIS depends on the behaviour of its marine glacier systems and whether they enter unstable mechanisms (Pattyn and Morlighem, 2020). Overall, these instabilities are essentially triggered by a loss of ice-shelf buttressing. In turn, ice shelves are directly affected by oceanic and atmospheric conditions (see sections 1.2.3.1.2 and 1.2.3.1.3 below). This raises concerns about the existence of a tipping point or critical threshold for the WAIS (Lenton et al., 2008; McKay et al., 2022) and the AIS more generally, beyond which the ice sheet can undergo rapid, abrupt, and irreversible changes. Indeed, when such self-sustained retreat is triggered, grounding-line retreat may continue even with no additional forcing (Seroussi et al., 2017; Joughin et al., 2014). Pattyn et al. (2018) have suggested that a key threshold for the stability of ice shelves, and thus the stability of the ice sheet, lies between 1.5 and 2°C of mean annual air temperature above present. The existence of a tipping point, on the other hand, implies that ice-sheet changes are potentially irreversible. The AR6 states that mass loss from the AIS is irreversible on decadal to millennial timescales (Fox-Kemper et al., 2021) and that returning to a preindustrial climate may not be sufficient to prevent or reverse substantial Antarctic mass losses (hence giving rise to a hysteresis behaviour; Garbe et al., 2020). Reversibility, however, may be possible over large climate cycles, such as a glacial-interglacial cycle (Pattyn and Morlighem, 2020). More recently, results from Urruty et al. (2022); Reese et al. (2022) suggest that there is no indication of current irreversible or self-sustaining retreat, suggesting that present-day grounding-line retreat is driven by external climate forcing alone (Urruty et al., 2022), as had already been suggested by Gudmundsson et al. (2019), but that they may evolve towards such a retreat under current climate conditions (Reese et al., 2022). For the Antarctic ice sheet, tipping points are known to exist for warming levels that could be reached before the end of this century (DeConto and Pollard, 2016; DeConto et al., 2021; Golledge et al., 2019; Chambers et al., 2021; McKay et al., 2022).

1.2.3.1.2 Ice–ocean interactions

Basal melt is the result of a positive thermal forcing, i.e., water warmer than the local freezing point (which depends on ocean salinity and water pressure) getting in contact with the ice shelf. Therefore, the magnitude of sub-shelf melting is essentially controlled by the strength of the heat flux transported into the sub-ice shelf cavities, which in turn is regulated by the transfer of water masses onto the continental shelf. The latter is locally affected, among others, by coastal upwelling, sea-ice formation and internal ocean dynamics. In contrast, iceberg calving is largely dominated by the ice dynamics.

Future ocean conditions will, therefore, exert a critical influence on ice discharge by means of basal melting and its influence on ice-shelf buttressing (Depoorter et al., 2013; Rignot et al., 2013; Paolo et al., 2015; Gudmundsson et al., 2019, see section 1.2.1.2). However, even for present-day, a good understanding of the drivers of ocean trends remain limited, essentially due to the sparsity of observations, and this especially around East Antarctica (Schmidtko et al., 2014; Stokes et al., 2022). Ocean-driven mass loss has been observed in both the WAIS and the EAIS over the past decades (section 1.4), driven by enhanced delivery of ocean heat to the ice shelves (Jenkins et al., 2018; Holland et al., 2019; Rintoul et al., 2016). In West Antarctica, such increase in ocean heat flux is likely driven by localised wind-forced perturbation along the continental slope, which facilitates the penetration of warm CDW on the continental shelf (Jenkins et al., 2018). This wind-forced perturbation seems to be modulated by tropical Pacific climate variability (Dutrieux et al., 2014; Paolo et al., 2018), which has lately been shown to be linked to anthropogenic forcing, with persistent wind anomalies under strong future greenhouse gas scenarios (Holland et al., 2019). In contrast, a recent study suggests that the warm CDW detected close to several main outlet glaciers from the EAIS may be associated to an interdecadal poleward shift of the westerlies over the Southern ocean, which is predicted to persist into the 21st century (Herraiz-Borreguero and Garabato, 2022). In addition, changes in sea ice formation and extent are projected to drive an increase in basal melting of the Antarctic ice shelves by allowing an increased penetration of warmer modified CDW and Antarctic surface water (ASW) into the ice shelf cavities (Naughten et al., 2018).

Basal melting of the Antarctic ice shelves is thus expected to increase in a warming climate (Timmermann and Hellmer, 2013; Naughten et al., 2018; Little and Urban, 2016). More specifically, using a Sea Ice/Ice-Shelf ocean model forced with atmospheric outputs from models from phase 5 of the Coupled Model Intercomparison Project (CMIP5), Naughten et al. (2018) project an increase of the total Antarctic ice shelves mass loss between 41% and 129% over the 21st century. While an increase in basal melting is projected to occur in all sectors of the Antarctic ice sheet, it is the Amundsen Sea sector (followed by the Bellingshausen Sea sector) which is expected to experience the largest augmentation, and this due to an increase in warm CDW penetration on the continental shelf (Naughten et al., 2018). In contrast, the Ross ice shelf (which, as shown in Figure 1.5, currently experiences low sub-shelf melt rates due to the intrusion of cold and dense shelf water into ice-shelf cavities, similar to other large ice shelves; Adusumilli et al., 2020; Schmidtko et al., 2014) is expected to experience the lowest increase in sub-shelf melting, primarily driven by atmospheric warming (Naughten et al., 2018). Contrastingly, Hellmer et al. (2012, 2017) have shown that such current cold ice-shelf cavities may become more similar to the ASE in the future, with derivatives of CDW penetrating in the ice shelf cavities under the Filchner-Ronne ice shelf, dramatically increasing local basal melting. Overall, projections of future Antarctic ocean conditions remain uncertain, and this essentially because atmosphere–ocean general circulation models do not resolve important processes, such as the circulation within the cavities below the ice shelves, which has been highlighted above as of crucial importance to determine sub-shelf melting patterns and magnitude. Only few coupled ice sheet-ocean simulations, enabling to account for the strong two-way interactions between sub-ice shelf circulations and the shape of the ice-shelf base (Fyke et al., 2018), exist (e.g., Seroussi et al., 2017; Kreuzer et al., 2021; Pelletier et al., 2022). For these reasons, there is little consensus regarding future change in CDW, and more generally low confidence in future change in the temperature of Antarctic ice shelf cavities (Fox-Kemper et al., 2021).

In turn, sub-shelf melting in itself is likely to modify ocean conditions via the input of freshwater in the ocean. Recent studies suggest that the input of freshwater from ice shelves in the ocean may create a positive feedback by stratifying the water column and trap warm water below sea surface, thus resulting in ocean warming and increased melt rates near the grounding line around most Antarctic margins (Silvano et al., 2018; Bronselaer et al., 2018; Golledge et al., 2019; Sadai et al., 2020). This mechanism, which is not included in most climate models projections, may result in a positive feedback that could increase the mass loss from the Antarctic ice sheet (Golledge et al., 2019). Strong uncertainties thus remain in determining how global warming relates to ocean dynamics and hydrography, potentially increasing sub-shelf melt (Pattyn and Morlighem, 2020).

1.2.3.1.3 Ice–atmosphere interactions

Projections of future Antarctic SMB are uncertain due to an increase in competing processes in a warming climate. First, snowfall is projected to increase under atmospheric warming due to increased atmospheric saturation water vapour pressure, the availability of more open coastal water and changing cloud properties (Lenaerts et al., 2019). More specifically, analyses of CMIP models have found Antarctic temperature sensitivity for accumulation (precipitation minus sublimation) of approximately 5 ± 1% per °C (Frieler et al., 2015), and for precipitation of around 4 to 9% per °C (Bracegirdle et al., 2020). Therefore, as snowfall is the main component of AIS present-day SMB (see section 1.4), we may expect an increase of SMB in a warming climate (Lenaerts et al., 2019). According to CMIP5 and CMIP6 models, by the end of the century, Antarctic SMB will be dominated by snowfall (Payne et al., 2021; Kittel et al., 2021), with an increase in SMB of 6.0 to 9.9% per °C (Previdi and Polvani, 2016), thereby mitigating future sea-level rise. Nonetheless, under high-end emission scenarios, Kittel et al. (2021) show that the increase in total Antarctic SMB by the end of the century will, in fact, be characterised by strong local differences, with an increase in SMB over grounded ice, but a decrease in SMB over the ice shelves.

Indeed, warming of the atmosphere also promotes rainfall and melting at the ice sheet surface. This is particularly illustrated by an ice core on the northeast Antarctic Peninsula which documents rapid melt intensification since the mid-twentieth century (Abram et al., 2013). In the future, according to global and regional atmospheric modelling, Antarctic surface melting and runoff is projected to increase with warming (Trusel et al., 2015; Kittel et al., 2021), and this approximately twofold by 2050, independent of the RCP forcing scenario (Trusel et al., 2015). However, the relationship between temperature and melting is highly non-linear (mainly as a result of the melt–albedo positive feedback; Zeitz et al., 2021). Therefore, while the presence of significant rainfall and surface runoff is unlikely to spread far south of the Antarctic Peninsula by 2100 under low-emissions scenarios (Trusel et al., 2015), melt rates associated with different future scenarios diverge considerably around mid-century, resulting in a wide range of values for 2100 (Trusel et al., 2015).

According to Bell et al. (2018), the projected increase in surface melt could impact the mass balance of the AIS by three primary modes:

(i) First, increased surface melting leads to direct runoff and ice-sheet thinning, thus counterbalancing the increase in snow accumulation mentioned above. Over the ice shelves, Kittel et al. (2021)

suggest that surface runoff anomalies may exceed precipitation anomalies (causing SMB to become negative, hence potentially weakening the ice shelves and their buttressing effect) with a regional warming of +2°C.

(ii) Second, increased surface melt may lead to the injection of melt water to underlying bedrock and the suglacial hydrological system, potentially increasing ice mass loss through enhanced basal sliding and triggering a fundamental shift in the dynamics and mass balance of the AIS. However, it is important to note that this mode of impact has not been documented in Antarctica yet, though it is widespread in Greenland.

(iii) Finally, the presence of liquid water at the ice sheet surface causes hydrofracturing as water propagates into crevasses, widening and deepening them (Scambos et al., 2000; Kuipers Munneke et al., 2014). On the ice shelves, this could lead to significant weakening of the ice so that they become more vulnerable to calving (Bassis and Walker, 2012). More drastically, this may eventually lead to ice shelf collapse if sufficient melt water and cracks are available, as has been observed in the Antarctic Peninsula at the beginning of the century (Scambos et al., 2004). Such link between surface melting and ice shelf collapse was also showed by the the ice core (mentioned above) documenting rapid melt intensification since the mid-twentieth century coincident with numerous ice shelf collapses (Abram et al., 2013). In addition, hydrofracturing is considered a precursor for MICI (section 1.2.3.1.1) if it leads to the production of high, mechanically unsustainable ice cliffs (Bassis and Walker, 2012; Pollard et al., 2015). Lai et al. (2020) have shown that 60±10% of ice shelves (by area) both buttress upstream ice and are vulnerable to hydrofracturing if inundated with water. Nonetheless, most studies identify significant potential ice-shelf collapse by 2100 under only the unmitigated scenarios (Trusel et al., 2015; DeConto and Pollard, 2016).

Overall, when and where each mode of melt water impact are activated in a wetter, warmer Antarctica will to some extent control the future contribution of the Antarctic ice sheet to sea-level rise (Bell et al., 2018). Anyway, warmer futures are associated with more intense and extensive runoff that increases the risk of hydrofracture-induced destabilisation, especially if concentrated in areas that also provide buttressing. On the grounded portions of East and West Antarctica, surface lowering due to runoff and connectivity to the bed could become significant by 2100 in certain regions (Bell et al., 2018). For these reasons, the AR6 states that there is only medium confidence that the future contribution of Antarctic SMB to sea level this century will be negative under all greenhouse gas emissions scenarios (Fox-Kemper et al., 2021). However, refreezing of melt water in ice shelves (Kuipers Munneke et al., 2014) is expected to delay or prevent melt water runoff (Hubbard et al., 2016). In addition, studies show that melt water in Antarctica can be displaced laterally in flow networks (Kingslake et al., 2017), and sometimes even drain into the ocean, which might prevent melt water from destabilising the ice shelves (Bell et al., 2017).

Finally, it is important to note that feedbacks can arise between the ice sheet and the atmosphere as a consequence of ice-sheet geometry changes (Fyke et al., 2018). Notably, a well-known of such feedback is the temperature-based SMB–elevation feedback (Oerlemans, 1981; Edwards et al., 2014), which is associated with the changes in atmospheric temperature with changing elevation of the ice sheet surface (i.e., lapse rates). The higher (lower) the ice sheet becomes, the less (more) mass will be accumulated at the surface due to less (more) precipitation in greater altitude, and the less (more) surface melt will be triggered due to lower (higher) air temperatures. In addition to the relatively straightforward temperature effects, changing ice-sheet topography will also drive regional feedbacks as large-scale topographic ice-sheet changes impact local atmospheric dynamics (such as katabatic wind strength), which in turn modify the atmospheric circulation and associated ice-sheet accumulation and melting patterns.

1.2.3.1.4 A stabilising aspect: interactions with the solid Earth and the local sea level via glacial isostatic adjustment

Glacial Isostatic Adjustment

Isostatic adjustment is the process of adjustment of the Earth lithosphere (the crust and the rigid upper part of the mantle) towards a hydrostatic equilibrium in which it is regarded as floating on the asthenosphere, i.e., the underlying viscous mantle (Gregory et al., 2019). Therefore, isostatic adjustment occurs in response to changes in the mass load of the lithosphere, typically over multi-millennial timescales determined by the viscous flow of the mantle beneath the lithosphere. In addition to this viscous response, as the solid Earth is a viscoelastic material, an elastic response of the lithosphere occurs almost instantaneously, on annual timescales, in response to changes in load. Although it is smaller in magnitude compared with the potential viscous isostatic response, it is much more rapid, and hence responsible for significant vertical land movement due to any recent historical changes in land ice. At present this concerns, for example, West Antarctica (Larour et al., 2019), where the majority of current Antarctic mass changes occur (Shepherd et al., 2018; Rignot et al., 2019).

The shape of the geoid, on the other hand, depends on the geographical distribution of mass over the Earth. Therefore, changes in the distribution of surface mass (e.g., the decrease in the mass of an ice sheet and the resulting addition of mass to the ocean) induce changes in the shape of the geoid, and therefore of the sea surface. If an ice sheet melts, the net volume of water in the oceans increases, but the gravitational force exerted by the (now smaller) ice sheet on the ocean decreases. This leads to a redistribution of water from the near field of the ice sheet to the far field. Typically, such a redistribution of the water leads to a sea-level fall within 2000 km of the ice sheet, while, in the far field, the migration adds to the general increase in ocean volume, leading to a sea-level rise in excess of the global mean sea level (Mitrovica et al., 2011). In addition, the shrinking of the ice sheet and the transferal of water to the ocean causes the solid Earth to deform due to isostatic adjustment as mentioned above, and this redistribution of mass inside the Earth further alters the shape of the geoid. Finally, alteration of the distribution of mass throughout the Earth system perturbs the magnitude and direction of the Earth rotation vector (Wu and Peltier, 1984), which will, in turn, instantaneously alter the shape of the geoid (Milne and Mitrovica, 1998) and cause elastic as well as, over longer timescales, viscous deformation of the solid Earth (Han and Wahr, 1989), thus further altering the shape of the geoid. Whatever the cause, redistribution of the ocean mass itself has gravitational effects, and thereby the ocean affects its own mass distribution. These effects of Earth **G**ravity, Earth **R**otation and viscoelastic solid Earth **D**eformation (GRD) resulting from the redistribution of mass between terrestrial ice and water reservoirs and the ocean occur simultaneously and thus interact through numerous feedbacks (Whitehouse, 2018). In this framework, Glacial Isostatic Adjustment (GIA) is ongoing GRD in response to past and current changes in the distribution of ice and water at the Earth's surface. Since GIA affects both the geoid (considered here equivalent to the sea surface) and the height of the sea floor topography through vertical land movement, it induces relative sea-level changes (Mitrovica et al., 2001), where the relative sea level (RSL) is the difference between the sea surface and the sea floor (Gregory et al., 2019).

Feedbacks between GIA processes and ice dynamics

GIA processes influence ice-sheet mass changes through different feedback mechanisms. First, GIA-induced changes of the relative sea level along the ice-sheet margin affect the local water depth and hence the position of the grounding line (Whitehouse et al., 2019). Since the thickness of ice at the grounding line, in turn, controls the amount of ice flowing across it (and thereby ice mass changes and grounding-line movements; Schoof, 2007; Weertman, 1974), GIA has the potential to stabilise a marine ice sheet undergoing MISI (Gomez et al., 2012, 2013, 2015; Konrad et al., 2015; Gomez et al., 2018; Larour et al., 2019), and could even lead to a re-advance of the grounding line in some areas (Kingslake et al., 2018). Indeed, grounding-line retreat leads to a decrease of the local water depth through both the bedrock uplift and the sea-surface fall associated with the ice mass loss. The grounding-line retreat may thus slow down as the height above hydrostatic equilibrium increases inland (see Figure 1.7). In addition, relative sea-level changes due to GIA can influence the degree to which ice shelves are able to stabilise the ice sheet. For example, a local decrease in water depth may enhance grounding of the ice shelf at ice rises, thereby stabilising the ice sheet, while an increase in water depth can lead to ungrounding of an ice rise, thus enhancing the ice flow across the grounding line (Matsuoka et al., 2015). GIA can also affect the ice dynamics by altering the shape and slope of the bed near the margins of the ice sheet, where ice mass loss is occurring. For example, the gradient of a reverse bed slope may be reduced by differential solid-Earth rebound (Adhikari et al., 2014). Finally, GIA processes can influence ice dynamics through the feedback between isostatically-driven ice surface elevation change and surface mass balance (van den Berg et al., 2008, section 1.2.3.1.3) as well as changes in the surface gradients and therefore the dynamic state of the ice sheet (Konrad et al., 2014).

The strength of GIA feedbacks depends on the pattern and the rate at which the solid Earth responds to ice-sheet changes. Both depend in turn on the rheological properties of the solid Earth, in particular the lithosphere thickness and the upper mantle viscosity, respectively. Several studies have shown that the AIS lies on a region of the solid Earth that is characterised by a strong lateral variability in rheological properties, with a thin lithosphere and a low-viscosity upper mantle beneath West Antarctica and a thick lithosphere and a more viscous upper mantle beneath East Antarctica (Ritzwoller et al., 2001; Morelli and Danesi, 2004; Heeszel et al., 2016; Chen et al., 2017; Pappa et al., 2019; Lloyd et al., 2020; Swain and Kirby, 2021). A low-viscosity upper mantle and a thin lithosphere (referred to as a weak Earth structure), as observed under the WAIS, will produce a faster and more localised viscoelastic response of the solid Earth to ice-load changes (contrary to a thicker lithosphere that acts to dampen and smooth the solid-Earth response or a high-viscosity upper mantle that generates a slower response), hence emphasizing the local relative sea-level fall and facilitating stabilising feedbacks (Gomez et al., 2015; Konrad et al., 2016). More specifically, recent evidence suggests very low mantle viscosities in some areas of West Antarctica, inducing solid-Earth response times on decadal rather than millennial timescales (Nield et al., 2014; Barletta et al., 2018), which is orders of magnitude faster than previously assumed. The West–East dichotomy in Antarctic Earth structure may play a crucial role in the future evolution of the AIS (Kaufmann et al., 2005; van der Wal et al., 2015; Hay et al., 2017; Nield et al., 2018), especially since considering that rapid uplift could occur in active areas of Antarctica such as Thwaites Glacier (Barletta et al., 2018). This could lead to a significant slowdown in Antarctic mass loss over the next centuries (Gomez et al., 2015). However, although the physics of GIA is well understood, major uncertainties remain in determining rheological properties of the Antarctic solid Earth with precision and absolute values of mantle viscosity and lithosphere thickness remain poorly constrained (van der Wal et al., 2015; Hay et al., 2017; Gomez et al., 2018; Whitehouse et al., 2019).

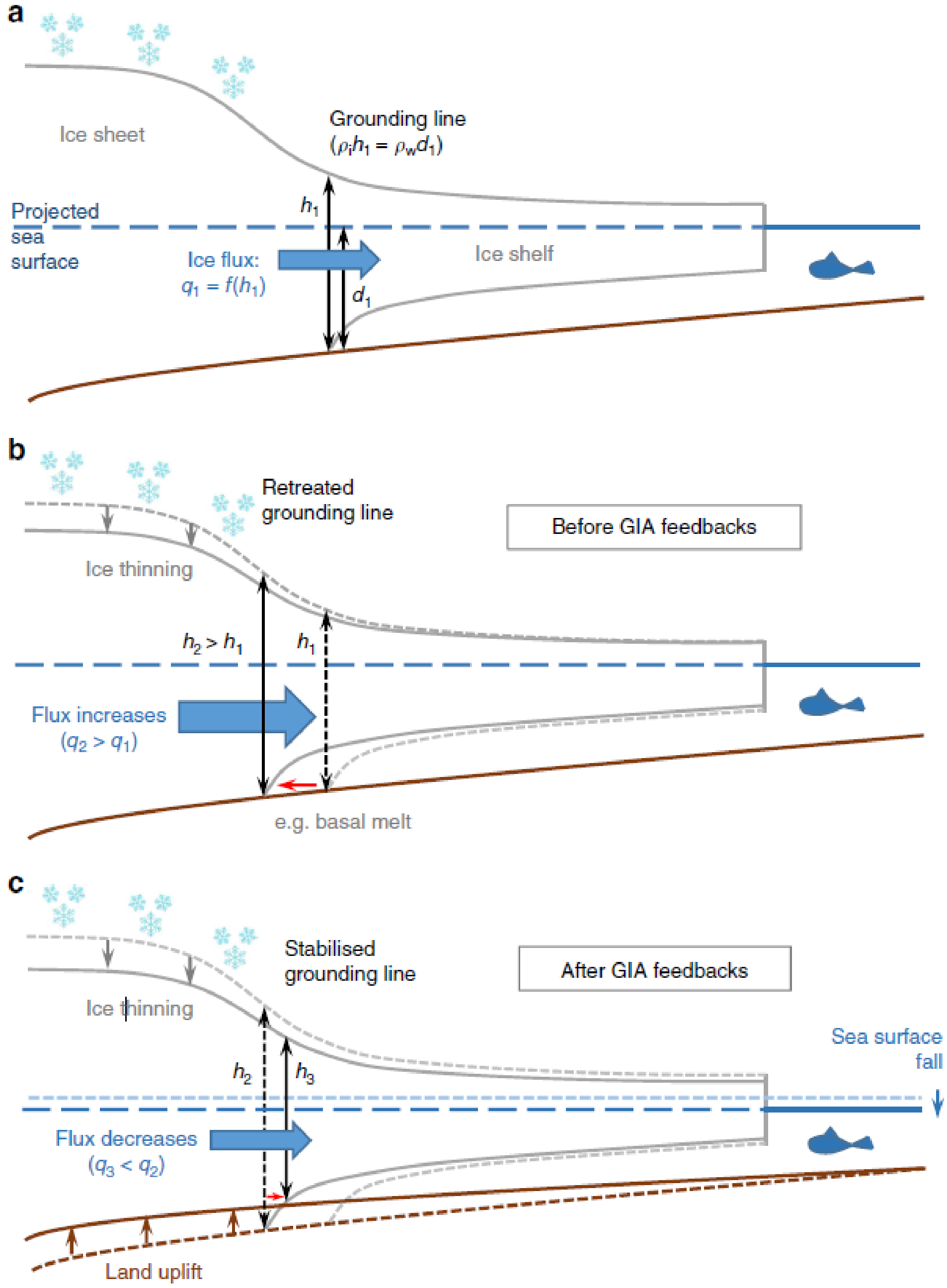

Figure 1.7: **Stabilising effect of GIA on ice dynamics.** (a) Flux across the grounding line (q_1) depends on ice thickness there (h_1). The position of the grounding line also depends on the local water depth (d_1), the density of ice (ρ_i), and the density of seawater (ρ_w). (b) Before GIA feedbacks: the grounding line retreats into deeper water on a downward sloping bed (red arrow). Increase in ice thickness at the grounding line ($h_2 > h_1$) results in an increase in ice flux towards the ocean ($q_2 > q_1$), triggering further grounding line retreat. (c) After GIA feedbacks: grounded ice loss triggers land uplift and sea surface lowering. The resulting decrease in water depth and hence ice thickness ($h_3 < h_2$) at the grounding line results in a decrease in ice flux across the grounding line ($q_3 < q_2$), which acts to stabilise the grounding line position. Initial configurations shown are by the dashed lines, and new configurations are shown by solid lines [Figure Credit: Whitehouse et al. (2019)].

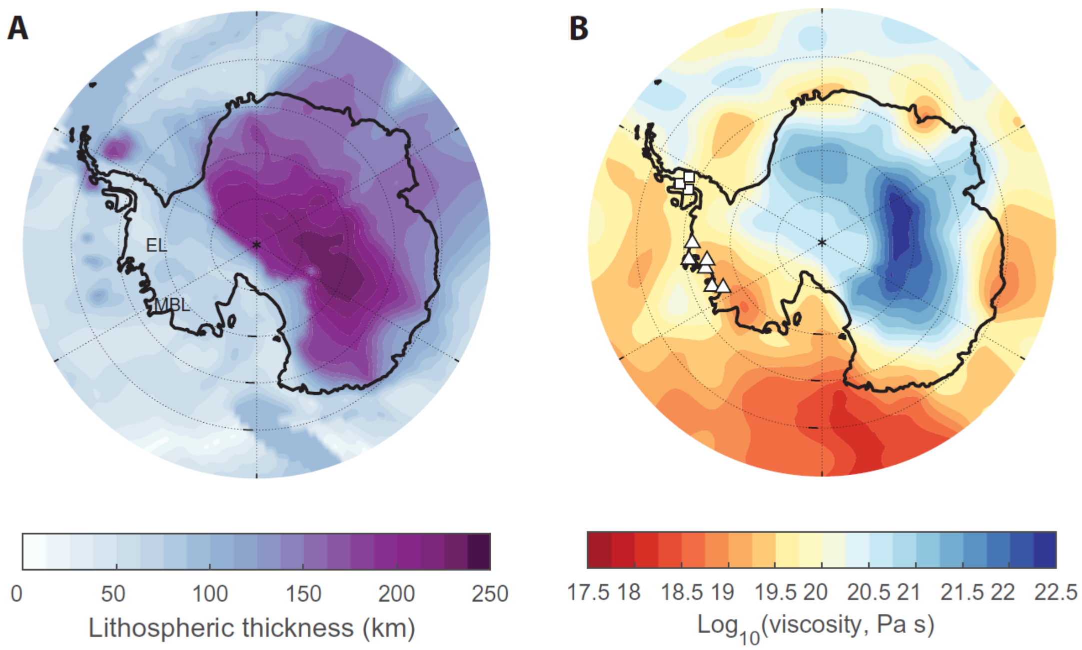

Figure 1.8: Viscoelastic properties of the Antarctic ice sheet. (A) Elastic lithospheric thickness and (B) mean viscosity from the base of the lithosphere to 400-km depth across Antarctica and the Southern Ocean for the model $V3D_{SD}$ used in Pan et al. (2021) [Credit: Figure from Pan et al. (2021)].

Overall, GIA feedbacks are not expected to substantially slow sea-level rise from marine-based ice in Antarctica over the 21st century (Larour et al., 2019; DeConto et al., 2021), but these processes could become important on multi-century and longer time scales (Gomez et al., 2015). However, new projections for Pine Island Glacier (Kachuck et al., 2020) support previous work (Barletta et al., 2018) suggesting that the lower mantle viscosity in the ASE may lead to a negative feedback on decadal time scales, therefore pointing to a potential stabilisation of grounding line retreat in this area over the 21st century already.

1.2.3.1.5 Ice sheet processes: uncertainties in substrate conditions and ice rheology

Major uncertainties in the response of ice sheets to environmental forcing are due to subglacial processes, such as the underlying bedrock and substrate conditions, basal friction, and ice rheology, which are difficult to observe. Especially, the mechanics of basal friction and how it varies spatially remain largely unknown (Pattyn and Morlighem, 2020). Lowry et al. (2021) have addressed the influence of uncertainties in substrate conditions and ice rheology, which impact ice flow and sliding. Their results show that these uncertainties are especially important over decadal timescales in regions with retrograde bed slopes prone to MISI, such as the Amundsen Sea Embayement and the Aurora Basin. In addition, Kazmierczak et al. (2022) have shown that subglacial hydrology modulates the basal sliding response of the AIS to climate forcing. Especially, they suggest that subglacial hydrological models that lead to low values of effective pressure (i.e., the ice overburden pressure minus the subglacial water pressure) in the grounding zone increase the ice-sheet sensitivity to climate changes (Kazmierczak et al., 2022). In addition, they show that the modulation of subglacial hydrology is a function of the power in a Weertman-type sliding law, where more plastic sliding laws significantly increase the sensitivity of the ice sheet (as already suggested elsewhere; Brondex et al., 2019; Sun et al., 2020; Bulthuis et al., 2019).

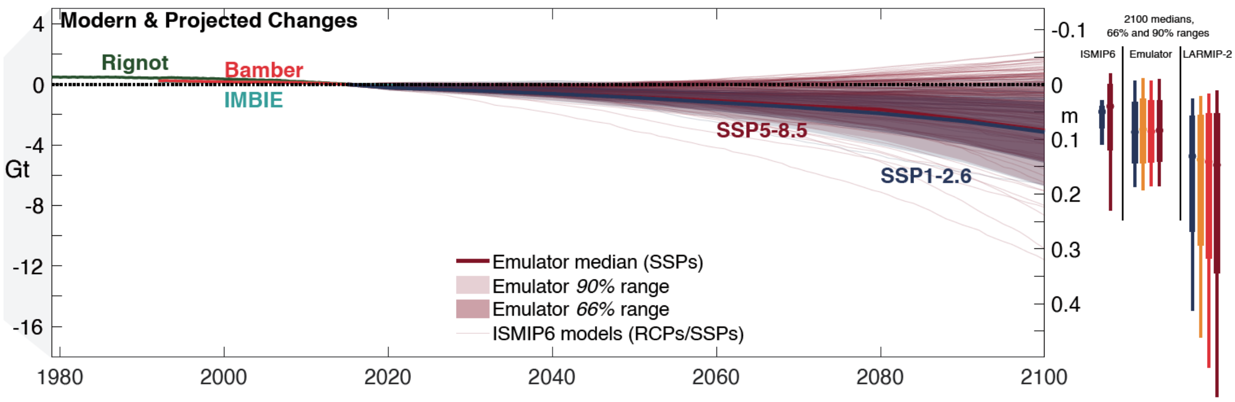

Figure 1.9: Cumulative mass loss (and sea level equivalent) since 2015, with satellite observations shown from 1993 (Bamber et al., 2018; Shepherd et al., 2018) and observations from 1979 (Rignot et al., 2019), and projections from Ice Sheet Model Intercomparison Project for CMIP6 (ISMIP6) to 2100 under RCP8.5/SSP5-8.5 and RCP2.6/SSP1-2.6 scenarios (thin lines from Seroussi et al., 2020; Edwards et al., 2021; Payne et al., 2021) and ISMIP6 emulator under SSP5-8.5 and SSP1-2.6 to 2100 (Edwards et al., 2021, shades and bold lines;). Bar and whiskers on the right represent the 17th–83rd and 5th–95th percentile ranges for ISMIP6, ISMIP6 emulator, and LARMIP-2 (Levermann et al., 2020) including surface mass balance (SMB) at 2100 [Figure Credit: AR6 – Fox-Kemper et al. (2021)].

1.2.3.2 Future projections: what can we expect?

Projections to 2100

Recently, the IPCC assessed in its AR6 that the median and likely (66%) range of sea-level contributions of the AIS by 2100 were 0.11 (0.03–0.27) m under SSP1-2.6 and 0.12 (0.03-0.34) m under SSP5-8.5. These estimates were based, notably, on results from the Ice Sheet Model Intercomparison Project (ISMIP6; Seroussi et al., 2020; Payne et al., 2021, Figure 1.9), which represents the most comprehensive and up-to-date synthesis of sea-level projections from the AIS by 2100, using 11 ice-sheet models forced by six CMIP5 (Seroussi et al., 2020) and four CMIP6 (Payne et al., 2021) climate models. Overall, projections from the ISMIP6 ensembles as well as from its statistical emulation (Edwards et al., 2021) led to a lower and narrower range than AR6 estimates (-0.01 to +0.10 and -0.01 to +0.09 m under SSP1-2.6 and SSP5-8.5, respectively, with an identical median of 0.04 m, were obtained by the emulated projections; Edwards et al., 2021). However, this lower estimate may be associated with the fact the ISMIP6 ensemble excluded the historical dynamic response (Aschwanden et al., 2021; Fox-Kemper et al., 2021), which has been estimated to be $0.33 \pm 0.16\,\mathrm{mm\,yr^{-1}}$, i.e., 0.03 ± 0.01 m in 2100 (Fox-Kemper et al., 2021).

In addition to ISMIP6, recent projections that contributed to AR6 estimates have also arisen from another multi-model intercomparison project, LARMIP-2[1] (Levermann et al., 2020). The majority of these recent projections indicate (in agreement with projections prior to AR6, e.g., Bulthuis et al., 2019; Golledge et al., 2019) that, by the end of the century, the AIS will overall lose mass and contribute to SLR, under all emissions scenarios, with most thinning occurring in the Amundsen Sea sector in WAIS and Totten Glacier in EAIS (Fox-Kemper et al., 2021; Stokes et al., 2022), thus similar to the current trends in AIS mass changes (section 1.2.2). No clear dependence on the emissions scenario emerges, partly due to opposite scenario dependencies between WAIS and EAIS (Edwards et al., 2021) as well as to large individual variations across climate and ice sheet models projections (Seroussi et al., 2020; Edwards et al., 2021; Fox-Kemper et al., 2021). Climate and ice-sheet models do not project that the AIS response will be the same

[1]Linear Antarctic Response to basal melting Model Intercomparison Project phase 2

under high or low GHG emissions in 2100, but rather there is no consensus on the sign of the continent-wide change, essentially arising from uncertainties in the future evolution of the EAIS (Fox-Kemper et al., 2021; Stokes et al., 2022; Edwards et al., 2021; Mengel and Levermann, 2014). A recent estimate suggests a small positive contribution to SLR from the EAIS at 2100, but with a wide range depending on the scenario (-4 to +22 cm for the 5th to 95th percentiles; Stokes et al., 2022). Nevertheless, the AR6 states that *'it is likely that the Antarctic ice sheet will continue to lose mass throughout this century under all emissions scenarios, i.e., that dynamic losses driven by ocean warming and ice shelf disintegration will likely continue to outpace increasing snowfall* (Fox-Kemper et al., 2021). It also underlines, however, that the upper end of the projections is not well constrained, especially under high emissions scenarios, in particular due to projections including MICI, which project higher contributions to SLR from the AIS of 0.34 (0.19 to 0.53) m by 2100 under RCP8.5 (DeConto et al., 2021).

Projections beyond 2100

While there is no consensus on the sign of AIS mass change by 2100, ice mass loss is expected to dominate on the long term (Fox-Kemper et al., 2021). By 2300, the ranges of projections from AR6 are –0.14 to +0.78 m SLE under RCP2.6/SSP1-2.6, and –0.27 to 3.14 m SLE under RCP8.5/SSP5-8.5 (Fox-Kemper et al., 2021). In contrast, including hydrofracturing and MICI mechanisms, Antarctic projections from DeConto et al. (2021) contribute much more to GMSL than the assessed ranges under sustained very high GHG emissions, that is, around 7–14 m by 2300.

The majority of AIS projections after 2300 are projections that, in fact, estimate the long-term changes of the AIS in response to early-millennium warming, i.e., until 2100 (e.g., Chambers et al., 2021; Golledge et al., 2019; Rodehacke et al., 2020; Lipscomb et al., 2021) or 2300 (e.g., Golledge et al., 2015; Bulthuis et al., 2019). These 'committed' mass changes are due to both the high inertia in the Antarctic system (the ice sheet responds slowly to imposed perturbations) and the fact that current or coming climate change may trigger self-sustained ice loss due to positive feedback mechanisms (such as, e.g., the marine ice sheet instability) which may unfold over longer timescales (Golledge et al., 2019; Clark et al., 2020; Chambers et al., 2021).

Overall, most studies project that only part of WAIS would be lost under 1–2°C warming above pre-industrial levels on multi-centennial (limiting the AIS contribution below 0–1.3 m SLE; e.g., Golledge et al., 2015; Bulthuis et al., 2019) to multi-millennial (contribution below 1.6 m SLE; Rodehacke et al., 2020; Garbe et al., 2020) timescales. Majority or total loss of WAIS under 1–2°C warming (therefore producing a sea-level contribution exceeding 2 m SLE) only occurs under the higher end of the warming range, or high ocean warming and/or high basal melting around WAIS (Lipscomb et al., 2021), or MICI (DeConto and Pollard, 2016; DeConto et al., 2021). Under $\sim$2–3°C of warming, complete or near-complete WAIS collapse is projected in most studies after multiple millennia (e.g., Garbe et al., 2020), with continent-wide mass losses of $\sim$2–5 m SLE or more. This could occur faster, i.e., on multi-centennial timescales, under very high basal melting (Lipscomb et al., 2021) or widespread ice shelf loss (Sun et al., 2020) and/or MICI (DeConto et al., 2021). If warming exceeds $\sim$3°C above pre-industrial levels, part of the East Antarctic ice sheet (usually the Wilkes basin) is projected to be also lost (although with low confidence) on multi-millennial timescales, with total AIS mass loss equivalent to around 6–12 m SLR or more (Garbe et al., 2020; Fox-Kemper et al., 2021). Under these higher warming scenarios, mass loss could be much smaller (less than 2 m SLE, especially at multi-centennial timescales) if the dynamic response is small (e.g., with less responsive sliding laws; Bulthuis et al., 2019).

As a summary, on longer timescales of a few centuries to several millennia, the projected SLR contribution of the AIS for low warming levels is limited to below 1–2 m, although with a probability distribution that is not Gaussian and presents a long tail towards high values due to potential MICI (DeConto and Pollard, 2016; DeConto et al., 2021; Fox-Kemper et al., 2021). Under stronger forcing, grounding-line retreat into the central WAIS region, as well as in some marine basins of the East Antarctic ice sheet is expected, thereby contributing several meters to global mean sea level rise (Bulthuis et al., 2019; Ritz et al., 2015; Pattyn and Morlighem, 2020; Golledge et al., 2019; DeConto et al., 2021; Garbe et al., 2020). Although the time of onset of the collapse is quite different across models and scenarios, all models produce WAIS collapse under unmitigated emission scenarios on multicentennial time scales (Pattyn and Morlighem, 2020). Projections are, however, less convergent on the future fate of the EAIS. Under sustained emissions and warming, modelling projections point to contributions from EAIS marine-based sectors (essentially the Wilkes and Aurora basins, and also potentially Recovery) reaching 1–3 m or more by 2300, and around 2–5 m by 2500 (Stokes et al., 2022). In contrast, if warming is limited so well below +2°C above pre-industrial levels, the contribution from the EAIS to global mean SLR would likely remain below +0.5 m at 2500 (Stokes et al., 2022).

1.2.3.3 Summary

As a conclusion, the main drivers of future AIS change will likely be surface mass balance, sub-shelf melting and ice shelf disintegration (Fox-Kemper et al., 2021), which, despite current uncertainties, are all expected to increase in a warming climate. The future Antarctic contribution to global-mean sea level thus has two parts: increasing snowfall (essentially in the interior terrestrial regions of the ice sheet), which is expected to reduce global-mean sea level by a few centimetres this century (Kittel et al., 2021), and ice discharge into the ocean, which is very uncertain. The latter is determined by the outflow of ice across the grounding line. Ice discharge can increase if buttressing by ice shelves is reduced by (1) ice-shelf thinning, essentially caused by enhanced oceanic melting due to circulation changes or direct warming, or (2) partial or total ice-shelf collapse, caused by widening of surface crevasses by melt water due to atmospheric warming. Therefore, the future fate of the AIS under a warming climate is dynamically tied to changes of the floating ice shelves.

Given the susceptibility of ice shelves to changing atmospheric and oceanic conditions, and of grounded ice to shrinking ice shelves, we can expect an increasing contribution from Antarctica to SLR on relatively short time scales, i.e., by the end of the century. Beyond 2100, deep uncertainties remain: owing to strong positive feedback mechanisms, some regions of the ice sheet may reach a tipping point, potentially leading to high rates of SLR. These regions vulnerable to irreversible mass loss are regions whose bedrock is currently grounded deep below sea-level. This is the case in most of the WAIS, but also in some regions of the EAIS. In addition, while solid Earth feedbacks could reduce ice loss, MICI might give contributions higher than the likely ranges.

The majority of projections suggest that the trigger of future Antarctic mass loss will occur in the WAIS, with a potential threshold for its instability close to 1.5°C–2°C (Pattyn et al., 2018; Fox-Kemper et al., 2021; McKay et al., 2022). Above this threshold, WAIS collapse would likely occur, unfolding over centuries to millennia (Garbe et al., 2020). The future behaviour of the EAIS remains much more uncertain. While only some glaciers in the Aurora basin currently show signs of mass loss, all EAIS marine basins (and especially the Wilkes basin) seem vulnerable to future ocean forcing (Stokes et al., 2022; Mengel and Levermann, 2014). Therefore, the major uncertainty in future Antarctic mass loss arises from the EAIS, which has a far greater sea-level contribution potential (52.2 m SLE; Morlighem et al., 2020).

Overall, these uncertainties in the magnitude and the timing of future AIS mass loss essentially stem from our poor knowledge of both the drivers of change and key mechanisms that operate in the dynamics of marine ice sheets (Pattyn and Morlighem, 2020).

1.3 Physically-based models of Antarctic mass change

Projections of ice-sheet contributions to SLR are established using ice flow models that compute the evolution of ice sheets under given climate scenarios. These ice-sheet models are multi-physics numerical models incorporating at best our physical understanding of the processes involved in various sub-systems such as the ice dynamics, surface and basal processes, and thermodynamics. We have highlighted in section 1.2.3.2 how projections of the AIS contribution to sea-level changes are affected by uncertainties. In particular, we have shown that even projections by the end of the century (e.g., Seroussi et al., 2020) are associated with relatively large uncertainties. While some of this range is related to uncertainties in the climate response to a given emissions scenario and differences in initialisation, ice-sheet model projections are also subject to uncertainties in the representation of physical processes. Many of these models were constructed to study the evolution of ice sheets across glacial–interglacial cycles, and are not therefore ideally suited to making projections for this century (Pattyn, 2018). Indeed, few ice-sheet models are currently able to produce simulations that fit observations of historical and current AIS mass changes within observational uncertainty (Aschwanden et al., 2021). Accordingly, the past decade has seen the modelling community repurpose these models, increasing confidence in the skill of ice-sheet models (particularly in the interactions with boundary conditions, such as ice–ocean, ice–atmosphere and ice–bedrock), but they still lag behind other areas of the climate system. In this section, we provide an overview of the principles and main challenges behind the simulation of the behaviour of the AIS using numerical ice-sheet models.

1.3.1 Ice-sheet models

As developed in section 1.2, many of the observed ice-sheet mass-loss processes are not driven directly by ice surface melting but instead involve the flow of ice. Indeed, ice flows through gravity-driven viscous creep, with the ice sheet spreading under its own weight (with the rate of ice deformation which is a highly nonlinear function of the temperature). A proper representation of the ice flow and dynamics is therefore crucial to reproduce and understand AIS changes.

The flow of an ice body is described by the following equation expressing the balance of linear momentum:

$$\rho_i \frac{d\boldsymbol{v}}{dt} = \nabla \cdot \tau + \rho_i g \tag{1.1}$$

where ρ_i is the ice density, $\boldsymbol{v}$ is the velocity field, τ is the Cauchy stress tensor, and g is the gravitational acceleration.

Ice sheets move as very slowly flowing viscous fluids. By ‘slowly flowing’ it is meant that the ratio of inertial forces to viscous forces (the Reynolds numbers) are very low, in fact so low that inertial/acceleration terms can safely be ignored (Hindmarsh, 2018). Removing them from the mechanical balance equation above results in a simpler set of equations (each corresponding to one of the spatial dimensions under consideration) known as the Stokes equations. Some ice-sheet models (called ‘Full Stokes’; Durand et al.,

2009) solve the full system of linear momentum equations. However, due to the considerable computational effort needed, approximations to these equations are often used, such as higher-order, shallow-ice and shallow-shelf approximations (SIA and SSA, respectively). Such approximations relate on the fact that the ice-sheet flow may be simplified by different ice flow regimes – one dominated by internal deformation (vertical shear) and another dominated by sliding (membrane stresses) at the ice–bedrock interface – and the switching from one to another associated with the pressure dependence of the ice melting point.

1.3.2 The main challenges in modelling the Antarctic ice sheet

1.3.2.1 Paradigm shift in ice sheet modelling: the challenge of modelling grounding-line migrations

When the first ice-sheet models were designed, in the 1990s, ice sheets were believed to be a slow component of the climate system and were therefore employed at a rather coarse resolution to investigate ice-sheet changes during glacial-interglacial cycles (Pattyn, 2018). Though the theoretical developments around the possibility of rapid mass change arising from marine ice sheets were proposed in the 1970s by Weertman (1974), thus before the first ice-sheet models, it is only after the mathematical proof of the MISI theory (see section 1.2.3.1.1) was put forward in the beginning of the century by Schoof (2007) that more focus was given to the importance of the ice shelves (hence contradicting the previous idea that the ice sheet and the ice shelves are mechanically uncoupled) and to a proper representation of the migration of the grounding line, i.e., the limit between grounded and floating ice, defined by the following flotation criterion:

$$\rho_i h = \rho_w (z_{sl} - b) \tag{1.2}$$

where ρ_i and ρ_w are the ice and water densities, respectively, h is the ice thickness, z_{sl} the sea-level elevation and b the bedrock elevation.

By providing an analytical solution to which models could rely and conform, the study by Schoof (2007) had a profound impact on ice-sheet model development. It led notably to model intercomparisons, such as the Marine Ice Sheet Model Intercomparison Project (MISMIP; Pattyn et al., 2012, 2013), which contributed to a significant improvement of the representation of grounding-line migration in ice-sheet models. The grounding line is an area characterised by a transition from a shear-dominated ice flow in the grounded ice sheet to an ice flow dominated by longitudinal stretching in the ice shelves. For this reason, to accurately capture grounding-line migration, MISMIP numerical experiments showed that it is necessary to resolve the transition zone at a sufficiently fine resolution (depending on the choice of basal sliding relation and melt rate at the grounding line; Gladstone et al., 2017), unless parameterisations based on analytical solutions (Schoof, 2007) are used (Pollard and DeConto, 2012b; DeConto and Pollard, 2016; Pattyn, 2017). In addition, model intercomparisons highlighted the fact that resolving the grounding-line migration is dependent of the level of physical approximation (see section 1.3.1) and requires the use of approximations to the flow of ice that are more complex than SIA, ranging from SSA to the computationally-expensive full-stokes models (Pattyn et al., 2013). Nonetheless, certain effects, such as buttressing, are not yet quantified/verified accurately, though new analytical tools are becoming available (Haseloff and Sergienko, 2018).

1.3.2.2 Basal sliding

Most models of ice dynamics assume that basal sliding can be described by a friction law that relates shear stress τ_b at the bed to sliding velocity u_b at the bed in the form $\tau_b = f(u_b)$ (Schoof and Hewitt, 2013). Sliding is often a dominant component in ice flow (Engelhardt and Kamb, 1997) (especially in the fast-flowing ice streams which drain about 90% of the total Antarctic ice flux; Bennett, 2003) and the choice of constitutive relation $f(u_b)$ can be at least as important as the details of ice rheology (Schoof and Hewitt, 2013; Ritz et al., 2015). Unfortunately, the ice–bed interface being usually out of reach, the formulation of a friction law has been a long-standing problem in glaciology, and various laws describing different physical processes at the roots of basal sliding of ice have been developed over the years (Weertman, 1957; Budd et al., 1979; Schoof, 2005; Tsai et al., 2015). Ice-sheet models typically use friction laws that depend on the basal velocity linearly or nonlinearly (Weertman, 1957), which is generally a good approximation for a hard bedrock (Brondex et al., 2019). Many Antarctic ice streams, however, are known to be lying on soft beds that have a layer of deformable till. Recent studies and laboratory experiments suggest that the rheology of the till is plastic at large strain, with a critical strength τ^* depending on effective pressure N, i.e., the difference between ice overburden pressure and water pressure (Alley et al., 1986). New mathematical models of basal sliding (based on the physics of the substrate or the interactions with it) have therefore been developed to account for both soft and hard beds (Gagliardini et al., 2007; Schoof, 2005; Tsai et al., 2015; Zoet and Iverson, 2020; Minchew and Joughin, 2020). The inclusion of these physically-derived friction laws into ice-sheet models and their validation with observations on a larger scale are critical to further improve the predictive skills of numerical models.

1.3.2.3 Model initialisation

Forward simulations performed with numerical ice-sheet models require an initialisation procedure that precedes their use for projections, by providing initial boundary conditions and surfaces (see for example Figure 2.2) to the ice-sheet model. Since ice sheets have a slow response time, their initial conditions may influence their evolution for centuries to millennia (Seroussi et al., 2019). The initial state of the model thus represents a key aspect of projecting future Antarctic mass loss with dynamical ice-sheet models, and this especially for low-emissions scenarios and the short term because the anticipated mass loss is relatively small in comparison to the total mass of the ice sheets (Pattyn et al., 2018; Seroussi et al., 2019).

Two main families of initialisation strategies are employed at present. Since ice-sheet models were initially applied for palaeo-climatic studies on long time scales, initialisation was generally obtained from a long spin-up over glacial-interglacial periods, leading to a steady-state ice sheet (both in terms of geometry and thermodynamics). However, for predictions on shorter time scales (decades to centuries), a stable spin-up generally leads to an ice sheet geometry and flow velocities far different from the ones currently observed, which is one of the reasons why such ice-sheet models may respond differently than observations suggest. Moreover, using a steady-state for initialising the ice sheet, though it assures that the internal properties of the ice sheet are consistent with each other, prevents models from properly accounting for the dynamical mass losses observed over the last decade, as the present-day ice sheet is not in steady state (section 1.2.2). Motivated by the increasing ice-sheet imbalance of the ASE glaciers over the last decades, and supported by the recent increase in satellite data availability, alternative data-assimilation methods are progressively used to evaluate unknown fields using time-evolving states accounting for the transient nature of observations and the model dynamics (e.g., Morlighem et al., 2013). The data being assimilated are typically satellite-derived surface flow speeds, thinning and thickening rates, or ice-sheet surface elevation. Such methods have a greatly improved representation of the current geometry and surface velocity (Seroussi et al., 2019; Morlighem et al., 2013; Berdahl et al., 2022). Combinations of spin-up and data-assimilation methods, using long transient spin-ups integrating simple inverse methods to match present-day ice sheet geometry have been developed (Pollard and DeConto, 2012a), thereby combining the best of the two approaches. The different existing initialisation methods lead to large differences in the initial conditions from which projections are made and therefore create a significant spread in projected contributions to future SLR — even when forced with similar datasets (Goelzer et al., 2018; Seroussi et al., 2019; Berdahl et al., 2022). Model initialisation thus remains a major limitation and source of uncertainty for predictions of the AIS response to climate change based on numerical ice-sheet models.

1.3.2.4 Simulating the ice-sheet interactions with its boundaries

Ice sheets interact directly and indirectly with every component of the Earth system through a series of processes and feedbacks. In turn, these Earth-system components exert a control on ice-sheet evolution. This two-way interaction is the basis of numerous ice sheet–Earth system feedback loops that modulate (sometimes strongly) the ice-sheet response to external climate forcing (Fyke et al., 2018; Oerlemans, 1981). More specifically, the crucial importance of the ice-sheet interactions with the ocean, atmosphere and solid Earth for simulating the Antarctic contribution to future sea-level evolution has been elaborated in section 1.2. For this reason, numerical ice-sheet models should ideally be coupled with computational models of these other components of the Earth system, in order to appropriately represent the interactions and feedbacks between them and the ice sheet. Unfortunately, such coupled simulations are hampered by the high computational cost of most climate, ocean, and Earth models. In addition, different spatial and temporal resolutions between the different components of the Earth system are used, thus generally requiring asynchronous coupling and appropriate downscaling techniques (see for example Pelletier et al., 2022). Owing to these challenges, interactions between an ice sheet and its environment are often prescribed in a parameterised way (e.g., Tarasov and Peltier, 1999; Zweck and Huybrechts, 2005; Garbe et al., 2020; Pollard and DeConto, 2012b; Bulthuis et al., 2019; Le Meur and Huybrechts, 1996). While allowing for a direct representation of the interactions between the ice sheet and the main components of the Earth system, these parameterisations may suffer from limitations in the representation of physical processes and involve a set of free parameters that need to be determined (Tarasov and Peltier, 1999; Le Meur and Huybrechts, 1996). Here, we provide more details on the commonly-used methods to simulate the interactions between the ice sheet and, on one side, the ocean and the atmosphere, i.e., the climatic components of the Earth system, and on the other side, the solid Earth and relative sea-level.

1.3.2.4.1 Ocean and atmosphere: coupling with climate models

Ice-sheet models need climate data to capture the past and project the future evolution of ice sheets and glaciers. Indeed, atmospheric and oceanic forcings are the primary drivers of ice-sheet change, and the majority of present-day mass loss (essentially the ASE) is driven by changes in ocean circulation (Rignot et al., 2019; Shepherd et al., 2018; Paolo et al., 2015). As explained above, the optimal way to consider and reproduce the interactions between the ice sheet and both the ocean and the atmosphere is to perform coupled simulations allowing to consider the different potential feedbacks between the different Earth system components (see sections 1.2.3.1.3 and 1.2.3.1.2). Unsurprisingly, simulations performing two-way coupling between the ice sheet and either the atmosphere (e.g., Le clec'h et al., 2019; van Kampenhout et al., 2019) or the ocean (e.g., Seroussi et al., 2017; Golledge et al., 2019), have displayed different sensitivities to climate change than uncoupled ones, even though they remain limited to decadal or multidecadal time scales. Simulations that include a complete coupling between ice, ocean, and atmosphere is currently the subject of ongoing research and remain few in number (Siahaan et al., 2021; Pelletier et al., 2022).

Instead of such two-way couplings, ice-sheet models are typically forced by independent atmospheric and oceanic forcings. More specifically, a common intermediate solution used by standalone continental-scale ice models is that the evolution of the climate forcing is provided by climate models (e.g., Seroussi et al., 2020; Payne et al., 2021). We therefore speak of a one-way (or 'offline') coupling, i.e. the ice-sheet model receives information (for example, the amount of snowfall) calculated by the atmospheric (ocean) model, but the atmospheric (ocean) model does not receive any updated variables from the ice-sheet model (for example the ice sheet surface elevation). For this reason, applying such one-way coupling as climate forcing is associated with an important drawback related to the fact that climate models consider a static ice-sheet geometry, which could lead to significant biases both on the oceanic (due to the non-consideration of changes in the shape of the sub-shelf cavities; Reese et al., 2018a; Jenkins, 1991; Asay-Davis et al., 2017) and atmospheric (due to the non-consideration of changes in ice-sheet elevation and geometry) forcings, thereby neglecting potential feedbacks, especially for long-term projections (Oerlemans, 1981; Kittel et al., 2021; Le clec'h et al., 2019).

For the atmosphere interface, the optimal and more robust way to evaluate the future Antarctic SMB in order to force projections from ice-sheet models is to use outputs from regional climate models (RCMs), themselves forced at their boundaries by projections from global climate and earth system models (ESMs). Polar-oriented RCMs perform best compared to observations, in particular for simulating precipitation patterns (Fettweis et al., 2013; Kittel et al., 2021). Indeed, the use of high-resolution ($\sim$10s of km) is necessary to resolve the boundary layer processes and interactions that drive SMB, such as topography, precipitation and sublimation (Lenaerts et al., 2019). Unfortunately, outputs from such RCMs downscaling ESMs are not yet available on timescales longer than the end of the century. A common alternative is the use of ESMs outputs directly (e.g., Seroussi et al., 2020). However, this involves several compromises related to their coarse resolution and their low sophistication to represent important physical processes of polar regions (Kittel et al., 2021). An alternative to using SMB directly derived from a climate model is to simulate it within the ice-sheet model using parameterisations, so that the evolving ice sheet topography can dynamically alter the SMB (Edwards et al., 2014). The most common approach is the use of a simple positive degree-day (PDD) scheme (e.g., DeConto et al., 2021; Golledge et al., 2019; Garbe et al., 2020), which parameterises SMB as a function of temperature and precipitation (typically supplied from a regional or global climate model) and possibly applies a simple snowpack model (Janssens and Huybrechts, 2000). The SMB–elevation feedback is usually incorporated through the use of a lapse rate correction of the air temperature, the latter altering the precipitation rate through a scaling factor (DeConto et al., 2021; Garbe et al., 2020; Rodehacke et al., 2020). A more intermediate option is the use of Energy Balance models (EBM; e.g., van de Wal and Oerlemans, 1994). These simpler and faster models have been shown to have biases of the same order as RCMs compared with observations and therefore remain useful tools for long-term simulations or coupling with ice sheet models (Fettweis et al., 2013). Note that when projections from climate models are not available, a simpler common method to apply an atmospheric forcing consists in modifying present-day fields (usually derived from RCM projections) by applying a spatially-uniform shift in surface temperature and correcting the precipitation rate for such change using a Clausius-Clapeyron-like relation (Garbe et al., 2020; Golledge et al., 2015; Bulthuis et al., 2019).

For the ocean interface, an additional complicating issue lie in the fact that most climate models do not consider some areas that are important for ice sheets, such as the ocean beneath the floating ice shelves. Indeed, sub-shelf melt is particularly determined by the hydrographic properties of the water entering the ice-shelf cavity (section 1.2.3.1.2). Resolving the circulation of the water masses within that cavity is therefore required. Ideally, this would be done by high-resolution ocean circulation models to link large-scale ocean circulation to sub-shelf melt (Seroussi et al., 2017; Favier et al., 2019). Given the immaturity and high cost of such models, alternative approaches are needed for generating basal melt rates in ice-sheet models, especially for multi-century and multi-millennium projections of the effects of ocean melting on ice sheet dynamics (Asay-Davis et al., 2017). Instead, melt rates at the base of floating ice shelves are usually determined through simplified ocean circulation models (i.e., physically-based parametrisations) that approximate the local thermal forcing (i.e., the difference between the in situ ocean temperature and the in situ freezing temperature) based on ocean properties simulated by ocean models for the region in front of the ice shelf (Reese et al., 2018a; Lazeroms et al., 2018, 2019; Jourdain et al., 2020; Burgard et al., 2022). Simple parameterisations relate sub-shelf melting to ocean ambient properties as a function of ice-shelf depth (Beckmann and Goosse, 2003; Holland et al., 2008; Jourdain et al., 2020; Burgard et al., 2022), in either a linear or a quadratic fashion, which leads to higher melting close to the grounding line (Burgard et al., 2022). Such simple parameterisations typically disregard the contribution of accretion of marine ice to the basal mass balance, which has been shown to be significant for some ice shelves (Bernales et al., 2017; Adusumilli et al., 2020). Alternatively, more sophisticated ocean-model couplers are designed to represent the ocean overturning circulation within the ice shelf cavities, i.e., the advection of ocean heat into the cavity and the subsequent transformation of ocean properties within a melt water plume that flows from the grounding line to the ice front along the ice shelf base. Such more complex parameterisations link far-field ocean temperature and salinity to sub-shelf melting either via a box model of overturning circulation within ice-shelf cavities (Olbers and Hellmer, 2010), such as the PICO (Reese et al., 2018a) and PICOP (Pelle et al., 2019) models, or via a plume model, a two-dimensional basal melt rate parametrisation (Lazeroms et al., 2018, 2019) based on the theory of buoyant melt-water plumes (Jenkins, 1991). In all cases, these parameterisations, which give rise to a variety of melt patterns (Favier et al., 2019; Burgard et al., 2022), allow to approximate two-way interactions between the ice sheet and the ocean by representing the influence of changes in ice-shelf geometry on the pattern of basal melt rates. Similarly to the atmosphere, if not provided by climate models, a simple common method to approximate far-field ocean temperature changes consists in relating changes in ocean temperature to atmospheric temperature changes (Bulthuis et al., 2019; Garbe et al., 2020).

Finally, while iceberg calving is responsible for the other part of the ice mass loss at the oceanic margins of the Antarctic ice sheet, its representation and quantification remains one of the grand challenges of ice sheet modelling, and no general calving law exists yet, which profoundly limits our ability to model catastrophic calving events (Pattyn and Morlighem, 2020). Similarly, current models of MICI (e.g., DeConto and Pollard, 2016; DeConto et al., 2021) rely on quasi-empirical parameterisations extrapolated from limited observations to simulate retreat (Crawford et al., 2021).

1.3.2.4.2 Solid Earth and relative sea-level: solving the Sea Level Equation

We have elaborated in section 1.2.3.1.4 why the evolving shapes of the solid Earth and the adjacent geoid act as fundamental boundary conditions on the dynamics of the modeled ice sheet. The spatially-varying evolution of RSL changes can be determined by solving the so-called sea-level equation (SLE; Farrell and Clark, 1976), which determines changes to the gravity field, and therefore how melt water

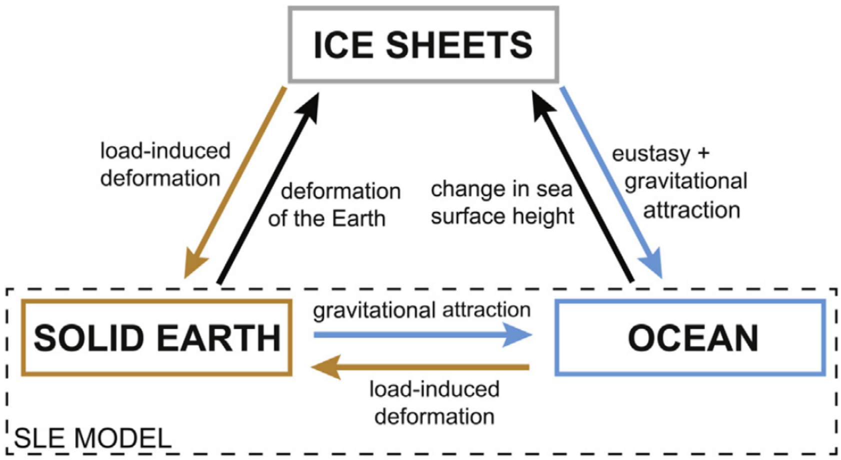

Figure 1.10: **Interactions between ice sheets, the solid Earth and the ocean.** Ice sheets are modelled with an ice-sheet model, the response of the solid Earth is typically calculated using a viscoelastic Earth model, and changes to the ocean (sea surface) reflect changes in the shape of the geoid. Together, the solid Earth and ocean components represent the change in RSL as determined by solving the SLE (dashed box) [Figure Credit: de Boer et al. (2017)].

is redistributed across the ocean, enabling to relate the geoid (i.e. the sea-surface) and the bedrock in a gravitationally self-consistent solution (Gregory et al., 2019).

Full self-gravitating viscoelastic solid-Earth models (SGVEMs) incorporate gravitational, rotational and bedrock deformational responses to ice-ocean mass redistribution and are thus able to solve the sea-level equation (e.g., de Boer et al., 2014; Gomez et al., 2015; Pollard et al., 2017; Gomez et al., 2018). However, they are very expensive in terms of computational time, essentially since calculations to determine the change in sea level due to the melting of an ice sheet must be carried out iteratively due to the multiple feedbacks to be considered because of the interplay of the different components of GIA processes (Whitehouse, 2018, see Figure 1.10).

On the other side of the spectrum are ELRA models (Elastic Lithosphere-Relaxed Asthenosphere), often used in conjunction with ice-sheet models because of their computational efficiency. They consider an elastic lithosphere, defined by a given effective lithosphere thickness and a relaxation equation for asthenospheric response with a characteristic response time as a function of asthenosphere viscosity. However, the ELRA model ignores the gravitational attraction of the oceans by the ice mass (and thus gravitationally-consistent sea-surface changes; Konrad et al., 2016, 2014), meaning that it does not solve the SLE, making full SGVEM an optimal choice for properly approximating the effects of GIA feedbacks on ice dynamics.

A complicating issue is that several recent studies (i.e., Kaufmann et al., 2005; van der Wal et al., 2015; An et al., 2015; Heeszel et al., 2016) suggest that Antarctica is located on a region of the solid Earth that displays large spatial variations - across several orders of magnitude - in rheological properties (Whitehouse et al., 2019, see section 1.2.3.1.4). However, accounting for lateral variations in Earth structure leads to a considerable increase in computational cost, which explains why, apart from some exceptions (e.g., Gomez et al., 2018), most coupled ice sheet–solid Earth models omit them.

1.4 Thesis objectives and outline

We have elaborated in this introduction how the interactions between the AIS and its surrounding environment, namely the atmosphere, the ocean, and the bedrock, strongly influence its stability, especially by triggering or dampening instabilities threatening the ice sheet. In addition, we have highlighted how large uncertainties remain in determining how these interactions will evolve under a warming climate and thereby influence the future evolution of the AIS.

In this context, two main objectives of this thesis may be highlighted. Using a numerical ice-sheet model, we wish to:

1. Contribute to the estimation of the future contribution of the AIS to sea-level changes and its uncertainty.
2. Assess the influence of uncertainties approximating the current limits of our scientific understanding in the interactions of the AIS with the other components of the Earth system on future AIS mass change.

In particular, since interactions with the solid Earth generally occur at relatively long timescales and ice sheets are known to have a relatively low response time and inertia, we will investigate the future evolution of the AIS with a specific focus on the long-term, i.e., centennial to multi-millennial timescales.

To do so, the manuscript of this thesis consists of five chapters.

Following the above general introduction (Chapter 1), **Chapter 2** provides a description of the methodology applied in order to produce credible projections of AIS mass changes. It also describes 'f.ETISh', the ice-sheet model used in this thesis, as well as the main contributions that have been made to its development in the framework of this thesis. Finally, a new approach for estimating the sea-level contribution from marine ice sheets in ice-sheet models is presented in a standalone section taken from a paper published in *The Cryosphere*:

Goelzer, H., **Coulon, V.**, Pattyn, F., de Boer, B., and van de Wal, R.: Brief communication: On calculating the sea-level contribution in marine ice-sheet models , *The Cryosphere*, 14, 833–840, https://doi.org/10.5194/tc-14-833-2020, 2020.

Chapters 3 and 4 then present the main results obtained during this thesis. These chapters correspond to papers that have been published or will be submitted to international peer-reviewed journals. Each introduction to these chapters presents the concepts needed to (re)place them in their respective contexts. Consequently, there may be some redundancy with the introduction of these different chapters as well as with Chapter 1, which is however necessary for each chapter to be self-sufficient. More specifically,

- **Chapter 3** investigates the influence of GIA feedbacks on the long-term stability of the AIS. Especially, a probabilistic assessment of the influence of uncertainties in viscoelastic properties on the response of the Antarctic ice sheet to future warming is realised in a standalone section taken from a paper published in *Journal of Geophysical Research: Earth Surface*:

  Coulon, V., Bulthuis, K., Whitehouse, P. L., Sun, S., Haubner, K., Zipf, L., and Pattyn, F.(2021).

Contrasting Response of the West and East Antarctic ice sheets to glacial isostatic adjustment, *Journal of Geophysical Research: Earth Surface*, 126, doi:10.1029/2020JF006003.

In addition, this chapter evaluates the influence of intra-regional variability in viscoelastic properties on Antarctic long-term stability, with a specific assessment of the influence of GIA feedbacks at high spatiotemporal resolution.

- **Chapter 4** is a draft of a publication that will be submitted shortly:

  Coulon, V., Klose, AK., Kittel, C., Winkelmann, R., and Pattyn, F. Projections of long-term future Antarctic ice loss with a historically-calibrated ice-sheet model, in preparation.

  It studies the relative importance of surface mass balance and ocean-induced melt processes on the long-term sensitivity of the Antarctic ice sheet as well as their associated uncertainties.

Finally, **Chapter 5** summarizes the main results of this PhD thesis and presents perspectives for future research.