Past Antarctic ice sheet dynamics (PAIS) and implications for future sea-level change

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Past Antarctic Ice Sheet (PAIS) dynamics and implications for future sea-level change

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The legacy of the Scientific Committee on Antarctic Research's (SCAR) PAIS strategic research programme includes not only breakthrough scientific discoveries, but it is also the story of a long-standing deep collaboration amongst different multi-disciplinary researchers from many nations, to share scientific infrastructure and data, facilities, and numerical models, in order to address high priority questions regarding the evolution and behaviour of the Antarctic ice sheets (AIS). The PAIS research philosophy is based on data-data and data-model integration and intercomparison, and the development of "ice-to-abyss" data transects and paleo-environmental, extending from the ice sheet interior to the deep sea. PAIS strives to improve understanding of AIS dynamics and to reduce uncertainty in model simulations of future ice loss and global sea level change, by studying warm periods of the geological past that are relevant to future climate scenarios. The multi-disciplinary approach fostered by PAIS represents its greatest strength. Eight years after the start of this program, PAIS achievements have been high-profile and impactful, both in terms of field campaigns that collected unique data sets and samples, and in terms of scientific advances concerning past AIS dynamics, that have measurably improved understanding of ice sheet sensitivity in response to global warming. Here, we provide an overview and synthesis of the new knowledge generated by the PAIS Programme and its implications for anticipating and managing the impacts of global sea-level rise.

1. Research focus of the PAIS programme

Ice sheet and sea-level reconstructions from the past "warmer-than-present" climates of the last 34 million years provide powerful insights into the long-term response of the polar ice sheets to climate changes projected for the twenty-first century. Proximal geological evidence shows the onset of large-scale Antarctic glaciations occurred around the Eocene/Oligocene Transition at approximately 34 Ma (EOT, **e.g. Barrett et al., 1989; Hambrey et al., 1991; Coxall et al., 2005; Escutia et al, 2011; Passchier et al., 2013; Galeotti et al., 2016; Passchier et al., 2016**). Since then, AIS evolution and variability, recorded in the direct geological archives have been widely compared with benthic oxygen isotope proxy record of deep-water temperature and global ice volume from far-field deep ocean locations (e.g. **Zachos et al., 2001**; **De Vleeschower et al., 2017**). Together, they reveal fluctuations in ice volume and extent that can be explained by changes in atmospheric CO_2 and astronomical forcing (**Naish et al., 2001; Pälike et al., 2006; Naish et al., 2009a; Patterson et al., 2014; Hansen et al., 2015; Galeotti et al., 2016; Levy et al., 2019**). Looking at the broad picture, the EOT really marked the transition into an ice-house and cool-house world (**Miller et al., 1991; 2020b; Westerhold et al., 2020**). As the initial cryosphere evolved, only one pole, the South Pole, hosted a continental-size ice sheet. Geological evidence from the Arctic revealed that small ice-sheets or ice caps may have formed in the Northern high latitudes as early as the Eocene (e.g. **Eldrett et al 2007**; **Tripati & Darby, 2018**). Then, along with the gradual decrease in atmospheric CO_2 (**Figure 1a**), Greenland glaciated from the Late Miocene, rapidly followed by episodic and extensive glaciations of most of the high-latitude Arctic margins, after the onset of Northern Hemisphere glaciations 2.7 Ma (**e.g. Thiede et al., 2011**).

During the past 34 million years, changes in global climate and polar ice volume have been paced by orbital forcing – the Milankovitch cycles. These regular glacial-interglacial cycles, which are amplified by internal Earth system feedbacks, occur on a background of secular change over millions of years driven by global plate tectonics and the carbon cycle (**Figure 1**). Rapid stepwise transitions between climate states (e.g. EOT) correspond to thresholds in the Earth system often linked to a combination of tectonic reorganisation, atmospheric greenhouse gas composition and extreme astronomical forcing (**Zachos et al., 2001, 2008; Levy et al., 2019**). Our knowledge of the evolution of the AIS and its influence on global climate is now widely documented by proximal ice and sediment core records (e.g. **EPICA Community Members, 2006**; **Bereiter et al., 2015; Barrett, 2007; McKay et al., 2016; Escutia et al., 2019; Levy et al., 2019).**

The circum-Antarctic seismic stratigraphic records of the continental margins reveal erosive unconformities indicative of at least six main periods of massive Antarctic ice sheet (AIS) advances (e.g. **Steinhauff and Webb 1987; Larter et al., 1997; Cooper et al., 1999, 2011; Brancolini et al., 1995a, 1995b; De Santis et al., 1999; Donda et al., 2007; Kristoffersen and Jokat, 2008**; **Bart & DeSantis, 2012; Gohl et al., 2013; Lindeque et al. 2016; Gulick et al., 2017**). Not all the unconformities have been dated but **Hochmuth et al., (2019, 2020)** recently provided the first attempt of pan-Antarctic correlation of the various regional seismic unconformities (**Figure 1b**). These seismic unconformities likely correspond to: early Antarctic glaciations after the EOT (e.g. **Galeotti et al., 2016**), a transient glaciation at Oligocene-Miocene boundary ~23-21 Ma (e.g. **Naish et al., 2001;** summarised in **Wilson and Luyendyk, 2009**), ice sheet re-advance at the end of the Mid-Miocene Climatic Optimum (MCO) about 15.8-14.2 Ma (e.g. **Levy et al., 2016**), the cooling and expansion of the East Antarctic Ice Sheet (EAIS) at the Middle Miocene Climate Transition (MMCT) ~13.8 Ma (e.g. **Lewis et al., 2007**; **2008; Levy et al., 2016; Pierce et al., 2017**), cooling after a period of warmth known as the Late Miocene Cooling ~ 8 - 5 Ma (LMC) (**McKay et al., 2009; Herbert et al., 2016; Gulick et al., 2017**), and cooling and expansion of marine-based ice during the Plio-Pleistocene Transition (PPT) 3-2.5 Ma (**Naish et al., 2009a; McKay et al., 2012a; Patterson et al., 2014**) and possibly to the Mid-Pleistocene Transition ~1 Ma (e.g. **O'Brien et al., 2007).** These periods evidenced by positive excursions in the deep sea benthic $\delta^{18}O$ isotope records (**Figure 1a**) correspond to episodes of global cooling and ice volume growth periods, generally associated with a decline in atmospheric CO_2 below a threshold (**Figure 1b**) for triggering the expansion of terrestrial or marine-based ice and/or the onset of perennial sea-ice.

Between these cooling periods, multi-proxy global climatic reconstructions and proximal Antarctic geological climate and ice sheet reconstructions provide evidence for intense and brief warm periods during which:

- atmospheric CO_2 levels, surface temperatures and global sea level rose well above present-day levels during the MCO (17-15 Ma) and the middle Pliocene Warm Period (mPWP, 3.3 -3 Ma) (e.g. **Miller et al., 2020b** for a review); both periods were characterised by CO_2 concentrations higher than 400 ppm and up to 800ppm for some specific intervals of the MCO (see **Figure 1a** and references therein).
- atmospheric CO_2 levels were near pre-industrial levels, i.e. lower than 300 ppm, but were associated with warmer global surface temperatures and higher sea-levels than today during the "super interglacials" of the Pleistocene. This was likely driven by astronomical forcing. Examples include marine isotope stage (MIS) 31 (1.081-1.062 Ma), and specific warm Late Pleistocene interglacials such as MIS 11 (425-395 ka) and MIS 5e (130-116 ka) (e.g. **Dutton et al., 2015; Miller et al., 2020b**).

These past warm periods are policy-relevant as they provide accessible examples of how the AIS responded to warmer-than-present global temperatures, comparable to those projected for the coming decades to centuries (**IPCC AR5, 2013; IPCC SCROCC, 2019**). However, using these past warm periods to inform our understanding of the AIS sensitivity under different atmospheric CO_2 levels remains a challenge, in part because Earth system boundary conditions were subtly different than today, and the duration and intensity of these past warm periods was highly variable (e.g. **Dutton et al., 2015; DeConto and Pollard, 2016; Colleoni et al., 2018a; Bracegirdle et al., 2019; Noble et al., 2020**).

For example, the duration of those past warm periods differs, from about ~ 2 million years for the MCO, approximately 300 thousand years for the mPWP to a few millennia for some Pleistocene interglacials. Thus, an approach focusing on specific MCO and mPWP glacial-interglacial cycles and interglacials (with similarities to our present interglacial) is being developed. For example, the Pliocene Model Intercomparison Project community is focussing its ongoing mPWP model-data comparison on the M2-KM5c (3.264-3.205 Ma) and KM5c-KM2 (3.205-3.130 Ma) intervals (**Figure 1b**) (**Haywood et al., 2016**). By using more appropriate forcing and boundary conditions for climate model simulations, discrepancies between models and data generally decrease (e.g. **Otto-Bliesner et al, 2017**).

In most simulations of future ice sheets evolution, model projections typically extend only until the policy horizon of 2100 CE (Common Era). However, some ice sheet models have run projections out as far as 2500 CE (e.g. **Golledge et al., 2015, DeConto and Pollard, 2016; Clark et al., 2016**). The recent IPCC special report **on "Ocean and Cryosphere in a Changing Climate" (IPCC SROCC, 2019)** utilises these projections and the results of a structured expert judgement approach (**Bamber et al., 2019**) to present projections to 2300 CE, that to some extent account for the long-term thermo-dynamical response of the Greenland and Antarctic ice sheets and related instabilities (e.g. **Golledge et al., 2015, DeConto and Pollard, 2016**). Paleoclimatic changes are often considered at timescales of tens of millennia to millennia and, in few archives, at sub-millennial timescale. At such timescales, past reconstructions can inform long-term projections over a few millennia **(e.g. Golledge et al., 2020** for a review**)**, but some refinements at sub-millennial timescales to investigate some abrupt events of the near past are necessary to reconcile with the projections.

During past warm periods, global paleogeography, paleotopography and/or paleobathymetry can differ substantially from today. Periods prior to the Plio-Pleistocene Transition (3.0-2.5 Ma) were characterised by a very different continental and oceanic configuration that yielded changes in the proportion of emerged lands and their locations. This affected surface elevation, oceanic gateways and bathymetry, which in turn impacted on ocean and atmospheric circulation **(e.g. Dowsett et al., 2016; Herold et al. 2008; Kennedy et al., 2015, von der Heydt et al., 2016, Huang et al., 2017**), on global mean sea level changes (**e.g. Miller et al. 2020b)** and on heat transport compared to modern conditions. The Antarctic continent and its surface elevation have also evolved throughout the Cenozoic, with important consequences for ice sheet behaviour (e.g. **Colleoni et al., 2018b; Paxman et al., 2020**). Thus, direct comparison between the past and future AIS sensitivity to high levels of atmospheric greenhouse gases is not straightforward.

Most of the efforts of the PAIS programme, and its predecessor, the Antarctic Climate Evolution (ACE) programme, focused on the past warm periods (e.g. MCO, mPWP, warm interglacials of the Pleisotcene) (**Figure 1**). More specifically, ice sheets and climate simulations of the MCO emerged during the PAIS programme lifetime. Within its programme, PAIS promoted collaborative work within six specific sub-committees that addressed the following topics for almost all of the warm periods listed above:

- Palaeoclimate Records from the Antarctic Margin and Southern Ocean (PRAMSO)
- Palaeotopographic-Palaeobathymetric Reconstructions
- Subglacial Geophysics
- Ice Cores and Marine Core Synthesis
- Recent Ice Sheet Reconstruction
- Deep-Time Ice Sheet Reconstructions

Scientific advances related to each of these topics are extensively described in the previous chapters of this book. Many research projects that were initiated within the ACE programme concluded during the PAIS programme. Many of their findings continues to have a significant impact on the community of Antarctic researchers and well beyond. In the following sections, we highlight some of the key findings that have advanced our understanding of the Antarctic Ice Sheet dynamics, instabilities and thresholds during past warm periods. We conclude with a discussion of the PAIS legacy, and highlight emerging issues, knowledge gaps, needs and challenges to be addressed within the next decade by the observational and modelling communities.

2. Importance of evolving topography, bathymetry, erosion and pinning points

Ice-sheet-ocean-bedrock interactions are of major importance for understanding the dynamics of the AIS (**Mengel et al. 2014; Bart et al., 2016, Colleoni et al., 2018a; Whitehouse et al. 2019; Paxman et al., 2020**). Surface and basal boundary conditions determine the characteristics and regime of the ice flow. At the base of a terrestrial or marine-based ice sheet, the geothermal heat flux, the morphology as well as the nature of the bed (hard rock or soft sediments), affect the sliding of the ice, generate heat and yield basal meltwater. When the ice sheet advances, it erodes its bed and carries sediment. Eroded material is released into ice shelf cavities and onto the continental shelf at the grounding zone where the ice sheet floats, disconnecting from its bed. Some of this glacigenic detritus finds its way via glacial troughs and via channels across the continental slope and rise, and ultimately to the abyssal plain. Some eroded material is also carried by icebergs and deposited offshore as Iceberg Rafted Debris (IBRD). These sediments preserved in a wide range of marine environments provide a valuable archive of past ice sheet dynamics and coeval oceanic and atmospheric conditions.

The AIS substantially expanded 34 Ma and since that time has advanced and retreated numerous times (see Galeotti et al., this volume). As a result, the morphology of the bed below the ice sheet constantly evolved. Seismic stratigraphic records from the Antarctic continental margins (e.g. **Cooper et al., 1991; Eittreim et al., 1995; De Santis et al.; 1999**; **Whitehead et al., 2006; Gohl et al., 2013; Huang and Jokat., 2016**), clearly show that the shallow continental shelves have been prograding northward through time. Sediment isopach (thickness) reconstructions indicate that much of the sediments have accumulated, and accreted along the Antarctic continental slope and rise (see references in **Hochmuth and Gohl 2019; Hochmuth et al. 2020** for circum-Antarctic review and reconstructions), implying that a large volume of material has been eroded and removed from inland regions since the onset of continental glaciation (e.g. **Wilson et al., 2012; Paxman et al., 2019; Hochmuth et al., 2020)**.

The most recent circum-Antarctic reconstructions show that the morphology of the bed has evolved substantially over the past 34 million years (**Figure 2**). At the EOT, most of the West and East Antarctic sectors that are currently below sea level were instead above sea level (**Wilson et al., 2012; Paxman et al., 2019**) (**Figure 2**). With time, tectonic subsidence and erosion caused those sectors to have deepened below sea level. These reconstructions have significant implications for the understanding of the evolution of the AIS and for its ice flow. It is, indeed, much easier to grow an ice sheet on terrestrial surface than on a submarine bed. On such a restored and emergent topography, simulated Antarctic glaciations at the EOT produce a total ice volume greater than today and similar to that of the Last Glacial Maximum (LGM, ~21 ka) (**Wilson et al., 2013; Ladant et al., 2014**), even though atmospheric CO_2 levels were much higher than today, ranging from around 780 to 560 ppm (**Figure 1a**). At the EOT, the ice sheet did not expand across the continental shelves, because the ocean temperatures were too warm (e.g. **DeConto et al., 2007; Bijl et al., 2018**). In fact, geological evidence from the Antarctic Peninsula documents a faunal turnover from species adapted to temperate waters (+5°C) to species adapted to cold waters through the EOT (**Kriwet et al., 2016, Buono et al., 2019**).

Continental shelf evolution was critical for advances of the AIS across the marine realm after the EOT (e.g. **Paxman et al., 2020**). The evolution of the shallow continental shelves around Antarctica was connected to the evolution of the topography in the continent's interior. Various reconstructions (**Cooper et al., 1991**; **Eittreim et al., 1995; Brancolini 1995a,1995b; Huang and Jokat, 2016**; **Paxman et al., 2019; Hochmuth et al., 2020**) suggest that in most of the sectors, the continental shelf edge was located further south than today (**Figure 2**) and then prograded seaward over time (e.g. **Cooper et al., 1991**, **De Santis et al., 1999, Huang et al., 2014**). The stratigraphic records combined with existing Antarctic deep drilling sites suggest that the majority of the continental margin expansion occurred prior to the Pliocene (**De Santis et al., 1995, 1999, Hochmuth and Gohl 2019**), although in some sectors (e.g Amundsen Sea, **Gohl et al., 2013;** Prydz Bay, **O'Brien et al., 2007**; Wilkes Land, **Escutia et al., 2011**), progradation of the margin was still important throughout the Pliocene. The Middle to Late Miocene is a period of transition during which the Antarctic ice sheet margin advanced into a cooling ocean, grounding on the continental shelf. This is also when prominent marine-based sectors of the AIS developed (e.g., **Uenzelmann-Neben, 2019**), especially in West Antarctica (**Bart et al., 2003**). Numerical ice sheet simulations using new Antarctic Mid-Miocene paleogeographies, showed that during this period, the AIS became increasingly sensitive to oceanic conditions (**Colleoni et al., 2018b**), resulting in large glacial-interglacial changes in ice volume (**Gasson et al., 2016**). From the Pliocene onward, the Antarctic continental margin evolved very little. Erosion of the continental interior appears to have been less influential on the ice sheet since the Pliocene than during the earlier Oligocene and Miocene. The terrestrial ice sheet became more stable (**Passchier et al., 2011; McKay et al., 2012a; Gulick et al., 2017, Kim et al., 2018**). However, fluctuations of marine- based ice in deep subglacial basins still occurred, especially when atmospheric CO_2 was between 400-300ppm, during the early and middle Pliocene between 5 and 3 Ma (**Naish et al., 2009a; Pollard and DeConto, 2009; Cook et al., 2013; Cook et al., 2014; Patterson et al., 2014; Reinardy et al., 2015; Hansen et al., 2015; Bertram et al., 2018; Blackburn et al., 2020**). The relative stability of the terrestrial AIS is further supported by a recent study of cosmogenic nuclide concentrations (e.g. in-situ ^{10}Be) in a sediment core from the Ross Sea (ANDRILL Site AND-1B) and implying minimal retreat of the EAIS onto land during the last 8 million years (**Shakun et al., 2018**).

Another important aspect of the continental margin evolution is that its orientation or slope gradually changed from seaward dipping until the Early Pliocene, to landward dipping as it is now (**Cooper et al., 1991; De Santis et al., 1999**) (**Figure 3a**). This change in the bed morphology was caused by the numerous ice sheet advances and retreats and associated erosion and deposition of sediments from the bed. In turn, changes in the bed morphology then feedback on the ice sheet dynamics. Thus, since the Pliocene, the bed of the AIS marine-based sectors generally has been characterised by retrograde slopes, which favoured the potential of Marine Ice Sheet Instability (MISI) **(Jamieson et al., 2012; McKay et al., 2016; Colleoni et al., 2018b**). Prior to the Late Miocene, climatic conditions were generally warm, the continental shelves were less expanded in most of the Antarctic sectors and did not present strong retrograde slope. Consequently, the ice sheet could retreat relatively easily during warm climate episodes with strong surface melt and oceanic melt (e.g. **Levy et al., 2016, Gasson et al 2016**). At the end of the Miocene, climate gradually cooled, which favoured terrestrial EAIS stability, but concurrently, the retrograde slope of the bed favoured instabilities and fast AIS grounding line retreat in the marine-based sectors during phases of prolonged or exceptional warmth (**Cook et al., 2013, Pollard and DeConto 2009; Naish et al., 2009a**; **DeConto et al., 2012,; Pollard et al., 2015; DeConto & Pollard, 2016; Golledge et al. 2017a, Colleoni et al., 2018b; Levy et al., 2019; Blackburn et al., 2020**).

Fast retreat of the grounding line can, however, be slowed down or stopped by the occurrence of pinning points at the bed that provide a buttressing backstress that resists seaward ice flow (**Mengel and Levermann, 2014**). Pinning points or pinning areas can take different forms. Ice rises for example, form when an ice shelf anchors on a pre-existing bathymetric high, stabilizing the flow (**Matsuoka et al., 2015**) (**Figure 3b**). They can be tectonic structures or volcanic islands. Today, several ice rises are visible from the surface, for example, in the Ross Sea embayment (e.g. Roosevelt Island, Crary Ice Rise, Franklyn Islands and Ross Island) and in the Weddell Sea embayment (Berkner Island). **Halberstadt et al. (2016) and Simkins et al. (2018)** suggested that the retreat of the grounding line during the last deglaciation was slower in the western Ross Sea than in the eastern Ross Sea, which is characterised by a smoother bed.

Pinning points can also form temporarily due to the uplift of the bed as a result of glacio-isostatic adjustment (GIA) in ice sheet retreat and unloading of the crust (e.g. Whitehouse et al., 2018; **Figure 3c**). Numerical simulations show that ignoring GIA during ice sheet advance results in smaller, less extended ice sheets, than would occur if GIA was accounted for, because GIA creates pinning opportunities (e.g., **Colleoni et al., 2018b**). **Kingslake et al. (2018)** showed that during the early Holocene, the West Antarctic Ice Sheet (WAIS) in both the Weddell Sea and the Ross Sea temporarily retreated beyond its present-day grounding line position. It subsequently re-advanced potentially due to uplift of the bed due to GIA, (**Bradley et al., 2015**) and the occurrence of relief (e.g. Bungenstock Ice Rise, Weddell Sea) on which the ice shelf could pin. Similarly, high-resolution bathymetry acquired from a ridge under the Pine Island Glacier Ice Shelf has revealed geomorphological features that may be consistent with a retreat of Pine Island Glacier inland from its present position earlier during the Holocene and a subsequent re-advance to its early 20th century position (**Graham et al., 2013).** These pinning points can be subsequently eroded or can simply “resorb” after glacio-isostatic adjustment of the bed.

Finally, the ice sheet can build its own pinning points by accumulating sediments in grounding zone wedges (GZW) during deglaciations, which slows its retreat (e.g. Alley et al., 2007; Horgan et al., 2013; **Figure 3d**). An example of this effect was outlined in **Bart et al. (2017, 2018)**, who analysed a complex of GZWs that formed during the last deglaciation in the Whales Deep basin (Eastern Ross Sea). Proxy analyses and dating of sediment cores revealed that the first four GZWs were built during the first ~5000 years of a gradual 75-km southward ice sheet retreat, between ~17 and ~12.3 ka. They were characterised by low sedimentation rates (i.e., the GZWs are very thin) and sediment compositions indicate that the grounding line was pinned on those GZWs, whilst an extensive ice shelf formed during the retreat. The last three GZWs accumulated, with a clear aggradation sequence, in about 800 years between 12.3 ka and 11.5 ka, implying that the grounding line was not retreating during this brief interval. These GZWs were characterised by very high sedimentation rates and were thus relatively thick compared to the older ones. Sediment compositions seaward of those three GZWs indicate that the ice shelf there broke up at the very beginning of this time interval and never reformed, and that the grounding line remained pinned successively on top of those last three GZWs. After building the uppermost GZW, the grounding line stepped back by about 100 km within a few decades, resulting in a very brief, massive ice discharge of about 0.1 mm Sea Level Equivalent (SLE, **Bart and Tulaczyk, 2020**). Similar mechanisms and sequences have been inferred from the analysis of Pine Island Bay continental shelf multibeam and marine seismic data from the Amundsen Sea Embayment shelf (**Uenzelmann-Neben et al., 2007; Jakobsson et al., 2011; 2012; Klages et al., 2015**). Data show that the ice stream retreat was paused due to the built of GZWs during the last deglaciation. GZWs can also build as a consequence of ice sheet retreat in a narrow trough in which lateral edges serve as a pinning zone that slowdown the retreat. **Livingstone et al. (2013) and Jamieson et al. (2012, 2014)** mapped a series of GZWs in a paleo-ice stream trough in Marguerite Bay (Antarctic Peninsula). Numerical simulations have shown that ice stream retreat rates slowed as the grounding line passed the laterally narrow parts of the trough. If a constant sedimentation rate was assumed, GZWs could form, thus further slowing the retreat.

The presence or absence of pinning points influences ice sheet dynamics. The estimated rates of sea level change during past periods may have been affected by potential pauses or changes in the rate of AIS (or other ice sheets) advances or retreats, due to ice-bed interactions. These variations may become highly relevant, especially for the interpretation of sea level reconstructions since the LGM, or simulated ice volume changes at the sub-millennial scale (**Klages et al., 2017; Bart et al., 2018, Kingslake et al., 2018**). However, the erosion of potential paleo-pinning points, during the numerous phases of expansion of the AIS, makes it difficult to know the role of pinning points on past ice sheet variability at sub-millennial time scales in reconstructions older than the LGM. Therefore, pinning points are generally not resolved in paleo-ice sheet simulations, where bed topography cannot be reconstructed with sufficient accuracy or resolution (<5 km). Moreover, shallow ice approximation ice sheet models frequently used for paleo-ice sheet simulations need a smoothing of the bed morphology in order to enable numerical convergence in areas where the morphology is too steep for the applicability of the hydrostatic approximation. For example, the Parallel Ice Sheet Model (PISM, **Bueler and Brown, 2009**) proposes different levels of smoothing of the bed, and this model is frequently used within the PISM paleo-community (**e.g. Golledge et al., 2012, Albrecht et al., 2020**). Full-Stokes ice sheet models, in which no hydrostatic approximation is applied, do not require bed smoothing since the physics account for both horizontal and vertical shear of the ice flow. However, full-Stokes models are too computationally demanding and are still not usable for most paleoclimate applications (**Colleoni et al., 2018a** and references therein). Given that most paleogeographic reconstructions (e.g. **Paxman et al., 2019; Hochmuth et al., 2020**) are very coarse in spatial resolution and highly uncertain in terms of detail, bed smoothing in ice sheet simulations resulting in the loss of local pinning points is generally of lesser importance than the biased controls on ice sheet dynamics induced by uncertain bed morphologies (e.g. **Gasson et al., 2015**).

3. Reconstructions of Southern Ocean sea and air surface temperature gradients

Equator-to-pole surface temperature gradients influence Earth's latitudinal heat distribution. Reconstructions of meridional temperature gradients since the Late Cretaceous clearly show a gradual steepening during the transition from greenhouse to icehouse conditions as the polar regions cooled and ice sheets developed (e.g. **Zhang et al, 2019**). Reconstructions also show the emergence of oceanic fronts in the sub-tropics and high latitudes, especially in the Northern Hemisphere (e.g. **Zhang et al, 2019**). The development of the Southern Ocean frontal system is of importance for reconstructing past AIS. The Antarctic Polar Front (APF) is a region marked by elevated current speeds and strong horizontal gradients in seawater density, temperature, salinity. It is currently located at approximately 50°S in the Atlantic and Indian sectors, and around 60°S in the Pacific sector. During warm periods, proxies imply a substantial southward shift of the Antarctic Polar Front (APF) associated with a significant reduction in sea ice extent (**e.g. Taylor-Silva and Riesselman, 2018; Bijl et al. 2018; Sangiorgi et al. 2018; Salabarnada et al., 2018; Chadwick et al., 2020; Evangelinos et al., 2020**). Conversely, during cold periods, the Antarctic Polar Front shifts northward accompanied by a large expansion of the sea ice cover (**Gersonde et al., 2005**; **Kemp et al. 2010; McKay et al., 2012a**).

Fewer sediment cores have been recovered in the Southern Ocean high latitudes than in the North. South of 50°S, sea surface temperature (SST) proxy reconstructions are rare (**Figure 4**). The ACE programme and more recently, the PAIS programme increased the number of

SST and Sea Water Temperature (SWT) at 0-200 m depth records from the Southern Ocean's Antarctic margin. SST and SWT records are now available for the MCO from the continental shelf site ANDRILL AND-2A (Western Ross Sea, **Levy et al., 2016**) and from the continental rise at Integrated Ocean Drilling Program (IODP) Site U1356 (Adélie Land margin, **Sangiorgi et al., 2018; Hartman et al. 2018**). Mid- to late Pliocene SST records are available from ANDRILL-1B sediment core (Ross Sea shelf, **McKay et al., 2012a**) and from other cores on the continental rise and abyssal plains from the Indian sectors (see **Dowsett et al., 2013** for an SST compilation). MIS 31 SST records are available from continental rise sites such as Ocean Driling Program (ODP) Site 1101 (Antarctic Peninsula, **Beltran et al., 2020**) and IODP Site U1361 (Adélie Land margin, **Beltran et al., 2020**) and from site ODP Site 1094, (south of APF, South Atlantic sector, **Beltran et al., 2020**). For MIS 11 and MIS 5e, no Antarctic continental margins SST records are available so far. However, recent International Ocean Discovery Program (IODP) Expedition 374 to the Ross Sea (**McKay et al., 2019**), IODP Expedition 379 to the Amundsen Sea (**Gohl et al., 2019**), and expedition INS2017_V01 on the Sabrina Coast (**Armand et al., 2018; O'Brien et al., 2020)** have recovered highly expanded sedimentary sections from the continental rise. This promises upcoming high-resolution SST records for Pleistocene interglacials, for which most of the ice proximal SST information is still missing.

A comparison between MCO and mPWP global meridional proxy-based SST gradients highlights the difference between the two periods in the Southern Hemisphere and how much the global climate state has evolved between 17 and 3 Ma (**Figure 4a**). During the MCO, the air surface temperature gradient strengthened between 30°S to 40°S, as in the Northern Hemisphere, suggesting that sub-tropical marine frontal system was well developed. The meridional SST and SWT gradients were much weaker than today and a summer warming of 16°C to 22°C (± 5°C) compared to today, was observed in geochemical proxies at around 60-65°S on the East Antarctic continental rise (**Sangiorgi et al. 2018; Hartman et al., 2018**) and a warming of about 2°C to 12°C (± 5°C) compared to today is recorded in the Ross Sea (**Levy et al., 2016, Sangiorgi et al., 2018**) indicating a total absence of or rare occurrences of sea ice during this period. The mPWP was characterised by a meridional SST gradient weaker than today (e.g. **Brierley et al., 2009; Haywood et al., 2013**) and with a significant Arctic amplification, while the warming anomaly was more subdued in the high southern latitudes (**Figure 4a**). South of 55°S, East Antarctic continental rise summer SST were warmer than today by about 4°C to 6°C on the East Antarctic continental rise and by up to 7°C on the Ross Sea shelf (**McKay et al., 2012a**), indicative of a highly reduced, if not absent, summer sea ice cover in some sectors.

During both the MCO and the mPWP, the APF was probably more contracted towards high latitudes in all sectors around Antarctica (**Taylor-Silva and Riesselman, 2018; Sangiorgi et al., 2018**). The mPWP presents a meridional SST gradient steeper than during the MCO and it is possible that the APF might not have reached latitudes as poleward as during the MCO. The contrast between the mPWP and MCO meridional SST gradient south of 40°S, suggests that the gradient probably steepened during the Late Miocene (**Herbert et al., 2016**). However, in some sectors of the Antarctic margin, there is strong evidence for warmer conditions during the Early Pliocene compared to the mPWP, i.e. sea-ice reduction and warming west of the Antarctic Peninsula (e.g. **Hillenbrand & Ehrmann 2005**; **Escutia et al. 2009**) and in Prydz Bay (**Whitehead et al. 2005**). This was accompanied by a poleward shift of the APF (**Bart & Iwai 2012**; **Whithehead & Bohaty 2003; Escutia et al. 2009**).

During the Pleistocene, MIS 31, MIS 11 and MIS 5e meridional SST gradients highlight the strong impact of precessional astronomical forcing, with SSTs warmer than today especially in the Northern Hemisphere mid-to-high latitudes (**Figure 4b**). All three interglacials present SST gradients quite similar to today in the equatorial to sub-tropical latitudinal bands. The discrepancies between them emerge for latitudes poleward of 50°S. For MIS 31, alkenone and long chain diol analysis on sediment cores revealed an SST warming of 4°C to 12°C on the Adélie Land margins and the Antarctic Peninsula (**Beltran et al., 2020**). Proxies suggest reduced or even absent winter and summer sea ice in the Ross Sea and offshore Adélie Land margin; information is missing for other Antarctic sectors (see references for **Figure 6d**). The abrupt appearance of foraminiferal oozes and bioclastic limestone in the Ross Sea and coccolith-bearing sediments in Prydz Bay during MIS31 (**Bohaty et al., 1998; Scherer et al., 2003; Villa et al., 2008, 2012**) indicates a significant southward migration of the APF. In terms of the SST gradient, MIS 31 shows more similarities with the mPWP in the Southern Hemisphere than with the more recent Late Pleistocene interglacials.

No SST or SWT proxies south of 60°S are available for MIS 11, thus it is difficult to assess the magnitude of a potential warming closer to the Antarctic margins. Interpretation of diatom and geochemical changes could help to estimate SST or SWT from the Wilkes Subglacial Basin and the Ross Sea margins (see Wilson et al., this volume for references). North of 40°S, tropical to subtropical MIS 11 SSTs were warmer than modern by about 1-4°C, and between 50°S and 60°S, only a 1-2°C warming above modern is recorded (**Kunz-Pirrung et al., 2002**). Apart from just a single exception, no SST reconstructions are available for MIS 5e south of 60°S (**Capron et al., 2014, Hoffman et al., 2017; Chadwick et al., 2020**), and the reconstructed SSTs north of 60°S are similar to those of MIS 11 (**Figure 4b**). It is hence difficult to assess the magnitude of surface and ice proximal sub-surface ocean warming during MIS 5e. **Capron et al., (2014)** report warmer than present-day conditions that occurred for a longer time interval in southern high latitudes than in northern high latitudes. They also report an earlier MIS5e warming in the Southern Ocean starting from 130 ka compared with the Northern high latitudes and synchronous with Antarctic ice core records. Moreover, **Chadwick et al. (2020)** showed that the sea-ice minima and SST maxima were reached at slightly different times in three Southern Ocean sectors. During both MIS 11 and MIS 5e, the few existing records indicate a seasonally sea ice covered ocean (**Kunz-Pirrung et al., 2002; Wolff et al., 2006 Escutia et al., 2011; Wilson et al., 2018, Chadwick et al., 2020**). Ice core analyses by **Wolff et al. (2006)** suggested on the basis of sea-ice proxies in the EPICA Dome C ice core that winter sea ice was largely reduced during MIS 5e and MIS 11 in the Indian sector of Antarctica, and that summer sea ice was likely absent. In addition, similar proxy analyses on the EPICA DML ice core from Dronning Maud Land indicate a sea ice reduction in the Atlantic sector during MIS 5e (**Schüpbach et al., 2013**).

4. Extent of major Antarctic glaciations

This section focuses on glaciations that occurred during the Eocene-Oligocene Transition (EOT, ~34 Ma) (see Galeotti et al., this volume), during the Mid-Miocene Climatic Transition (MMCT, 14.5 to 13.5 Ma) (see Levy et al., this volume), the M2 glaciation (3.312-3.264 Ma) preceding the mPWP, and the Last Glacial Maximum (LGM, ~21 ka) (see Siegert et al., this volume).

The EOT was characterised by the development of a continental ice sheet on Antarctica (**Barrett, 1989; Hambrey et al. 1991; Wise et al., 1991; Zachos et al. 1992**) as atmospheric CO_2 level fell (e.g **DeConto et al., 2003**) and the Southern Ocean cooled (e.g., **Bijl et al., 2013**) as a result of the opening of ocean gateways. (e.g. **Kennett et al., 1977**) (**Figure 1**). Across the EOT, deep-see temperatures cooled by 3° to 5°C (e.g. **Liu et al., 2018**) as a consequence of decreasing CO_2 levels (**Pagani et al., 2005)**. Sedimentary cycles from a drill core in the western Ross Sea provided the first direct evidence of orbitally controlled glacial cycles between 34 million and 31 million years ago (**Galeotti et al., 2016**). Initially, under atmospheric CO_2 levels of ≥600 ppm, a smaller AIS, restricted to the terrestrial continent, was highly responsive to local insolation forcing. The establishment of the Antarctic Ice Sheet (AIS) is associated with an approximately +1.5 per mil increase in deep-water marine oxygen isotope values ($\delta^{18}O$) beginning at ~34 million years ago (Ma) and peaking at ~33.6 Ma (**Coxall et al., 2005; Bohaty et al., 2012**), with two positive $\delta^{18}O$ steps separated by ~200,000 years (Figure 1b). The first positive step in the isotope data primarily reflects a temperature decrease (**Lear et al., 2008**) (EOT-1, ~34.46-33.9 Ma); the second one has been interpreted as the onset of a prolonged interval of maximum ice extent at ~33.6 -33.7 Ma (EOT-2) (**Liu et al., 2009**). Stratigraphic unconformities identified from the continental margins of Antarctica (**Figure 1b**, e.g. **De Santis et al., 1995; Eittreim et al., 1995; Cooper and O'Brien 2004; Escutia et al., 2005; Whitehead et al., 2006; Gohl et al., 2013; Uenzelmann-Neben and Gohl; 2014, Gulick et al., 2017)** presumably correspond to one of these $\delta^{18}O$ excursions.

Galeotti et al. (2016) suggest that **a** continental-scale AIS with frequent calving at the coastline did not form until ~32.8 million years ago, coincident with the earliest time when atmospheric CO_2 levels fell below ~600 ppm. The atmospheric CO_2 threshold for the onset of large-scale Antarctic glaciations remains, however, uncertain and varies between 900 ppm to 560 ppm in numerical climate and ice sheet simulations (**DeConto et al. 2003; Ladant et al. 2014, Liakka et al., 2014, Gasson et al, 2014). Liakka et al., (2014)** showed that when accounting for vegetation-albedo feedbacks, large-scale Antarctic glaciations occurred when atmospheric CO_2 dropped between 1120 ppm and 560 ppm. **Ladant et al. (2014)** simulated a first Antarctic expansion at EOT-1 associated with a first sea level drop of about 10 meters (atmospheric CO_2 set to 900 ppm) and a second one coinciding with early Oligocene glaciation Oi-1 (33.4 to 33.0 Ma, **Miller et al.,1991; Zachos et al., 2005**) of about 63 meters (atmospheric CO_2 set to 700 ppm). Sequence boundary and ice volume proxies suggest that the extent of the AIS gradually increased across the EOT and expanded to either near-modern dimensions (**Miller et al., 2008; 2020a**) or as much as 25% larger than at present day (**Katz et al., 2008; Wilson et al., 2013**). Numerical ice sheet modelling studies show a large range of ice volumes across the EOT. Simulated glaciations lead to sea level fall clustered around 10 m SLE and 25 m SLE relative to present AIS volume (**DeConto and Pollard 2003; Pollard and DeConto 2005; Gasson et al., 2014; Ladant et al, 2014; Liakka et al, 2014; Wilson et al., 2013**) (**Figure 1d**), which corresponds to a total simulated AIS volume up to 83 m SLE. This is broadly in agreement with sea level falls up to 70 m estimated from low-latitude shallow-marine sequences (e.g. **Cramer et al., 2011**) (**Figure 1c**). Both **DeConto and Pollard (2003)** and **Ladant et al. (2014)** simulated isolated big ice caps over East Antarctic highlands presumably during EOT-1, that ultimately coalesced during Oi-1 (**Figure 5a**). This is supported by the geological evidence of glacimarine deposits in the Wilkes Land continental shelf and rise since 33.6 Ma (**Escutia et al., 2005, 2011**) and by glacial sediment transport to the continental slope of the Prydz Bay margin since ~35 Ma (**O'Brien et al, 2004**). No ice grounded on the continental shelf at this time (**Figure 5a**, e.g. **Barrett et al., 1989, 2007**), and the continental shelf edge was located farther South than present for most of the Antarctic margins (**Figure 2**).

The Mid-Miocene Climatic Transition (MMCT, ~14.8 - 13.5 Ma) is a period of global cooling following the extreme warmth of the Mid-Miocene Climatic Optimum (MCO). The onset of global climatic cooling at ~14.8 Ma marks the start of the MMCT (**Böhme, 2003; Flower and Kennett, 1993; Holbourn et al., 2014; Shevenell et al., 2008**). Disconformities in the ANDRILL AND-2A record (**Levy et al., 2016**) and across the Ross Sea (**De Santis et al., 1999**) **(De Santis et al., 1999**), pulsed deposition of ice-rafted debris offshore Prydz Bay and the Adélie Land margin (**Pierce et al., 2017**) together with an increase in sea ice indicators in the Ross Sea and off East Antarctica (**Levy et al., 2016; Sangiorgi et al., 2018**) and major turnover in Southern Ocean diatom species (**Crampton et al., 2016**), suggest marine ice sheet advances across the Ross Sea during glacial intervals for the first time since the onset of the MCO (**Figure 5b**). Ice sheet advance in the Wilkes Land margin is recorded by erosion of older sediments from the shelf (**Escutia et al., 2011**) and an increase in dinocyst assemblages from the seasonal sea ice zone south of the Antarctic Polar Front (**Sangiorgi et al., 2018**). Additionally, less well-dated erosional unconformities in the Weddell Sea (e.g. **Huang et al., 2014**), Amundsen Sea (e.g. **Lindeque et al., 2016; Uenzelmann-Neben and Gohl, 2012**), Bellingshausen Sea-Antarctic Peninsula (e.g. **Rebesco et al., 2006; Uenzelmann-Neben,**

2006), Sabrina Coast (**Gulick et al., 2017**) and in Prydz Bay (e.g. **Whitehead et al., 2006**), are attributed to MMCT marine-based ice expansion, and together imply that both the EAIS and the WAIS expanded onto the continental shelf at this time. An increase in glacial-interglacial amplitude in the far-field $\delta^{18}O$ data suggests that the AIS expanded further during successively, gradually colder glacial phases. This interval of increased glacial expansion culminated in a major step in the $\delta^{18}O$ record at 13.9 Ma (**Figure 1b**). During the MMCT, Southern Ocean SSTs cooled by about 6 °C (**Holbourn et al., 2007; Sangiorgi et al., 2008**). Bottom water temperatures generally cooled by 2 to 3 °C (**Cramer et al., 2011; Lear et al., 2015; Shevenell et al., 2008**) and global sea level may have dropped by as much as 50 m (**Miller et al., 2020a**), hinting at the possibility of some ice expansion in the Northern Hemisphere (**DeConto et al., 2008**). Summer temperatures in the Trans-Antarctic Mountains declined by >8 °C (**Denton and Sugden, 2005**) and this cooling has been linked with a shift from temperate climate wet-based glaciation with a dynamic ice sheet in a warm to temperature climate to a predominantly dry glaciation style with a more stable terrestrial ice sheet under moder-like Antarctic polar climatic conditions (**Lewis et al., 2008; Lewis and Ashworth, 2016; Sugden and Denton, 2004**).

The largest benthic foraminifera $\delta^{18}O$ shift during this period is of about 1.3 ‰ (e.g. **Holbourn et al., 2013**), partly corresponding to an estimated a sea level drop of 35-40 m from interglacial to glacial. Interestingly, numerical ice sheet models can only simulate such a large interglacial-to-glacial amplitude in ice volume (**Gasson et al., 2016, Colleoni et al., 2018a**), when using a reconstructed Mid-Miocene paleogeography (**e.g. Paxman et al., 2019**). Backstripped sea level data (**Miller et al., 2005, Kominz et al., 2008**) and calibrated benthic $\delta^{18}O$ sea level changes (**Miller et al., 2020a)** revealed potential sea level falls up to 10 to 20 meters below the present-day mean sea level during the MMCT (**Figure 1c**), implying a greatly expanded AIS, perhaps up to 30% larger than today. The compilation of simulated Antarctic ice volume contributions to global mean sea level for this cold period ranges between +10 to -20 m SLE relative to present (**Gasson et al., 2016, Colleoni et al., 2018a**) (**Figure 1d**, cyan squares).

The Mid-Pliocene M2 glaciation (**~3.312-3.264 Ma**) corresponds to a large transient increase in the deep-sea benthic $\delta^{18}O$ records (**Figure 1b**) with a cooling of at least 3.5°C preceding the peak of M2 (**Karas et al., 2020**) and an atmospheric CO_2 level drop of about 320-343 ppm (**de la Vega et al., 2020**). A compilation of climate proxy data suggests that during this glaciation, Greenland mountain glaciers expanded (**Thiede et al., 2011; Jensen et al., 2000**) and that other ice caps also grew in the Northern Hemisphere (**De Schepper et al., 2013; Tan et al., 2017**). In fact, this glaciation marks the end of global warmth of the early Pliocene (5.5 to 3.3 Ma) and the beginning of a step-wise transition towards bipolar cooling that culminated in continental-scale Northern Hemisphere glaciations ~2.7 Ma. Numerous sedimentary hiatuses, including the M2 glaciation, are observed in the AND-1B sediment record (western Ross Sea, **Naish et al., 2009a**) during the Mid to Late Pliocene. Continental margin morphology appears to have allowed the AIS to advance to the shelf edge during the M2 glaciation in the Ross Sea (e.g. **Kim et al, 2018, McKay et al., 2012a; McKay et al. 2019**), Prydz Bay (**O'Brien et al., 2007),** Sabrina Coast **(Gulick et al., 2017**), Wilkes Land (**Eittreim et al., 1995**; **Escutia et al., 1997**; **De Santis et al., 2003**), and the Antarctic Peninsula and Amundsen Sea (**Rebesco et al., 2006, Gohl et al., 2013**). The glacio-eustatic sea level drop of this period is represented by a major erosional sequence boundary on the New Jersey shelf (**Miller et al., 2005**) and global sea level fall of about 30 m at the M2 glaciation (**Naish and Wilson, 2009; Miller et al., 2012; Grant et al., 2019; Miller et al., 2020a**) (**Figure 1c**). Based on the review of circum-Antarctic evidence of grounding events and sequence stratigraphy, **Bart (2001)** suggested indeed, an Antarctic ice volume larger than today and almost as large as during the LGM, implying the existence of relatively small Northern Hemisphere ice sheets. In contrast, the compilation of simulated Antarctic ice volume, however, likely underestimates the ice sheet expansion during this glaciation (**Figure 1d**) and, instead, yields a global mean sea level rise up to 5 m above present-day mean sea level (**Pollard and DeConto, 2009; Tan et al., 2017; De Boer et al., 2017a).**

The Last Glacial Maximum (LGM, ~19-23 ka) is the most recent glaciation that occurred before present, and as such, is the best documented glaciation. Ice core records have shown that the atmospheric CO_2 levels dropped to about 185 ppm (**Lüthi et al., 2008**). Global climatic reconstructions revealed that the global mean temperature dropped by about 4-7 °C (e.g**. Schneider von Deimling et al., 2006; Tierney et al., 2020**) compared to present. Marine benthic and planktic foraminifera recorded a clear δ^{18}O increase (**Imbrie et al., 1984, Lisiecki and Raymo, 2005),** actually, observable concomitantly to all past cold glacial periods (**Imbrie et al., 1984, SPECMAP stack; Lisiecki and Raymo 2005, LR04 stack; Westerhold et al., 2020, CENOGRID; Zachos et al., 2008**, **Figure 1b**). Calibrated conversions of the δ^{18}O record or dated paleo coral reefs suggest a global mean sea level drop ranging between 80 m to 130 m relative to present (**Figure 1c**) (e.g. **Waelbroeck et al., 2009**; **Bard et al., 1990; Shackleton, 2000**), with a cluster between 110 m and 130 m below present-day mean sea level. Diatoms and radiolarians show that the sea ice cover was larger than today. Winter sea ice cover shifted northward by about 5 to 10° (e.g. **Gersonde et al., 2005**; **Benz et al., 2016**) and summer sea ice edge, although more uncertain, might have been located around 60.5°S (**Green et al., 2020** and references therein). Compilation of climate proxies suggest a potential strengthening and equatorward shift of the Southern Hemisphere westerlies (e.g. **Kohfield et al., 2014; Lamy et al., 2014; Struve et al., 2020**), which remains debated (e.g. **Kim et al., 2017**; **Sime et al., 2016; Lamy et al., 2019**).

There is persuasive evidence from the geological record to indicate that the AIS was larger than present around the time of the global sea level lowstand at ~20 ka, although the extent of this expansion is well constrained at only a few sites around the continental margin (**Whitehouse, 2018**). Both marine and terrestrial geological data indicate that at the LGM, the AIS almost extended to the continental-shelf break in most sectors (**Eittreim et al., 1995**; **Anderson et al., 2002, 2014; Hillenbrand et al., 2012, 2014; The RAISED Consortium, 2014; Mackintosh et al., 2014; Arndt et al., 2017; Bart et al., 2018**) (**Figure 5c**), as during many previous Pleistocene glaciations (e.g. **Escutia et al., 2003**). **However, the AIS** did not advanced up to the continental shelf edge in Prydz Bay (**O'Brien et al., 2007; Mackintosh et al., 2014; Wu et al., 2021,** in the Western Ross Sea (**Halberstadt et al., 2016; Prothro et al., 2018**) and in parts of the Amundsen Sea (e.g. **Larter et al., 2014; Klages et al., 2017**). Furthermore, the scenario for ice advance in the Weddell Sea embayment remains uncertain (**The RAISED consortium, 2014; Whitehouse et al., 2017; Nichols et al. 2019**). Ice sheet expansion during the LGM led to a thickening of the AIS of several hundreds of meters almost in all sectors, especially around West Antarctica as supported by exposure data (**see** Siegert et al., this volume). On the Antarctic plateau, ice core δ^{18}O isotopes records suggest that elevation increased of 270 to 660 meters between the LGM and present-day (**Werner et al., 2018**). Over West Antarctica, the increase in elevation during the LGM is up to 850 to 1800 meters (**Werner et al., 2018**).

The relatively small number of proximal geological records on AIS extent and thickness during the LGM prevents an accurate constraint on LGM ice volume. Distal, deep ocean benthic foraminifera δ^{18}O records may provide overall ice volume estimates, but do not allow disentangling contributions from individual contribution of each ice sheets at the LGM (**Simms et al., 2019**, and references therein; **Clark and Tarasov, 2014**). Ice sheet modeling is one of the possible approaches to simulate the volume of the AIS at the LGM. Such modeling has yielded an increase in ice volume of 5.9 to 19.2 m of sea level equivalent (SLE) (**Bentley, 1999; Huybrechts, 2002**) in the late 1990 and early 2000s. With the improvement of ice sheet models and climate forcing, the range of AIS contributions to sea level change at LGM has narrowed to about -5 to -12 m SLE (e.g. **Huybrechts, 2002; Golledge et al., 2012; Gomez et al., 2013; Maris et al., 2014; Briggs et al., 2014; Quiquet et al., 2018, Sutter et al., 2019**), with a cluster around -7 to -8 m SLE (**Figure 1d** and references therein). Another approach is to inverse AIS by means of glacial-hydro isostatic adjustment (GIA) models, which describe the viscous response of the solid Earth to past changes in surface loading by ice and water

(**Whitehouse et al., 2018**). This approach has also been used in combination with direct ice sheet modeling (e.g. **Whitehouse et al., 2012b**) and/or by making use of constraints on ice thickness from reconstructions based on exposure age dating, as well as satellite observations of current uplift (**Whitehouse et al., 2012b; Ivins et al., 2013; Argus et al., 2014**). Estimates from GIA modelling for the AIS contribution to global mean sea level amount to -5 to -30 m SLE with most of the contributions smaller than -13 m SLE (**Figure 1d**). Older studies had estimated large sea level contributions generally above 15 m (e.g. **Nakada et al.,2000; Huybrechts, 2002; Peltier and Fairbanks, 2006; Philippon et al., 2006; Bassett et al., 2007**), but more recent modeling studies and reconstructions have refined these estimates to below 13.5 m (**Mackintosh et al., 2011; Whitehouse et al.,2012a; Gomez et al., 2013; Argus et al., 2014; Briggs et al., 2014**) with an average contribution of about -10 m SLE (e.g. **Simms et al., 2019** and references therein).

Despite those improvements, AIS contributions to sea level changes at the LGM remains poorly constrained (**Simms et al., 2019**, and references therein, **Clark and Tarasov, 2014**) and this has global consequences on the assessment of past, present and future sea level changes. Land ice retreat in both hemispheres during the last deglaciation have produced a residual GIA signal that still affects present-day sea level changes measurements **(Martín-Español et al., 2016)**. This residual signal is estimated from modelled reconstructions of global land ice thickness changes, and spatio-temporal deglaciation history (e.g. ICE-5G to ICE 7G, GLAC-1, ANU; **Peltier et al., 2004; Peltier et al., 2015; Roy and Peltier, 2018; Tarasov and Peltier, 2002, 2003; Lambeck and Chappell, 2001; Lambeck et al., 2002)**. To date, there is no consensus on AIS volumes at the LGM and through the last deglaciation, A compilation of cosmogenic exposure ages from low-elevation sites shows that the AIS substantially thinned throughout the Holocene, but mainly after the MWP-1A (**Small et al., 2019**). **The RAISED consortium (2014)** provided partial pan-Antarctic grounding line position at ~15 ka, 10ka and 5ka. In the ice sheets deglaciation scenarios ICE-6G and ICE-7G (**Peltier et al., 2015; Roy and Peltier, 2018**), the AIS extent at LGM has been set up to its present-day extent, but with grounded ice filling the embayments currently occupied by the Ross Ice Shelf, by the Ronne-Filchner Ice Shelf and by the Amery ice shelf. This surely affects the calculation of GIA and its residual signal and, as such, the assessment of post-glacial and present-day land ice contribution to on-going sea level changes (**Martín-Español et al., 2016**).

At regional scale, the inclusion of realistic, spatially variable relative sea-level forcing through coupled simulations of 3-D ice-sheet and GIA-modulated sea-level change results in a stabilising effect on marine-grounded ice sheet dynamics (**Gomez et al., 2010**). The grounding line, in fact, advances or retreats in response to the regional ice fluctuation. The latter triggers viscoelastic solid Earth rebound as well as a change of the local geoid height in response to the variation of the gravitational pull (e.g. **Stocchi et al. 2013**). In particular, the predicted increase in the volume of the WAIS during the last glacial cycle is smaller in the coupled simulations due to negative feedbacks associated with an increase in near-field water depth. The latter stems for the combination of ice-driven solid Earth subsidence and counterintuitive local sea-level rise caused the gravitational attraction of the growing ice sheet's mass (**Gomez et al., 2013; De Boer et al., 2014b, 2017b; Konrad et al., 2014**). At global scale, **Gomez et al. (2020)** showed that the retreat of Northern Hemisphere ice sheets during the last deglaciation and associated sea level rise directly impacts on the dynamical behaviour of the AIS and conditions its own retreat. Modelled AIS sensitivity on different paleobathymetries since the Mid-Miocene shows that the position of the AIS advance on the continental shelf depends on glacio-isostatic adjustment (generating pinning points, see **Section 2**) and on the magnitude of global mean sea level changes (**Colleoni et al., 2018b; Paxman et al., 2020)**. A similar relationship between the AIS stability and global mean sea level changes has recently been inferred from a North Atlantic deep-ocean benthic δ^{18}O record of the Plio-Pleistocene Transition and the early Pleistocene (**Jakob et al., 2020**). Based on a range of Mg/Ca paleothermometer calibrations, the sea level record suggests that the gradual expansion of the Northern Hemisphere ice sheets, and the consequent substantial lowering of global mean sea level, led to an increasing stability of the terrestrial EAIS (**Jakob et al., 2020**). Other studies also highlight the sensitivity of the marine-based sectors of the AIS to rapid sea level rise at millennial to sub-millennial time-scales, such as the impact of rapid sea level rise during the various meltwater pulses episodes of past deglaciations (e.g. **Golledge et al., 2014; Petrini et al., 2018; Turney et al., 2020**).

5. Antarctic Ice Sheet response to past climate warmings

Assessing the AIS behaviour during past periods warmer than today can inform on the magnitude and timing of past and future sea level changes, as well as on various mechanisms triggering ice sheet retreats that can vary through time (i.e. atmospheric and/or oceanic warming). At millennial to sub-millennial scales, the crossing of tipping points caused by Earth's climate system feedbacks can cause rapid ice sheet retreats. One example is ocean warming triggering MISI. Because the global climatic state has been constantly evolving, the conditions necessary to cross these tipping points have also evolved as well. To highlight this aspect, several policy-relevant warm periods in the geologic past have been analysed, based on a few climatic and glaciological indicators synthesized in **Figure 6**. Note that except for the sea ice (**Figure 6d**), all the other variables are expressed as anomaly relative to their present-day value (20^{th} century for MAT and SST).

The compilation of global mean sea level changes and simulated Antarctic contributions is exhaustive and illustrates a key focus of the paleo polar community has been producing over recent decades. We do not discard computed estimates of AIS contribution to global mean sea level change that could appear out of the range of data. Instead, we consider such values as part of the uncertainties associated with uncertain models physics and boundary conditions. References for all compiled data and simulated ice volumes are provided in the caption of **Figure 1**.

The Mid-Miocene Climatic Optimum (MCO, 17-14.8 Ma). The MCO presents an interesting analogue for assessment of climate projected for the next decades to centuries (**Steinthorsdottir et al., 2020**). At that time, Antarctica hosted the only existing continental-size ice sheet. Geological proxy data indicate atmospheric CO_2 concentrations generally varied between 300 ppm and 600 ppm on glacial-interglacial (orbital) time scales during much of the MCO (**Foster et al., 2012; Greenop et al., 2014**), but it may have reached values as high as 840 ppm (**Retallack, 2009**) (**Figure 1a**). The limited existing geological proxies of terrestrial Antarctic temperature, from the Ross Sea (**Warny et al., 2009**) and off Adélie Land (**Sangiorgi et al., 2018**), indicate a surface air temperature warming of approximately 14°C to 25°C relative to today. Comparison with the global Mean Annual Temperature (MAT) clearly emphasises strong polar amplification occurred during the MCO (**Goldner et al., 2014**). Sub-Water Temperature (SWT) and SST reconstructions from the Ross Sea (ANDRILL-2A, **Levy et al., 2016**) and from the Adélie Land margins (**Sangiorgi et al., 2018, Hartman et al., 2018**) also support this polar amplification (**Figure 6c**). On the continental rise (paleo latitude 53°S, **Sangiorgi et al., 2018**), a 5-10°C SWT warming (likely summer) was recorded, but on the continental shelf (~77°S) this estimated warming was even larger, reaching 10-20°C through the MCO.

Together, far-field data and modelling experiments suggest a highly dynamic ice sheet during the MCO. The AIS was mostly responsive to eccentricity-modulated precession affecting local insolation and leading to widespread inland retreat of the land-terminating ice sheet on glacial-interglacial timescales (e.g. **Holbourn et al., 2013**). **Levy et al. (2019**) suggested glacial to interglacial ice volume fluctuations were of about 30 to 46 m SLE for a $\delta^{18}O$ shift of about 0.88‰, which was successfully simulated by **Gasson et al., (2016**) and **Colleoni et al.,**

(2018a) using idealised (but representative) mid-Miocene boundary conditions. Geological records recovered adjacent to the EAIS suggest it advanced and retreated many times through the TAM during the MCO (**Hauptvogel and Passchier, 2012**), but did not advance far beyond the coastline during glacial intervals (**Levy et al., 2016**). The cored interval spanning ~17 to 15 Ma at Integrated Ocean Discovery Program (IODP) Site U1521 **(McKay et al., 2019**) consists of diatom-rich mudstone and diatomite, which also indicates ice distal environments in the Ross Sea through the MCO.Sediments collected at IODP Site U1356, off the coast of the Adélie Land margin (East Antarctica), suggest open-water conditions at the site throughout the MCO (**Sangiorgi et al., 2018**). Modelling studies suggest the Wilkes Subglacial Basin remained free of grounded ice during warm interglacial episodes through the early to mid-Miocene **(Gasson et al., 2016; Colleoni et al., 2018a; Paxman et al., 2020**) but it is unclear whether grounded ice advanced across the region during glacial intervals prior to the MMCT (**Pierce et al., 2017**). Mg/Ca calibrated sea level and sequence boundary estimates suggest Global Mean Sea Level (GMSL) rise ranging from +20 meters to + 30 meters above present (**Miller et al., 2005; Kominz et al., 2008**) (**Figure 6e**). The compilation of simulated AIS contributions to GMSL vary between +15 m SLE to + 35 m SLE (**Langebroek et al 2009; Gasson et al. 2016; Colleoni et al., 2018b; Stap et al. 2019**) (**Figure 6e**). Such a large range mostly results from the use of different ice sheet models, different bed topographies and bathymetries, and different climate forcing in the mid-Miocene experiments. The range of potential Antarctic ice sheet GMSL contribution was significantly to +16 to +17 meters by **Gasson et al. (2016**), and **Colleoni et al. (2018b)** using an idealised Mid-Miocene paleotopography similar to that of **Paxman et al. (2019**), a prescribed atmospheric CO_2 of 500 ppm.

The mid-Pliocene Warm Period (mPWP, 3.3 - 3 Ma) is considered as one of the most geologically accessible and relevant examples of climate change driven by atmospheric CO_2 levels equivalent to present-day one (**Naish & Zwartz, 2012; Masson-Delmotte, 2013**; **Haywood et al., 2016**). Atmospheric CO_2 levels ranged between 300 ppm and 450 ppm and global mean temperature was about 2-3°C warmer than present during the warmest interglacials (**Masson-Delmotte, 2013**) (**Figure 6a & 6b**). One of the striking characteristics of the mPWP is that the SST proxy compilations reveal a meridional temperature gradient weaker than today (**Figure 4**), characterised by expansion of tropical to sub-tropical bands, no boreal and reduced austral summer sea-ice and thus a strong northern and southern polar amplification (e.g. **Lunt et al., 2012; Haywood et al., 2020**). Modelling showed that such SST patterns reflected a weaker Hadley circulation than today (**Brierley et al., 2009; Haywood et al., 2020**). In the Southern high-latitudes, the coastal Antarctic region was up to 6°C warmer than today (e.g. **McKay et al., 2012a;** compilation in **Dowsett et al., 2012**) (**Figure 6b**) mostly due to the fact that summer sea-ice and the ice sheet had retreated and the APF had contracted to more southern latitudes (**Taylor-Silva and Riesselman, 2018**). Evidence documents episodic sea ice in the Ross Sea, and offshore Adélie Land and in Prydz Bay (e.g. **McKay et al., 2012a;** compilation in **Dowsett et al., 2012**) (**Figure 6d**). Seasonal sea ice was likely present in the Weddell Sea (**Burckle et al. 1990**). Reconstructed SSTs show a pan-Antarctic warming of up to 5°C on the continental slope and rise (**Whitehead and Bohaty, 2003; Escutia et al., 2009**). SSTs also show a warming up to 6°C in the Ross Sea (**McKay et al., 2012**)(**Figure 6c**), which was likely caused by the sea-ice albedo feedback, and decreasing local albedo due to the retreat of coastal land ice.

Global mean sea level reconstructions (paleo-shorelines and sequence stratigraphy) indicate a sea level rise between about +15 m to +28 m (**Wardlay and Quinn, 1991; Dwyer and Chandler, 2009; Kulpecz et al 2009; Naish and Wilson, 2009; Sosdian and Rosenthal, 2009; Miller et al., 2012; Winnick and Caves, 2015 ; Dimitru et al 2019; Miller et al., 2020a**), whereas Mg/Ca paleothermometry calibration of benthic $\delta^{18}O$ records suggest a GMSL up to +40 m above present (**Figure 6e**). GMSL changes based solely on benthic $\delta^{18}O$ records, however, yield large uncertainties (±15m) (e.g. **Raymo, 2018**).GMSL change amplitudes larger than +30 m above present can only be explained by melting the terrestrial sectors of the AIS, but retreat of the the EAIS in the Ross Sea since 8 Ma appears not likely because a recent study that found extremely low concentrations of cosmogenic ^{10}Be and ^{26}Al isotopes in the ANDRILL AND-1B marine sediment core (**Shakun et al., 2018**). In addition, many of these peak GMSL estimates (e.g. **Miller et al., 2012**; **Hearty et al., 2020**) have not been corrected for regional deviations due to tectonics, glacio-isostatic adjustment, and dynamic topography (**Raymo et al., 2011**; **Rovere et al., 2015**; **Dumitru et al., 2019**). A reassesment **of Grant et al. (2019)** based on far-field data implies GMSL during the warmest mid-Pliocene interglacial was no higher than +21 m (**Grant & Naish, 2021**). This new estimate is very close to the average of +20 m above present provided by sea level reconstructions based on sequence stratigraphy and paleo-shore lines.

The compilation of simulated Antarctic ice sheet contributions to GMSL ranges from ~ +3 m SLE to +15 m SLE (**de Boer et al., 2017; Pollard and DeConto 2009; Pollard & DeConto, 2012; de Boer et al., 2015; Austerman et al., 2015; Gasson et al., 2015, Yan et al., 2016; DeConto & Pollard, 2016; Dolan et al., 2018**) (**Figure 6e**). Although there is no observational evidence of a potential melting from the Greenland Ice Sheet so far, recent transient numerical simulations suggest that the Greenland ice sheet melting could have contributed up to about 6 m SLE to GMSL rise (**De Boer et al., 2017).** Based on this estimate, the lower bound GMSL rise (+15 m) implies a contribution of the AIS no larger than 9 m SLE. Considering the upper bound of GMSL rise of about +20 m to + 28 m above present, the maximum contribution of the AIS thus ranges between +15 to +22 m SLE, implying melting of the WAIS and all marine-based sectors of the EAIS (e.g. **DeConto and Pollard, 2016; Golledge et al., 2017**). Site ANDRILL AND-1B in the Ross Sea recorded numerous occurrences of open-marine conditions suggesting frequent retreats of the Ross Ice Shelf during the mPWP (**Naish et al., 2009**). Provenance of fine-grained detritus offshore the Wilkes Subglacial Basin and ice-rafted debris offshore the Aurora Subglacial Basin and Prydz Bay was attributed to the retreat of marine-based sectors of the East Antarctic ice sheet (**Whitehead et al. 2006; Cook et al., 2013, 2014; Bertram et al., 2018; Blackburn et al., 2020**). Similar circum-Antarctic retreat of the marine-based sectors was simulated for one of the Early Pliocene interglacials (**Golledge et al., 2017**), supported by sedimentological and geological evidence of a circum-Antarctic warming events during that period (e.g. **Whitehead and Bohaty, 2003; Escutia et al 2009; McKay et al., 2012a**).

Marine Isotope Stage 31 (MIS 31, 1.081-1.062 Ma) is a prominent mid-Pleistocene interglacial categorised as “super interglacial” based on the expanded lacustrine sediment record from Lake El'gygytgyn in Siberia (**Melles et al., 2012**). It corresponded to an exceptionally high eccentricity and obliquity inducing particularly intense high-latitude summers. The level of atmospheric CO_2 is not well known for this interval but ranges from 300 to 420 ppm (**Honisch et al., 2009**) (**Figure 6a**). A circum-Antarctic warming has been inferred from sediment core analysis in the Ross Sea (**Scherer et al., 2008; Naish et al., 2009**), in Prydz Bay (**Villa et al., 2008**), on the Adélie Land margin and in the Antarctic Peninsula (**Beltran et al., 2020**). In particular, the presence of diatoms in the Cape Roberts sediment record (**Scherer et al. 2008)** suggested a 3-5 °C warming of upper ocean temperatures compared to today, with seasonally open-ocean (no summer sea ice) (**Figure 6d**). For the Adélie Land margin and the Antarctic Peninsula, **Beltran et al. (2020)** reconstructed summer SSTs that were on average were 3°C to 6°C warmer than today (**Figure 6c**). Similar warm conditions are recorded in the Ross Sea at site ANDRILL AND-1B as indicated by seasonal open-ocean conditions (**Figure 6d**) and suggesting a retreat of the Ross Ice Shelf (**Naish et al., 2009; McKay et al., 2012b**). Global mean sea level rise during MIS 31 is relatively poorly constrained from far-field records. Sea-level indicators preserved in coastal cliffs of the Northern Cape Province of South Africa and from Cape Range, Western Australia, suggest highstands not more higher than +15 – 16.5 m above present mean sea level (**Hearty et al., 2020; Sandstrom et al., 2020**). Note that those estimates are not corrected from GIA, dynamic topography and local tectonic. Far-field evidences of four consecutive Middle

Pleistocene Transition sea-level highstands between MIS 31 and MIS 35 were identified in a speleothem record from a western Sicily cave (Mediterranean Sea) (**Stocchi et al., 2017**). The peculiarity of this marine cave is that it has been last flooded between MIS 35 and MIS 31, and has been tectonically uplifted to higher elevations afterward. Among several GIA-modulated relative sea level scenarios, only those accounting for a significant AIS retreat up to about 25 m SLE at MIS 31 and 35, are capable to flood the marine Sicilian cave. The compilation of simulated AIS melting contribution to GMSL ranges from 2 to 10 m SLE (**DeConto et al., 2012; de Boer et al. 2013; Beltran et al 2020)** (**Figure 6e**). The upper bound of this range is in agreement with far-field, though uncorrected, sea level changes and indicate a large WAIS retreat, with a modest contribution from East Antarctic marine-based sectors. In fact, mineralogical provenance from IBRD from ODP Site U1090 (South Atlantic) and ODP Site U1165 (Prydz Bay) revealed that the EAIS retreated significantly over MIS 31 and particularly in the Prydz Bay region. However, other sectors of the EAIS were still characterized by active marine margins (**Teitler et al., 2015**). **Beltran et al. (2020)** suggested that the AIS retreat was caused by a stronger advection of Circumpolar Deep Water (CDW) resulting from the changes of the westerlies (subpolar jet). Such process was also inferred from changes in the geochemical composition of Holocene foraminifera shells from the Amundsen Sea and the aeolian dust from a West Antarctic ice core record. Both geological. evidence support the notion of enhanced advection of CDW onto the continental shelf due to a strengthening / poleward shift of the westerlies can drive WAIS retreat (**Hillenbrand et al., 2017**).

Marine Isotope Stage 11 (MIS 11, 425-375 ka) occurred close to the Mid-Bruhnes Event (**Figure 1**). It is the Late Pleistocene warm stage considered as one of the closest analogues to our future because astronomical forcing of a few time slices within MIS 11 are very similar to today (**Loutre and Berger, 2003**). MIS 11 is also the oldest middle Pleistocene interglacial categorised as a “super interglacial” based on lacustrine sediment records from the Lake El'gygytgyn in Siberia (**Melles et al, 2012**). Global mean air temperature was 1.5-3°C higher compared to modern temperatures (**Figure 6b**) although atmospheric CO_2 levels were around 280 ppm (**Figure 6a**). On the Antarctic plateau, the surface air temperature increased by 2°C to 3°C (**Jouzel et al., 2007; Uemura et al., 2018**). A polar amplification occurred during that period but was reduced compared to MIS 31 or older warm periods. MIS 11 is not really an intense but brief interglacial such as MIS 5e (130-116 ka, see below); its major characteristic is its longer duration of ~ 50,000 years (**Tzedakis et al., 2012**), which may have been key to ice sheet melting (**Irvali et al., 2020**). Reconstructed SSTs were not much warmer than modern temperatures (e.g. **Hodell et al., 2000; King and Howard, 2000; Becquey and Gersonde, 2002, 2003a, 2003b**). In fact, geological evidence supports the idea that a modest but sustain warming was at the origin of ice sheet retreat in the Wilkes Subglacial basin during MIS 11 (**Wilson et al., 2018; Blackburn et al., 2020**). Recent modelling studies, indeed, showed that the WAIS and part of the EAIS retreat could occur with a limited warming of +0.4°C if applied for a duration of 4 000 years (**Mas e Braga et al., 2021**). As with other past intervals, the absence of ice proximal oceanic temperature reconstructions is thus one of the critical gaps to constrain ice sheet simulations of this interval. Reconstructed GMSL from data suggest a rise of about +13 m above present sea level (**Raymo and Mitrovica, 2012; Roberts et al. 2012**) and up to +20 m during MIS 11 (**Kindler and Hearty, 2000; Hearty et al., 1999; Brigham-Grette, 1999**). Such a range of sea level rise implies the complete melting of both the Greenland Ice Sheet and the WAIS, which would account for about 12 m SLE, leaving about 8 m SLE from the EAIS melting (e.g. **Lythe et al., 2001; Warrick et al., 1996**. **Mas e Braga et al. (2021)** recently simulated a contribution from the WAIS around 4.3 - 4.5 m SLE and a contribution from the EAIS ranging from 2.3 to 3.7 m SLE. Sedimentological analyses from Erik Drift, Southeast Greenland reveal that most of South Greenland deglaciated during MIS 11 (**Reyes et al., 2014**). The compilation of simulated AIS contributions to GMSL ranges from ~ -3 m SLE to + 13 m SLE (**Tigchelaar et al. 2018; Sutter et al. 2019; Mas e Braga et**

al., 2021) (**Figure 6e**) and in absence of further geological constraints, it is difficult to refine this range.

Marine Isotope Stage 5e or Last Interglacial (LIG, 130-116 ka) was the most recent interglacial with temperatures warmer than today. It has long been considered as an analogue for the future climatic changes (**Jansen et al., 2007**). However, at the peak of the LIG, the astronomical forcing differed too much from the present-day to be a true analogue (**Ganopolsky and Robinson, 2011**). Nevertheless, the LIG presents a very useful time period for understanding the Earth System response (e.g. internal feedbacks in the climate system) to the Paris Agreement temperature targets (e.g. IPCC 1.5C Special Report). Atmospheric CO_2 concentration were low (**Figure 6a**), and reconstructed global mean temperature is estimated to have been about 0.5-2°C higher than today (**Masson-Delmotte et al., 2013; Hoffman et al., 2017**) (**Figure 6b**). The East Antarctic plateau recorded a warming up to 5.5°C at ~128.66 ka followed by a plateau around 2°C (**Petit et al., 1999; Watanabe et al, 2003; Jouzel et al., 2007**). A polar amplification thus occurred during this period (**Capron et al., 2017)**, and was broadly of the same magnitude than during MIS 11. Antarctic continental margin sediment records imply seasonal sea ice in most of the sensitive marine-based sectors (e.g. **Konfirst et al. 2012; Presti et al. 2011**) (**Figure 6d**). Global mean sea level rise is estimated to about +5.9 m to +9.3 m above present level from paleo-shorelines (**Dutton et al., 2015** and ref. therein) and up to almost +20 m based on calibration of benthic and planktonic $\delta^{18}O$ records (**Waelbroeck et al., 2009; Rohling et al., 2009**) (**Figure 6d**), also involving some Greenland Ice Sheet melting. However, ice core constraints and modelling studies suggest that the contribution from Greenland was likely about +2-3m (**Dahl-Jensen et al, 2013**), implying a significant meltwater contribution from Antarctica, although also a Greenland Ice Sheet contribution of up to +5.1 m SLE has been suggested (**Yau et al., 2016**). The compilation of simulated AIS melting contributions to GMSL range from about -2 m SLE to + 8 m SLE (**Figure 6e)** (e.g. **Huybrechts et al. 2002; Pollard and DeConto 2012; de Boer et al. 2015; Goelzer et al. 2016; Sutter et al. 2016; DeConto and Pollard 2016; Tigchelaar et al. 2018; Quiquet et al. 2018; Colleoni et al. 2018b; Sutter et al. 2019;** compilation in **De Boer et al., 2019)**. Antarctic ice core records of $\delta^{18}O$, considered as a proxy for ice volume changes, have been analysed in an attempt to better constrain the individual contribution of Antarctica to GMSL. Based on these analyses, numerical climate and ice sheet simulations suggest that part of the $\delta^{18}O$ signal could be explained by sea ice reduction rather than ice sheet retreat (e.g. **Holloway et al., 2016**). In absence of ice proximal ocean temperature reconstructions, as for other Late Pleistocene interglacials, it is very difficult to constrain the magnitude and timing of the Antarctic ice sheet retreat during this interval. The magnitude of oceanic warming required to trigger a large retreat of the marine-based sectors at that time varies between models from +2°C to +3°C relative to pre-industrial temperature (e.g. **Sutter et al., 2016, DeConto and Pollard, 2016, Turney et at., 2020**). However, **Turney et al. (2020)** also showed that with a modest ocean warming of +0.4°C, the major ice shelves disintegrated within 600 years. While continental margin sediments offshore from the Wilkes Subglacial Basin suggested a reduction of this marine-based sector of the EAIS during MIS 5e (**Wilson et al. 2018**), geological evidence from the WAIS are contradictory and suggests either that no major ice sheet retreat occurred (e.g. **Hillenbrand et al. 2002,2009; Spector et al., 2018; Clark et al., 2020**) or that considerable retreat took place (e.g. **Turney et al. 2020**).

6. Antarctica and global teleconnections: the bipolar seesaw

Inter-hemispheric heat transport, the so-called bipolar see-saw (**Stocker and Johnsen, 2003**), is another key process affecting AIS evolution. It regulates oceanic and atmospheric temperatures at sub-millennial to millennial time scales. The bipolar see-saw mechanism was hypothesized by **Stoker et al. (1998)** on the basis of observed asynchronous changes in ice core records between Greenland and Antarctica for some of the Dansgaard/Oeschger events that occurred during the last glacial cycle (**Blunier et al., 1998**). **Stocker and Johnsen, (2003)** hypothesized that icebergs melting and meltwater discharges close to the North Atlantic convection sites caused a substantial weakening of the Atlantic Meridional Oceanic Circulation (AMOC). Such a slowdown of the AMOC could have induced a gradual heat transfer to the South, with a lag of a few centuries to millenia, thus explaining the asynchronous temperature changes between Greenland and Antarctic ice core records during the last glacial period (**EPICA Community Members, 2006, Pedro et al., 2018**). Recent findings confirm that for example, the cooling of the Antarctic Cold Reversal is synchronous with the Bølling–Allerød warming in the Northern Hemisphere 14,600 years ago (**Stenni et al., 2011**). The Bølling–Allerød is coincident with the occurrence of meltwater pulse 1A (MWP-1A) that caused a rapid sea level rise of about 9 to 20 m (e.g. **Deschamps et al., 2012; Lambeck et al.,2014; Peltier et al.,2015; Liu et al., 2016**) at a rate of 4 meters/100 yr (**e.g. Peltier and Fairbanks, 2006; Deschamps et al., 2012; Carlson et al., 2012**). Although some studies have considered the AIS as a potential contributor to MWP-1A (e.g. **Clark et al., 1996; Bassett et al., 2007; Weaver et al., 2003; Golledge et al., 2014)**, most geological and glaciological studies argue against a large Antarctic contribution from either sector (e.g. **Licht et al. 2004; Bentley et al., 2010**, **The RAISED Consortium, 2014**; **Spector et al., 2017**). However, IBRD records from "Iceberg Alley" in the Scotia Sea showed recorded the occurrence of eight events between 20 ka and 9 ka, including the MWP-1A (e.g. **Weber et al., 2014**). **Etourneau et al., (2019**) showed that a +0.3–1.5 °C increase in subsurface ocean temperature (50–400 m) in the northeastern Antarctic Peninsula drove a major collapse and recession of the regional ice shelf during both the instrumental period and the last 9000 years. Modeling studies support the idea of a responsive marine-based sectors of the AIS at millennial time scales, driven by oceanic melting rather than by atmospheric forcing triggering fast ice sheet instabilities (e.g. **DeConto and Pollard (2009)**; **Golledge et al., 2014; Blasco et al., 2019; Lowry et al., 2019**).

Meltwater sources of such millennial oscillations are poorly constrained (e.g. **Clark et al., 2002, Peltier et al., 2005; Liu et al., 2016**). This limits our understanding of the causes of the events, i.e. warm water advection to the grounding line (e.g. **Golledge et al., 2014**) or bipolar seesaw caused by melting Northern Hemisphere ice sheets (e.g. **Menviel et al., 2011**), abrupt global mean sea level rise (e.g. **Clark et al., 2002; Golledge et al., 2014; Gomez et al., 2020**), atmospheric forcing (e.g. **WAIS Divide Project members, 2015**), or feedbacks within the climate system. Past, present and future meltwater-climate feedbacks have been widely studied and there is an extensive literature on modelling the global climate response to freshwater discharge from ice sheets (**e.g. Stammer et al., 2008; Roche et al., 2014; Boning et al., 2016; Bronselaer et al., 2018, Golledge et al., 2019; Sadai et al., 2020**). Results from these studies highlight the role of altered inter-hemispheric heat transport on the global climate both in the past and in the future. Different mechanisms respond to the freshwater at different timescales but the overall feedbacks loop spans the millennial scale (**Turney et al., 2020**). The sequence of those feedbacks loops is illustrated in **Figure 7** and is based on two recent contributions, i.e., **Turney et al. (2020)** for the LIG (130-116 ka), and **Golledge et al. (2019)** for projected climate changes until 2100 CE following RCP 8.5 emission scenario.

Turney et al (2020) reported evidence of substantial ice discharge across the Weddell Sea sector during the LIG based on a blue-ice core record. Substantiated by with climate and ice sheet simulations, they suggest that the ice discharge (and subsequent multi-meter global mean sea level rise) was caused by a millennial-scale oceanic warming following freshwater discharge in the Northern high latitudes (Heinrich event 11 at ~ 135-130 ka) and a weakening of the AMOC (**Böhm et al., 2015**). This mechanism corresponds to the bipolar see-saw. **Turney et al., (2020)** identified a loop of positive ice-sheet-climate feedbacks that further amplified the warming close to the Antarctic margin. **Grant et al. (2014)** identified two main meltwater pulses, one at 139 ka pre-dating Heinrich event 11 (135 ± 1 and 130 ± 2 ka) and one occurring at about 133 ka during this Heinrich event. **Marino et al. (2015)** found that Heinrich event 11 coincided with a rapid sea-level rise mostly explained by Northern hemisphere ice sheets deglaciation. The occurrence of this meltwater pulse supports the positive feedbacks described in **Turney et al. (2020)** and potentially explains the delayed timing of AIS contribution to the GMSL highstand at the LIG. A delay of a few thousands of years is supported by the idealised modelling study by **Blasco et al. (2019),** suggesting that the bipolar seesaw accumulated heat in the Southern Hemisphere, enhancing ocean warming on a millennial time scale during the last deglaciation. Similarly, **Clark et al. (2020)** suggested that the rate of global mean sea level changes during the LIG as well as spatial sea level variations could be explained by the responses of the Antarctic and Greenland ice sheets to Heinrich event 11 and associated climate feedbacks. The sequence of feedbacks in **Turney et al. (2020**) can be applied to other interglacials and is as follows **(Figure 7,** top**)**:

(1) Northern high-latitude freshwater was released during the Heinrich 11 event (~135 and 130 ka);
(2) Subsequently, a weakening of North Atlantic Deep Water (NADW) flow was observed, and heat was transferred gradually southward.
(3) An increased in meridional inter-hemispheric thermal gradient due to Northern high latitudes cooling induced a southward shift of the Inter-Tropical Convergence Zone (ITCZ) and of the Southern Hemisphere westerly winds (e.g. **Shevenell et al., 2011**).
(4) The southward shift and strengthening of the westerlies (e.g. **Hillenbrand et al., 2017, Etourneau et al., 2019; Lamy et al., 2019, Dickens et al., 2019**) drove an enhanced northward Ekman transport and a stronger southward advection of CDW on the continental shelf (e.g. **Hillenbrand et al., 2017, Minzoni et al., 2017**).
(5) Enhanced advection of CDW amplified the melting of the AIS and of the sea ice, triggering the AIS retreat. Northward transport of cool surface waters caused sea ice expansion and local atmospheric cooling.
(6) Large freshwater discharge caused a reduction in Antarctic Bottom Water (AABW) formation and a subsequent increase in NADW formation. Increase NADW formation led to heat transfer towards northern high latitudes, and thus a bipolar see-saw swing towards the north.

Golledge et al. (2019) simulated a similar ice-sheet-climate sequence of feedbacks by considering on-going and projected meltwater discharge from the Greenland and Antarctic ice sheets until 2100 CE at the same time. Results show that a slow-down of the AMOC occurs in response to Greenland Ice Sheet melting, and that projected meltwater discharge from Antarctica can trap heat of CDW at intermediate depths on the continental shelf (**Silvano et al., 2018, Bronselear et al., 2018; Sadai et al., 2020**), establishing a positive feedback establishes that further enhances AIS melting (**Figure 7**, bottom):

(1) Projected freshwater release at Northern and Southern high-latitudes;
(2) A weakening of NADW formation is observed, and heat is transferred gradually southward; In the South, freshwater stratifies the continental shelf waters.
(3) Increased in meridional inter-hemispheric thermal gradient induces a weak southward shift of the ITCZ and of the Southern Hemisphere Westerly winds.
(4) Southward shift of the westerly winds drives an enhanced northward Ekman transport compensated by a stronger southward advection of CDW on the continental shelves, which amplifies the melting of Antarctic ice shelves and sea ice.
(5) Continental shelf water stratification fosters a northward Ekman transport of cool surface waters associated with sea ice expansion and local atmospheric cooling. This mechanism further amplifies the advection of CDW to the AIS grounding line and initiates its retreat.
(6) Larger freshwater release further causes a reduction in AABW formation.

Compared to **Turney et al. (2020)**, the sequence of processes and feedbacks in 2100 remains incomplete and stops before all the heat from the North is transferred to South. This suggests that additional decades to centuries are needed for the effects of the bipolar see-saw on southern high latitudes to be felt.

7. The PAIS legacy: bridging the past and the future

7.1 The PAIS legacy

The PAIS legacy is clearly one of successful delivery on addressing high-level scientific priorities. Beyond this, it is also the story of a long-lasting network of collaborations, among different nations and researchers and striving to share scientific infrastructure and capability to investigate remote and challenging Antarctic regions and to address high-level scientific priorities. The multidisciplinary concept of the PAIS programme represented the key to its success. Eight years after the start of the programme, PAIS achievements are many, both in terms of field campaigns and in terms of scientific advances concerning Antarctic ice sheet dynamics. Several projects fostered by the PAIS programme fostered, which contributed to major scientific advances in constraining AIS contribution to past sea level changes, fostered by the PAIS programme, are briefly summarized below. This list is far from being exhaustive and the interested readers can refer to the other chapters of this book for more detailed descriptions of PAIS research outcomes and other time periods not discussed here.

Antarctic Ice Sheet sensitivity during past high-CO_2 worlds and its contribution to global sea-level change

Geological proxies from the Antarctic continental margin have improved reconstructions of ocean and land temperatures, sea-ice extent, ice sheet extent, subglacial hydrology, carbon cycle feedbacks and paleogeography for past warm climate states. This has provided improved boundary conditions for testing and developing ice sheet and climate models skills and performance, as well as evaluating model sensitivity. Significant outcomes include:

- Reconciling southern high-latitude meridional temperature gradients and polar amplification between model simulations and data during Greenhouse climates (e.g **Pross et al., 2012; DeConto et al., 2012**) and new knowledge of Antarctic margins SSTs and SWTs during the MCO, the mPWP and MIS 31 (**McKay et al., 2012**, **Levy et al., 2016**, **Sangiorgi et al., 2018; Hartman et al., 2018; Beltran et al, 2020**). Polar amplification is much larger during the MCO, mPWP and MIS 31 than during the Pleistocene interglacials. Those findings allow an estimate of Earth's climate sensitivity to high atmospheric CO_2 concentrations.
- Constraining equilibrium and transient ice volumes (e.g. **de Boer et al., 2015; DeConto et al., 2012; Pollard et al., 2015; DeConto & Pollard, 2016; Goelzer et al., 2016; Gasson et al., 2016; Golledge et al., 2017; Dolan et al., 2018; Clark et al., 2019, Stap et al., 2019**), and the contribution to global sea-level under past "warmer-than-present" climates (e.g. **Miller et al., 2012; Dutton et al., 2015; Miller et al., 2020a, 2020b**).
- Recognition of the importance of bedrock topography and paleobathymetry on past Antarctic ice volume reconstructions (e.g. **Gasson et al., 2016** building on **Wilson & Luyendyke, 2009; Hochmuth and Gohl, 2019; Paxman et al., 2019; Hochmuth et al., 2020, Paxman et al., 2020**) and sensitivity to ocean warming (e.g. **Colleoni et al., 2018a; Paxman et al., 2020**).
- Recognition of the sensitivity of marine-based sectors of the EAIS from models and data (e.g**. Cook et al., 2013, 2014; Reinardy et al., 2015; Bertram et al., 2018; Pierce et al., 2017; Scherer et al., 2016; Levy et al., 2016; Gasson et al., 2016; Aitken et al., 2016; Gulick et al., 2017; Simkins et al., 2017; Golledge et al., 2017b; Wilson et al., 2018; Blackburn et al., 2020***)*.

- Insights into the influence of mean climate state (CO_2) on the response of the AIS to orbital forcing (e.g. **Dolan et al., 2011; Patterson et al., 2014; Levy et al., 2019** building on concepts in **Naish et al., 2009**; **Stap et al., 2019, 2020, Sutter et al., 2019**).

Geological evidence of ocean forcing and marine ice sheet instability:

The potential for abrupt and non-linear “runaway” retreat of the marine-based sectors of the AIS due to marine ice sheet instability (MISI) and potentially also marine ice cliff instability (MICI) up until recently had only been mathematically simulated in ice sheet models.

- Geological observations of the last deglaciation and recent observations coupled with models have now identified MISI during the Holocene after atmospheric forcing had weakened in the Ross Sea (e.g. **Jones et al., 2015; McKay et al., 2016; Spector et al., 2017; Bart and Tulaczyk, 2020**), and potentially MICI in the Amundsen Sea sector (**Wise et al., 2017**) and Antarctic Peninsula (**Rebesco et al., 2014**)
- There are geological and modern oceanographic observations of oceanic warm waters reaching the grounding line of marine-based ice sheets (e.g. **Joughin et al., 2011; Schmidko et al., 2014; Hillenbrand et al., 2017**; **Rintoul et al., 2016**; **Smith et al., 2017; Hansen and Passhier, 2017**)

Improved temporal and spatial patterns of AIS retreat and its contribution to global Melt-Water Pulse 1A:

Improved geological and bathymetric constraints combined with ice sheet models have shown:
- An improved understanding of the extent and dynamics of the Last Glacial Maximum ice sheet and deglaciation into the Holocene (e.g. **Hillenbrand et al., 2014; Larter et al., 2014; O’Cofaigh et al., 2014; Hodgson et al., 2014; Golledge et al., 2013; Mackintosh et al., 2014; The RAISED Consortium, 2014; Anderson et al., 2014; Johnson et al., 2014; Lee et al 2017; McKay et al., 2016**).
- the AIS contributed to melt-water pulse 1A (**e.g. Golledge et al., 2014; Weber et al., 2014**), though not from all sectors (e**.g. Spector et al., 2017**) and to other millennial scale fluctuations with estimated contributions to global mean sea level up to 6 m SLE (**Blasco et al., 2019**, **Golledge et al., 2014**), at a rate of about 1 meter/century in the case of MWP-1A (e.g. **Gollege et al., 2014**).

A better understanding of ice-sheet-ocean interactions

- During the last deglaciation, proxy data and model simulations consistently find that ocean warming drove the ice sheet retreat in different sectors of Antarctica (e.g. **Hillenbrand et al., 2017; Crosta et al., 2018**, **Wilson et al., 2018**).
- Ocean warming is also thought to have accelerated the last deglaciation in the Ross Sea during MWP-1A (e.g. **Golledge et al., 2014**), although this finding is challenged by geological evidence from the Tran-Antarctic Mountains (e.g. **Spector et al., 2017**). Based on regional ice sheet simulations, atmospheric forcing can enhance, diminish or compensate for oceanic warming during the first half of the deglaciation, while during the second half, ocean warming clearly drove the end of the ice sheet retreat (**Buizert et al., 2015; Blasco et al., 2109; Lowry et al. 2019)**.
- Strengthening of the subpolar jet during deglaciation (e.g. **Lamy et al., 2020**) enhanced advection of CDW towards the continental margins. (**Hillenbrand et al., 2017; Minzoni et al., 2017; Salabarnada et al., 2018; Evangelinos et al., 2020**).
- Improved understanding of sedimentological facies indicative of sub-ice shelf environment opens new perspectives on quantifying the influence of the ocean on the AIS evolution through time (e.g. **Yokoyama et al., 2016**; **Smith et al., 2019**).
- Freshwater release from the Northern high latitudes can induce a bipolar seesaw, transferring heat to the Southern Hemisphere and fostering AIS retreat a few

thousands of years later (**Buizert et al., 2015; Blasco et al., 2019; Turney et al., 2020; Clark et al., 2020**). Likewise, freshwater release from Antarctica can stratify the ocean, reduces vertical mixing and the release of heat and gas to the surface, increase heat transport at the grounding lines of marine-based ice sheets (**e.g. Golledge et al., 2019; Silvano et al., 2019**)

Antarctic ice-Earth interactions and their influence on regional sea-level variability and Antarctic Ice Sheet dynamics

The importance of departures in regional sea-level changes from eustatic sea-level due to rotational, visco-elastic and gravitational changes as water mass is transferred between the ice sheets and the ocean has been identified in the far and near-fields of the AIS from paleo-reconstructions (e.g. **Clark et al., 2002; Milne et al. 2008, Raymo et al. 2011, Raymo and Mitrovica, 2012; Stocchi et al., 2013; Rovere et al., 2014**). This has been established through 1D and 3D glacio-isostatic adjustment models that couple (runtime or asynchronously) ice sheets and solid Earth processes constrained by both near-field and far-field geological reconstructions of sea-level changes. Important outcomes include:

- Role of Earth deformation processes (GIA and dynamic topography) on near-field sea-level changes and ice sheet dynamics (e.g. **Gomez et al., 2018; Stocchi et al., 2013; Austermann et al., 2015; Gomez et al., 2015; Whitehouse et al., 2017, 2019; Gomez et al., 2018; Pollard et al., 2017; Kingslake et al., 2018**).
- Impact of global gravitationally consistent sea level changes induced by Northern Hemisphere ice sheets fluctuations on the retreat of the AIS (e.g. Gomez et al., 2020).
- Impact of long-term global mean sea level changes on the stability of EAIS (e.g. **Shakun et al., 2018; Jakob et al., 2020**).

Improved interpretation of subglacial processes from mapping seabed

- Multibeam campaigns in different sectors of Antarctica have mapped the geomorphological footprints of paleo ice streams and their associated paleo-drainage networks (e.g. **Nitsche et al., 2013, The RAISED consortium, 2014; Simkins et al., 2017, Larter et al., 2019, Kirkham et al., 2019**) as well as other subglacial features (**Kuhn et al., 2017**; **Bart et al., 2018; Stokes et al., 2018; Dowdeswell et al. 2020**).
- Analysis of the characteristics of those geomorphological features inform the long-term mean and potential maximum rates of grounding line retreat (e.g. 1 to 10 km/yr, **Bart et al., 2018, Dowdeswell et al. 2020**), but also show that meltwater can enhance ice flow and cause ice surges and meltwater outbursts (e.g. **Simkins et al., 2017; Kuhn et al., 2017**). These reconstructions provide constraints on the ice flow regime during both advances and retreats and on the mechanics and dynamics of ice stream (**Stokes et al., 2018**).

Paleo-data calibrated ice sheets models provide revised global sea-level predictions for IPCC scenarios

A new generation of continental scale ice sheet models that simulate MISI and in one case MICI have been developed and tested by reconstructing past AIS volume and extent constrained by paleoclimate and paleo-ice extent data. These models have been used to simulate future Antarctic meltwater contribution to global mean sea-level changes based on the Representative Concentration Pathways. Implications include:

- That Antarctic contribution to global sea-level rise for the year 2100 CE and beyond may have been under-estimated in IPCC AR5 projections especially for high emission scenarios (e.g. **Golledge et al., 2015; DeConto & Pollard, 2016; Edwards et al., 2019; Golledge et al., 2020**).

- These paleo-calibrated AIS models show that a threshold for marine ice sheet stability may exist at ~1.5-2°C global warming above pre-industrial (e.g. around RCP 2.6, the target of the Paris Agreement) (e.g. **Golledge et al., 2015; Clarke et al., 2015; Pollard & DeConto, 2016**).
- Recent paleo-studies have stressed that a moderate local oceanic warming, lower than the upper bound of +1.5°C-2°C for pan-Antarctic ocean warming can also trigger fast ice sheet retreat if applied for a few centuries: **Beltran et al. (2020)**; **Turney et al., (2020); Golledge et al., (2017a); Bakker et al., (2017); Feldmann and Levermann, (2015)**. This highlights the importance of polar amplification for the fast response of polar areas under past and future global warming conditions.

7.2 Challenges for the next programmes

Gaps illustrated above highlight the necessity to assess whether or not the WAIS only partially retreated or totally disintegrated during past warm periods. Records of such massive ice sheet retreats are possibly located below the ice sheet. Locating subglacial drilling sites that could have recorded such extensive retreat represents a high priority challenge worthy of future field campaigns (e.g. **Bradley et al., 2012; Spector et al., 2018**). Similarly, it is urgent to assess the EAIS marine-based sectors sensitivity to oceanic and atmospheric warming during past warm periods (e.g**. Cook et al., 2013, 2014; Reinardy et al, 2015; Aitken et al., 2016; Gulick et al., 2017; Pierce et al., 2017; Wilson et al., 2018; Blackburn et al., 2020)** and their potential contribution to global mean sea level change (e.g. **DeConto and Pollard 2016; Paxman et al., 2020; Mas e Braga et al., 2021**) to refine their future contribution to on-ongoing sea level rise (e.g. **Golledge et al., 2017b; Rignot et al., 2019**).

Paradoxically, even though the Pleistocene interglacials are more recent and well documented in many places around the world, AIS fluctuations through time have destroyed most of the ice proximal geological evidence of these interglacials on the continental shelf, making direct records of the ice sheet's behaviour difficult to find. Only a few precious SST records are currently available from the Antarctic continental slope and rise and those records are indirect and cannot fill the gap of ice proximal ocean temperature records. This data gap directly impedes the validation of numerical paleo-climate and paleo-ice sheet numerical simulations. The interpretation of sedimentary facies and geomorphological features on the seafloor, however, does allows to infer the type of sub-glacial environments and thus the ice flow during past deglaciations to be reconstructed (e.g. **Smith et al., 2019; Simkins et al., 2017; Bart et al., 2018**; **Prothro et al., 2020**).

Another observational challenge is to recover records with sub-millennial temporal resolution for the different past warm periods. Such high-resolution archives can be recovered from the continental slope and rise, by drilling levee deposits and contourite systems, or from the on the continental shelf in overdeepened basins and fjords (e.g. ODP leg 178 The Palmer Deep **Domack et al., 2001**; IODP Exp 318**, Ashley et al., 2020**; IODP Expedition 374 Ross Sea, **McKay et al., 2019;** IODP Expedition 379 Amundsen Sea, **Gohl et al., 2019;** approved IODP proposal 732 Antarctic Peninsula, **Channell et al.**). High-resolution data represent the bridge between the past and the future, in particular for centennial to millennial-scale climate oscillations (e.g. **Weber et al. 2014; Bakker et al., 2017; Bracegirdle et al., 2019; Noble et al., 2020; Golledge et al., 2020**). High-resolution sedimentary data are also important for correlating marine sediment records with ice core records of the past 800,000 years. The on-going Beyond EPICA: Oldest Ice project (e.g. **Fischer et al. 2013; Parrenin et al., 2017;** https://www.beyondepica.eu/) will allow correlation with expanded sediment records from the Ross Sea (IODP Exp. 374) (**McKay et al., 2019**) including the MIS-31 super-interglacial event and the Amundsen Sea (IODP Exp. 379) (**Gohl et al., 2019**) across the Mid-Pleistocene Transition and from future expeditions, for example the IODP proposal 732-Full2.

To maintain pace with advances of the observational ice sheet community, the paleoclimate modelling community will need to expand efforts more in regional atmospheric and oceanic modelling for different past periods representing both glacial and interglacial contexts. Regional modelling is computationally expensive and also requires highly resolved boundary conditions at high frequency to capture the local variability of processes. Improved large-scale global climate simulations will also be required to support regional modelling. Many on-going data-model comparison initiatives already exist, and some of them focus on the periods described in this chapter, as for example the Paleoclimate Model Intercomparison Project (PMIP, now in phase 4) (**Kageyama et al., 2018**), the Pliocene Model Intercomparison Project (PLIOMIP, now in phase 2) (**Haywood et al., 2020**), the recently started Miocene Model Intercomparison Project (MIOMIP) (**Steinthorsdottir et al., 2020** and related special issue) and the Deep-Time Model Intercomparison Project (DEEPMIP) (**Lunt et al. 2017**). PMIP focuses on the Late Pleistocene and now also includes transient simulations of entire interglacials using coupled atmosphere-ocean models. DEEPMIP focuses mainly on the EOT and the Eocene warmth. More refined global mean sea level records are also necessary to better assess Antarctic ice volume fluctuations over the past 34 Myrs. Both MCO and mPWP periods are of high interest to assess Earth climate sensitivity to high CO_2 concentrations (similar to the projected ones) and global mean sea level rise (e.g. **Haywood et al., 2016**; **Steinthorsdottir et al., 2020**). Sequence stratigraphy of the continental margins is a powerful approach and the key to fill this gap. However, improvements are needed, especially to correct those records from glacio-hydro-isostasy (e.g. **Grant et al., 2019,2021**). Thus, coupled ice-sheet-GIA-sediment erosion and transport models are needed (e.g. **Pollard & DeConto, 2003, 2019, Whitehouse et al., 2019**).

Finally, while the climate and paleoclimate communities are currently putting efforts in the development of fully coupled Earth System Models, such models are too computationally demanding to allow for long-term transient simulations. With upcoming progress in scientific computing, and progress in the computing facilities themselves, using fully coupled Earth System Models now seems an achievable objective for paleo studies.

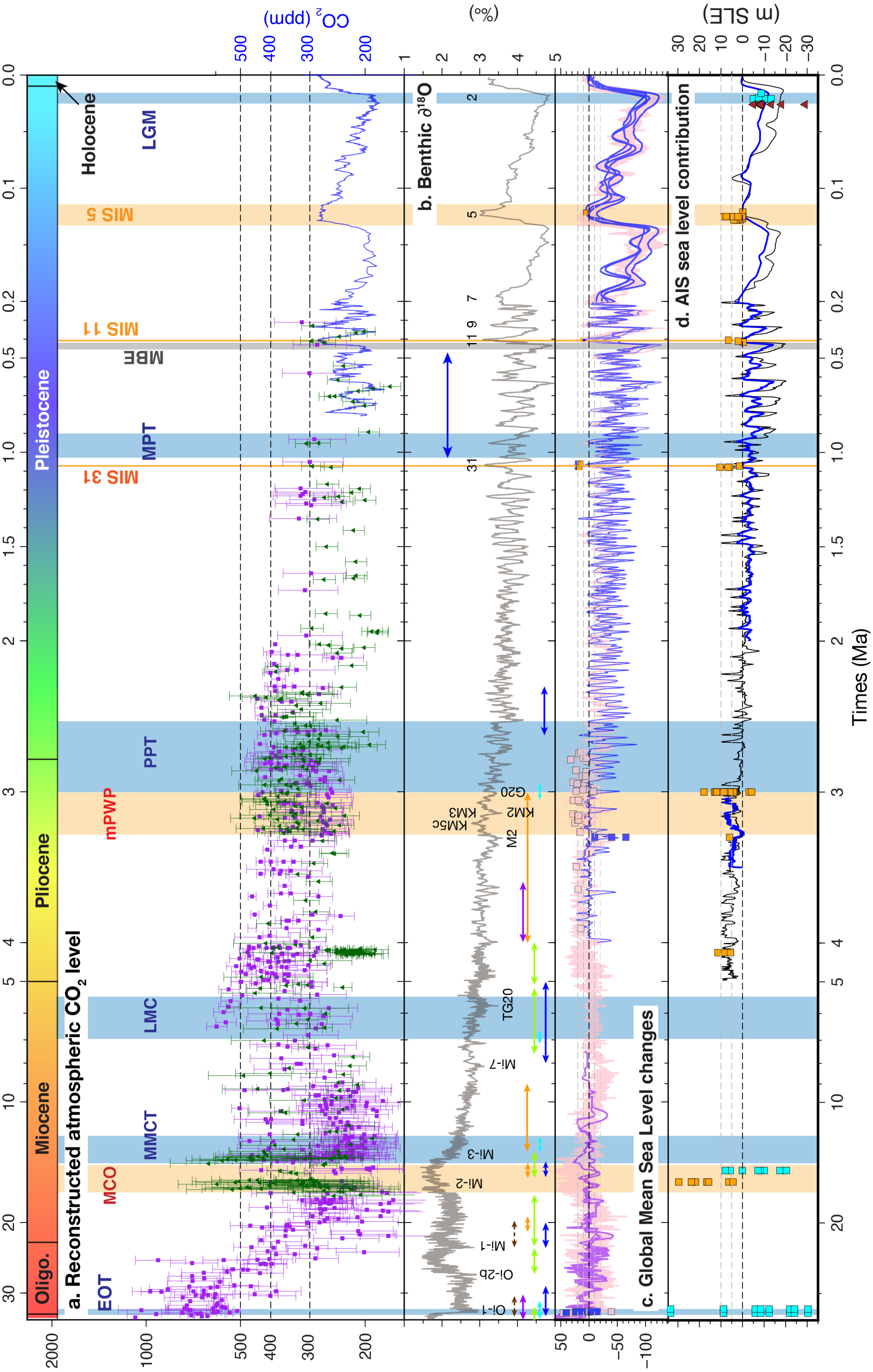
7.3 Long-term projections and the role of PAIS and future programs

Future projections of AIS evolution have shown large improvements over the past few years (**e.g. DeConto and Pollard, 2016; Pattyn et al., 2018; Edwards et al., 2019**). However, related uncertainties remain large, indicating that fundamental knowledge gaps still persist about ice sheet dynamics, and interactions with the atmosphere, ocean and the solid earth (**Whitehouse et al., 2019**). **Morlighem et al. (2020) released an updated** subglacial topography map revealing the high-resolution bed morphology of some of the glacial troughs and their potential in causing AIS instability in case of fast retreat of the grounding line. Many of them are still unexplored, despite their clear importance in reconstructing the AIS past, present and future dynamics. The release of the IPCC Special Report “The Ocean and the Cryosphere in a Changing Climate” (SROCC) in September 2019 (**IPCC, 2019**) showed that our understanding of the various contributions to GMSL change has improved since the last IPCC Assessment Report 5 (AR5) in 2013. After the release of AR5, further satellites observations revealed that Antarctic ice shelves were thinning faster than previously thought (**Paolo et al., 2015**), caused by observed warming in the surrounding ocean (**Pritchard et al., 2012**). Recent re-assessments of 20th century observations confirmed the AIS has been losing mass since the publication the publication of IPCC AR5 and that this mass loss strongly accelerated at the end of the 20th century (e.g. **Shepherd et al., 2018, Rignot et al., 2019**).

To precisely assess the AIS contributions to GMSL changes, the polar community has increased the monitoring and modelling of AIS evolution. Attention has been focused on ice shelf buttressing and on large partly marine-based drainage basins of the West and East Antarctic ice sheet (**Fürst et al., 2016**) (e.g. Pine Island Glacier, Thwaites glacier, Totten glacier, Recovery ice stream, Foundation ice stream) and ice-ocean interactions around Antarctica. The particularity of most of the marine-based sectors of the Antarctic ice sheet is that they are grounded on a bed with retrograde slope (**Joughin and Alley, 2011; Morlinghem et al., 2020**) or that their buttressing ice shelves are pinned on a sill with retrograde slope bed and are thus vulnerable to future MISI. New estimates of future GMSL rise from the **IPCC SROCC (2019)** amount to 0.43 m (0.29–0.59 likely range, RCP 2.6 scenario) and 0.84 m (0.61–1.10, likely range, RCP 8.5) in 2100 CE , with the possibility of multi-meter sea level rise by 2300 CE (**Golledge et al., 2015; Clark et al., 2016**) but with “deep uncertainty (**IPCC SROCC, 2019)**. Ice shelf loss is a key prerequisite for the onset of marine ice shelf instabilities. The large uncertainties in the most recent estimates of sea level rise from Antarctica mostly result from our inability to assess the potentially unstable behaviour of the marine-based sectors of the AIS, and in particular: the sensitivity of ice shelves to sub-ice shelf melting from below and surface warming above. These gaps inevitably lead to model-dependent results, particularly for processes that are parameterized (**Asay-Davis and Jourdain, 2017**). This is where the past can close those knowledge gaps, and provide necessary observational constraints to model the past and the future evolution of the Antarctic ice sheet (e.g. **Gasson and Keisling, 2020**).

The Earth’s past provides a natural laboratory for testing realistic cases of ice-sheet-climate-solid earth interactions at different timescales (**Bracegirdle et al., 2019**). The research produced within the PAIS programme has shown that the AIS potentially crossed its tipping point for major ice loss many times since the onset of large-scale glaciation (~34 Ma) under climatic conditions warmer than today. Past periods have the potential to identify thresholds for instability (e.g. **Naish et al. 2009; Cook et al., 2013; Weber et al., 2014; Wilson et al. 2018**) or large retreat/re-advance events (e.g. **Scherer et al., 2016; Golledge et al., 2017; Kingslake et al., 2018; Wilson et al. 2018**) and thus to provide credibility to future scenarios. Paleo-records also have the potential to reveal new mechanisms as for example the marine-ice cliff instability (MICI, **Pollard et al., 2015; ; DeConto & Pollard 2016).** MICI involves the fast disintegration of ice shelves by surface-melt induced hydro-fracturing that can trigger MISI, and rapid calving at thick, marine-terminating ice margins. This mechanism has been implied to explain rapid major mass loss from the WAIS and EAIS during MIS 5e and the mPWP (**Pollard et al. 2015; DeConto & Pollard 2016**) but many open questions about MICI and its possible role in past and future sea level rise remain (e.g., **Edwards et al. 2019; Pattyn et al., 2018**). Geological and glaciological evidence can also highlight feedbacks in the ice sheet-climate system **(Turney et al., 2020**) or processes that might not appear policy-relevant, but are indeed determinant in understanding the future sensitivity of the AIS and sea level rise to ongoing and projected climate changes (**Haywood et al., 2019**).

Figure1



a. Reconstructed atmospheric CO_2 level
b. Benthic $\delta^{18}O$
c. Global Mean Sea Level changes
d. AIS sea level contribution
Holocene
Pleistocene
Pliocene
Miocene
Oligo.
LGM
MIS 5
MIS 11
MBE
MPT
MIS 31
PPT
mPWP
LMC
MMCT
MCO
EOT
pCO2 (ppm)
CO2 (ppm)
(‰)
(m)
(m SLE)
Times (Ma)
G20
KM2
KM3
KM5c
M2
TG20
Mi-7
Mi-3
Mi-2
Mi-1
Oi-2b
Oi-1

Figure 1: Proxies and simulations synthesis over the past 34 million years. Note that the time scale is logarithmic **a.** Reconstructed atmospheric p[CO_2] levels are based on alkenone (**Pagani et al. 2005, 2010, 2011; Seki et al., 2010; Badger et al. 2013a, 2013b; Zhang et al., 2013; Super et al. 2018**) and on boron (**Honisch et al., 2009; Bartoli et al., 2011; Foster et al., 2012; Greenop et al., 2014; Martinez-Boti, 2015**). Late Pleistocene atmospheric CO_2 levels are based on the Antarctic ice core composite record from **Bereiter et al., (2015)**; **b.** Deep sea benthic $\delta^{18}O$ record from **Zachos et al. (2001a, 2008)**. Marine isotope stages (glacials and interglacials) discussed or named in the chapter are indicated and based on **Miller et al., (1991)** for the Oligocene and Miocene**,** on **Haywood et al. (2016)** for the Pliocene and on **Lisiecki and Raymo (2005)** for the Pleistocene. Note that marine isotopes stages EOT-1 (~34.46-33.9 Ma) and EOT-2 (33.7 Ma) are not indicated. Seismic stratigraphic unconformities from different Antarctic sectors are reported with arrows, based on **Hochmuth et al., 2020** and references therein (Ross Sea: dark blue; Wilkes Land: green; Weddell Sea: cyan; Amundsen Sea: orange; Cosmonaut Sea: brown; Prydz Bay: purple); **c.** Reconstructed (pink, purple) and simulated (blue) global mean sea level changes (GMSL). Proxy-based reconstructions: benthic $\delta^{18}O$ (**Miller et al., 2020a**) from EOT to present, backstripped sequence stratigraphy from New Jersey from EOT until the Late Miocene (**Miller et al., 2005**, **Kominz et al., 2008**). For the EOT (pink solid squares): **Pekar et al., (2002); Pekar and Christie-Blick, (2008); Lear et al. (2008), Katz et al., (2008), Miller et al., (2009); Bohaty et al. 2(012); Houben et al., (2012); Stocchi et al. (2013)**. For the Pliocene: converted benthic $\delta^{18}O$ record from **Dimitru et al., (2019)** until the PPT. Pink squares correspond to reconstructed Pliocene highstands (**Wardlay and Quinn, 1991; Dwyer and Chandler, 2009; Kulpecz et al 2009; Naish and Wilson, 2009; Sosdian and Rosenthal, 2009; Miller et al., 2012; Winnick and Caves, 2015; Dimitru et al 2019**) and to M2 glaciation (**Miller et al., 2005; Naish and Wilson, 2009; Dwyer and Chandler, 2009**). For Pleistocene: sea level reconstructions are taken from **Hearty et al., (2020); Sandstrom et al., (2020) f**or MIS 31 (uncorrected from GIA and dynamic topography) and from **Raymo and Mitrovica (2012)** and **Roberts et al. (2012)** for MIS 11. For the last 400 kyrs, reconstructed curves of sea level changes are from **Waelbroeck et al. (2002)** and from **Rohling et al. (2009).** For MIS5, paleoshorelines data are from the compilation in **Dutton et al. (2015)** and references therein. Simulated GMSL changes are from: **Bintanja et al., (2005**), **de Boer et al., (2015)** and (**Stap et al., 2017**) and from **Raymo et al (2006)** for MIS 31 (blue squares); **d.** Simulated Antarctic ice sheet melting contributions (meter Sea Level Equivalent, m SLE) to GMSL changes are from ice sheet simulations (squares and curves) and from Glacio-isostatic-Adjustment simulations (dark red triangles). Note that some of the reported simulated ice volumes do not refer to volumes above floatation. For the EOT: **DeConto and Pollard (2003), Pollard and DeConto (2005), Gasson et al (2014), Ladant et al (2014), Liakka et al (2014), Wilson et al (2013)**. For the Miocene: **Langebroek et al (2009), Gasson et al. (2016), Colleoni et al. (2018b), Stap et al. (2019)**. For Early Pliocene: **Pollard and DeConto (2009**, transient and black line), **Golledge et al. (2017**, orange squares). For mPWP to Late Pliocene: **de Boer et al. (2017**, transient blue line), **Pollard and DeConto (2009**, transient black line). Orange squares come from **Tan et al. (2017)** for M2 glaciation and remaining symbols are for the mPWP considering **Pollard and DeConto, (2012), de Boer et al., (2015), Austerman et al. (2015), Gasson et al. (2015), Yan et al., (2016), De Conto and Pollard (2016), Dolan et al. (2018).** When simulations were run with averaged mPWP climatic conditions, orange squares are indicatively plotted at 3 Ma. For the entire Pleistocene: blue line - **de Boer et al (2014)** and black line - **Pollard and De Conto (2009)**. For MIS 31: **De Conto et al. (2012)**, **de Boer et al. (2013), Beltran et al (2020)**. For MIS 11: **Tigchelaar et al. (2018), Sutter et al. 2019, Mas e Braga et al. (2020)**. For MIS 5: **Huybrechts et al. (2002), Pollard and DeConto (2012), de Boer et al. (2015), Goelzer et al. (2016), Sutter et al. (2016), DeConto and Pollard (2016), Tigchelaar et al. (2018), Quiquet et al. (2018), Colleoni et al. (2018b), Sutter et al. (2019).** For LGM based on ice sheet simulations: **Philippon et al. (2006), Mackintosh et al (2011), Golledge et al. (2012), Brigg et al (2013), de Boer et al (2013), Golledge et al. (2014), Pollard et al. (2016), Quiquet et al. (2018), Colleoni et al (2018b), Sutter et al (2019)**. For LGM based on glacio-isostatic adjustment simulations: **Peltier et al. (2004), Ivins and James (2005), Lambeck et al. (2014), Whitehouse et al. (2012), Ivins et al. (2013), Gomez et al. (2013), Argus et al. (2014)**. Cold periods of interest in this chapter are indicated with blue bars: EOT - Eocene-Oligocene Transition; MMCT - Mid-Miocene Climatic Transition; LMC - Late Miocene Cooling; PPT - Plio-Pleistocene Transition; MPT - Mid-Pleistocene Transition; MBE - mid-Brunhes Event; LGM - Last Glacial Maximum. Warm periods of mentioned in this chapter are indicated with orange bars: MCO - Mid-Miocene Climatic Optimum; mPWP - mid-Pliocene Warm Period, MIS 31, MIS 11, MIS 5.

Figure 2

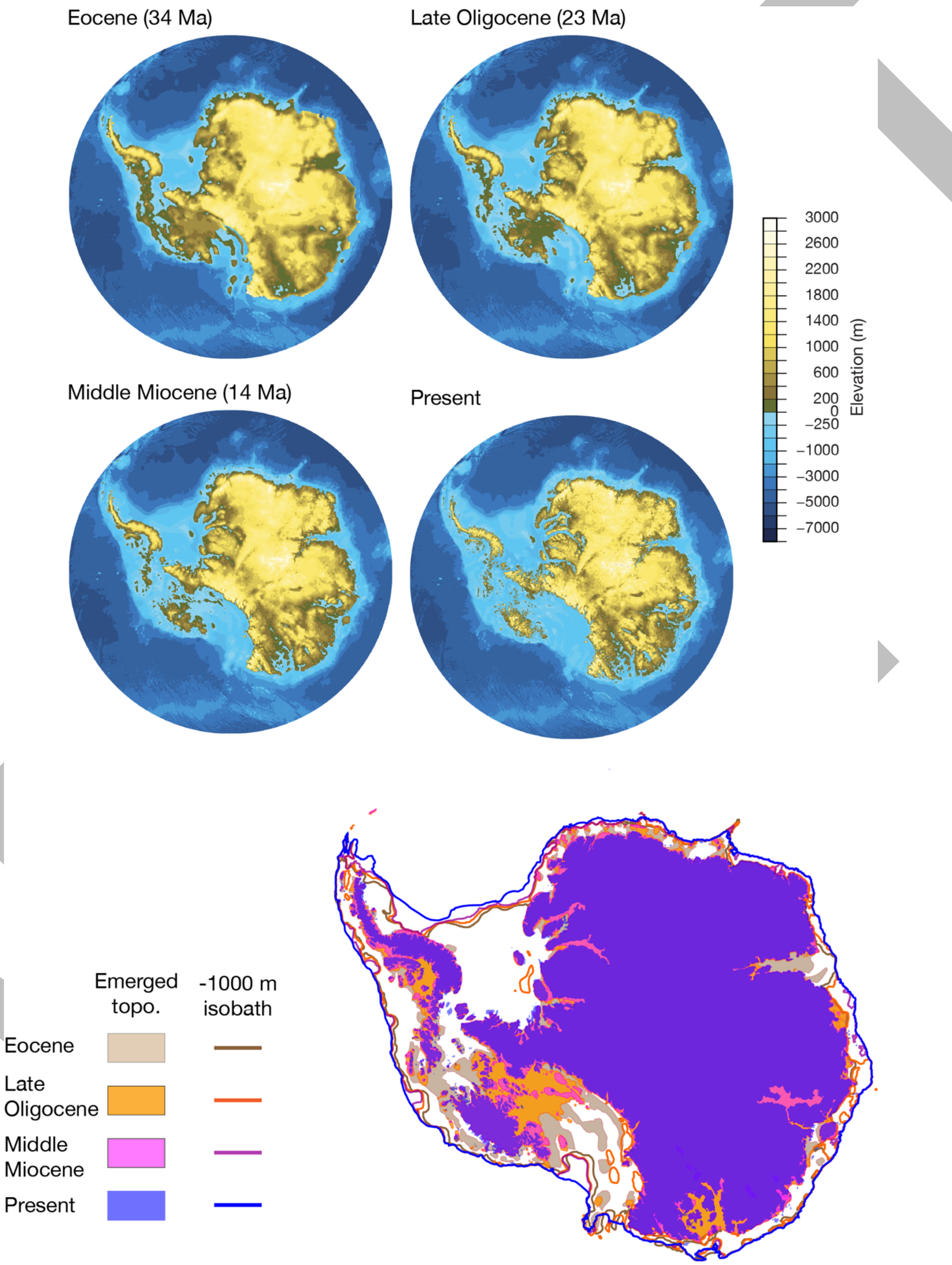


Figure 2 Top: Pan-Antarctic isostatically-relaxed paleogeographic reconstructions from **Paxman et al. (2019)** for the Eocene (34 Ma), the Late Oligocene (23 Ma), the Middle Miocene (14 Ma) and BEDMAP2 for modern pan-Antarctic geography (**Fretwell et al., 2013**). Bottom: superimposed Eocene (brown), Late Oligocene (orange), Middle Miocene (pink) and modern (blue) emerged topography. Isobath at -1000 meters for each time slices is also indicated.

Figure 3

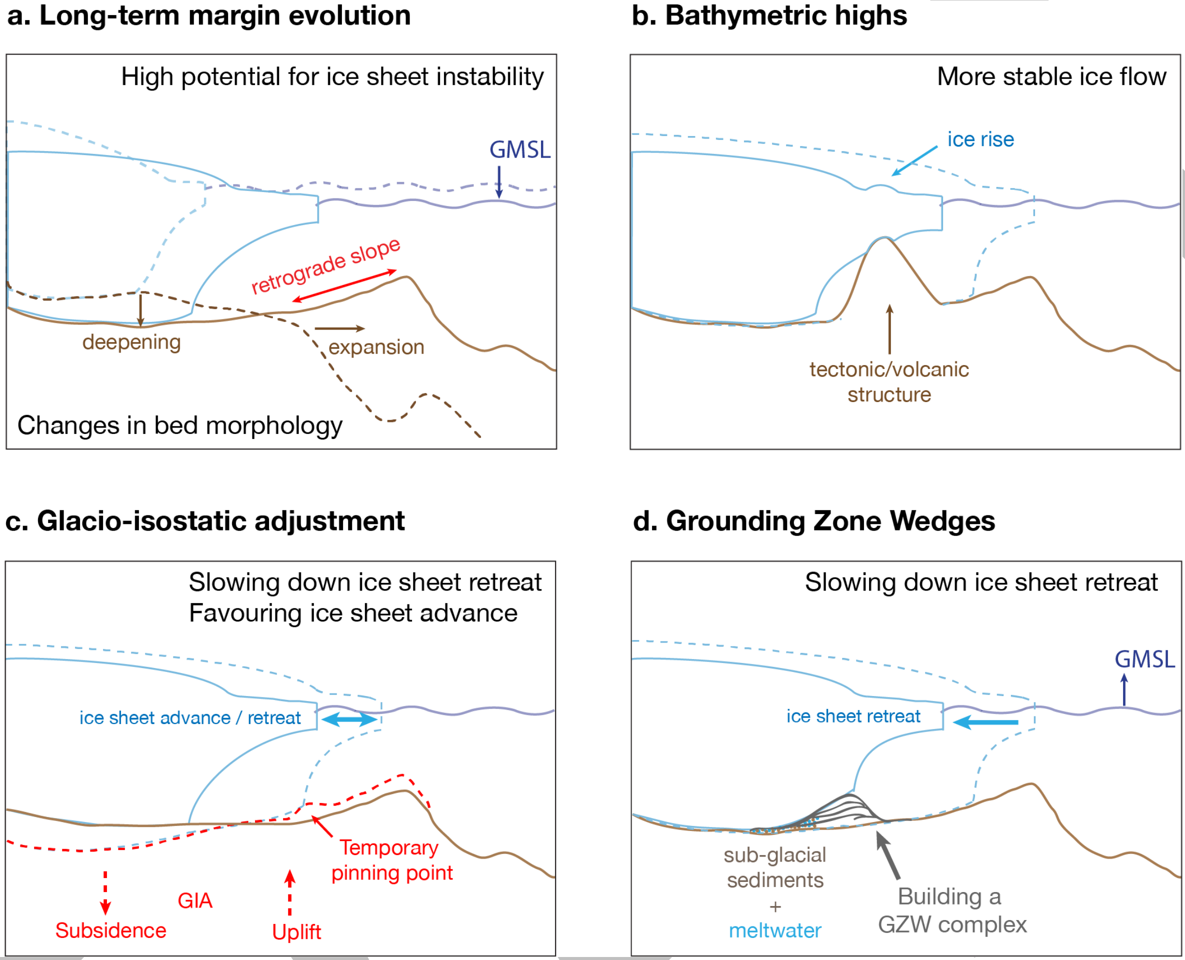


Figure 3 Schematics of the different ways by which an ice sheet can anchor on the bed. GMSL: global mean sea level; GZW: grounding zone wedge; GIA: glacio-isostatic adjustment.

Figure 4

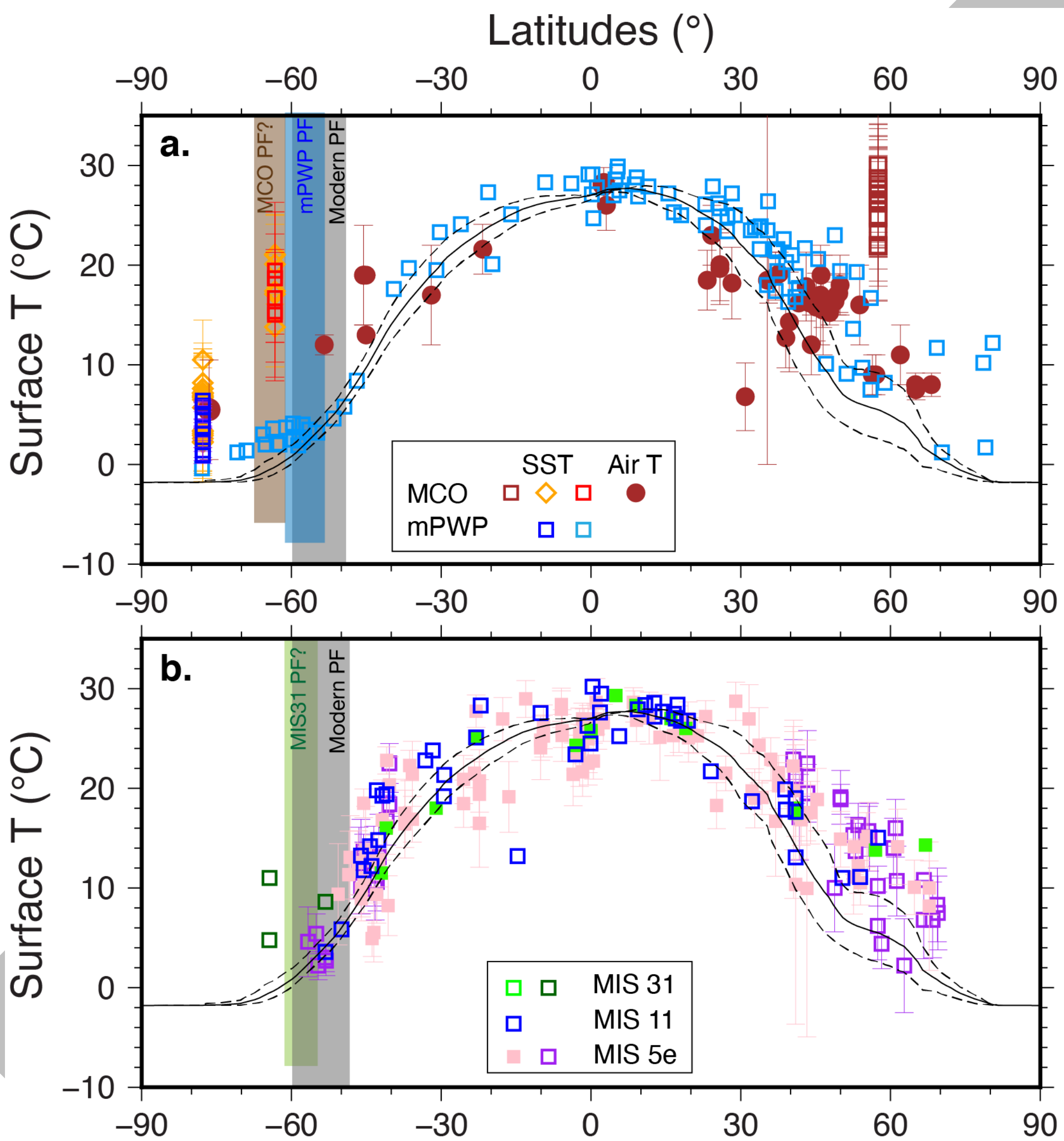


Figure 4: Proxies compilation of reconstructed meridional sea surface temperature (SST) for the MCO and the mPWP (a.), MIS 31, MIS 11 and MIS 5 (b.). Note that vertical colour bars correspond to our interpretation of approximated location of the zonally-averaged circum-Antarctic Polar Front. Mean Air Temperature (MAT) and SST proxies for the MCO are from **Golder et al. (2014)** for global compilation. Due to the paucity of data, we also include terrestrial proxies (MAT, dots). We also include from **Super et al. (2020)** BAYSPAR calibration of TEX_{86} for North Atlantic ODP Site 982 (brown open squares), from **Levy et al., (2016)** TEX_{86} Ross Sea SWT 0-200 m depth and Adélie Land margin (**Sangiori et al., 2018**; orange open diamonds) and **Hartman et al. (2018)** for TEX_{86} BAYSPAR calibrated Adélie Land margin SST records (red open squares). MAT and SST proxies for the mPWP are from **Dowsett et al. (2012)** global compilation (light blue open squares) and **McKay et al. (2012)** for TEX_{86} Ross Sea SST record (dark blue open squares). SST proxies for MIS 31 are from **Justino et al. (2017)** for global compilation and **Beltran et al. (2020)** for Antarctic Peninsula, Weddell Sea and Adélie Land margin SST records (dark green open squares). SST proxies for MIS 11 are from **Justino et al. (2017)**. SST proxies for MIS 5 are from **Capron et al. (2014)** (solid pink squares) and **Hoffman et al. (2017)** (open purple squares), both at 125 ka. Black continuous (mean annual) and dashed lines (boreal and austral summers) correspond to pre-industrial HadISST reconstruction from **Rayner et al. (2003)**. Note that many of the SST proxies plotted here from the various compilations tend to be more representative of boreal or austral summer conditions rather than of mean annual conditions.

Figure 5

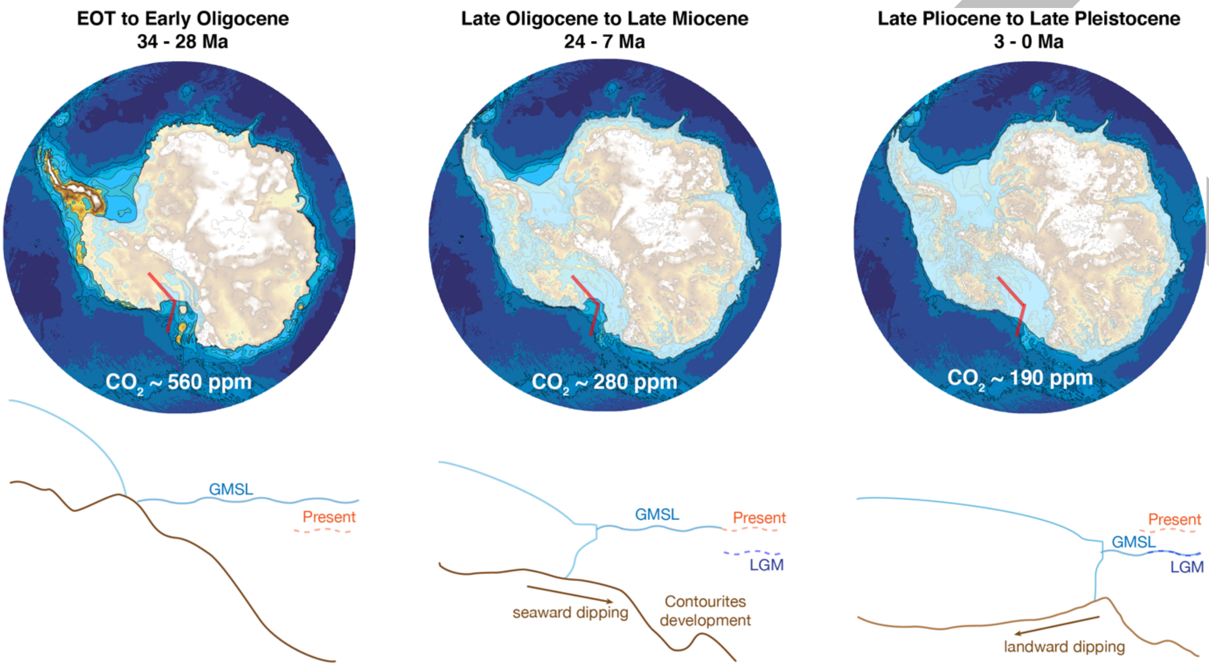


Figure 5: Simulated Antarctic ice sheet extent during past glaciations of different intervals. EOT to Early Oligocene (34 - 28 Ma): extent adapted from simulations by **Ladant et al. (2014)** with prescribed atmospheric CO_2 of 700 to 560 ppm. Late Oligocene to Late Miocene (24 - 7 Ma): extent adapted from simulations by **Gasson et al. (2016)** and **Colleoni et al. (2018b)** prescribing an atmospheric CO_2 of 280 ppm. Late Pliocene to Late Pleistocene (3 Ma to 0): extent adapted from **Colleoni et al. (2018b)** with prescribed atmospheric CO_2 of 190 ppm. Paleotopographies and bathymetries are from **Paxman et al. (2019)**. Note that ice shelves are not represented on the different panels. Schematics below each circum-Antarctic view corresponds to an idealised transect along the red lines indicated on the Antarctic maps above. Those schematics illustrate the evolution of the continental margin through time, with corresponding global mean sea level variations (GMSL, see **Figure 1c**) referred to present sea level (dashed orange line) and Last Glacial Maximum sea level (LGM, 21 ka, dashed blue line).

Figure 6

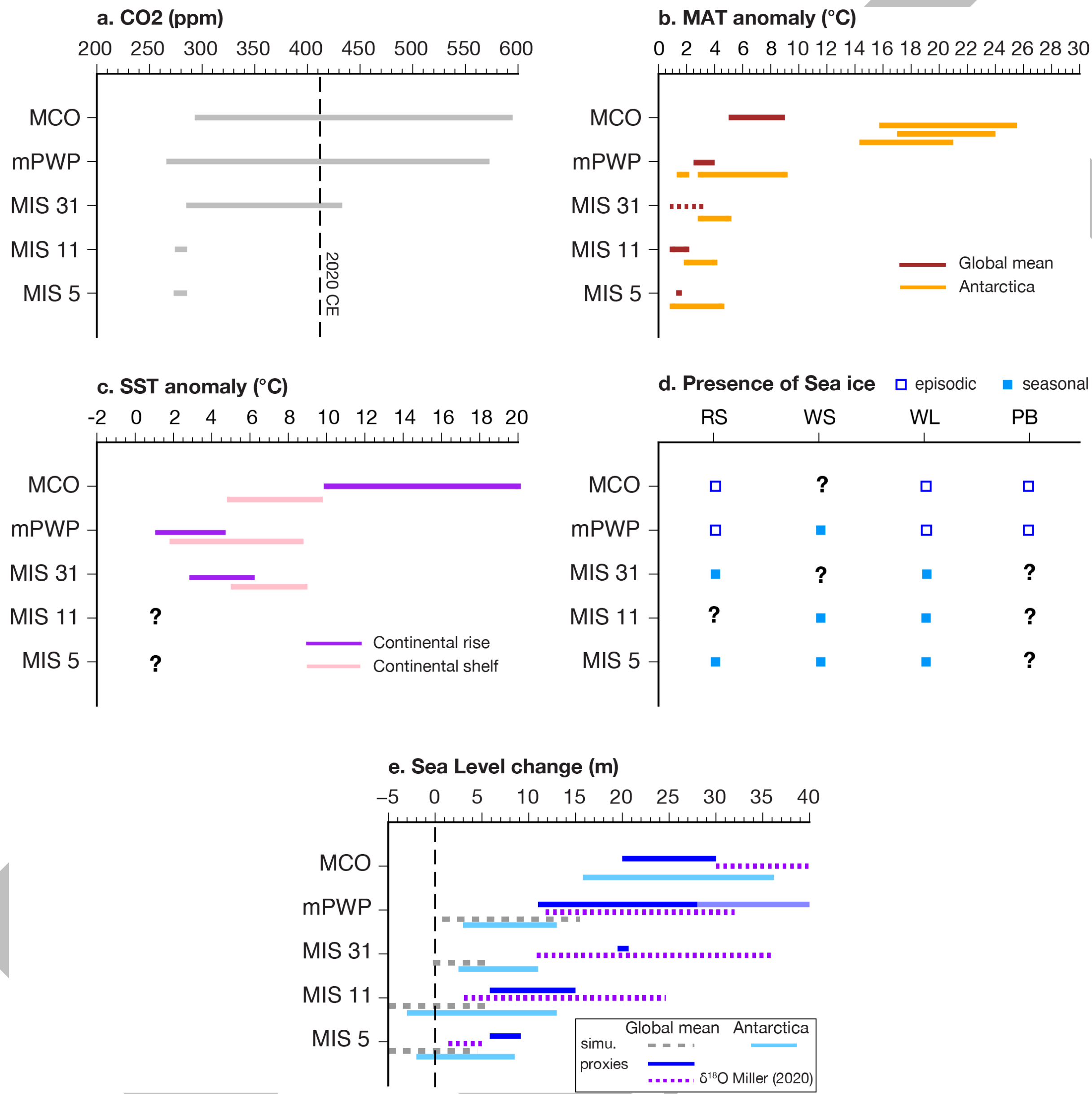


Figure 6: Main climatic indicators of each warm periods considered in the chapter (a. to d.) and associated simulated range of Antarctic contributions to global mean sea level changes (e.). Each panel shows ranges of climatic proxies or simulated quantities at global scale and for Antarctica relative to their present-day value. Note that for each range (global or Antarctic), minimum and maximum account for the minimum and maximum uncertainties of the represented proxies when found. **a.** atmospheric CO_2 levels, see Figure 1a for references. **b.** Global mean annual temperature (MAT, °C) anomaly relative to 20th century average: MCO - **Goldner et al. (2014;** mPWP - **Salzmann et al. (2013)** terrestrial proxies (see their Table S3b) and **Dowsett et al. (2012)** for SST compilation; MIS 31 - **Justino et al (2019)** averaged SST compilation also accounting for **Beltran et al. (2020)** Antarctic margin proxy-based SST. However, there is no northern high-latitude MAT or SST reconstructions. Proxies in the northern high latitudes suggest sea ice free conditions (**Detlef et al., 2018)** as during the mPWP**.** Given the high similarities with mPWP SST gradient **(Figure 4),** MIS 31 global MAT is tentatively extended to mean mPWP MAT (dashed line); MIS 11 - **Lang and Wolff (2011)** and MIS 5 - **Turney and Jones (2010)**; MAT (°C) for Antarctic region (including ice core records) anomaly are relative to 1990 at the closest weather station to the sediment cores location: MCO - **Warny et al. (2009)**, mPWP - **Passchier et al. (2011), Haywood et al. (2020)**, MIS 31 - **Scherer et al. (2008)**, MIS 11 - **Jouzel et al. (2007)**, MIS 5 - **Jouzel et al. (2007), Lang & Wolff (2011)**. **c.** Sea surface temperatures (SST, °C) anomaly

relative to present value at each core location: continental shelf values (pink) are from Cape Roberts or ANDRILL sites: MCO - **Sangiorgi et al. (2018)** are sea water temperature (0-200 m depth), mPWP - **McKay et al. (2012)**, MIS 31 - **Scherer et al. (2008)**. SST records from continental slope and rise (purple) are from: MCO - **Sangiorgi et al (2018)** are sea water temperature (0-200 m depth), **Hartman et al. (2018)**, mPWP - **Dowsett et al. (2012)** compilation of proxies below 60°S, MIS 31 - **Beltran et al. (2020)**. Note that many of the SST proxies could be representative of boreal or austral summer conditions rather than annual mean. **d.** Proxies for presence/absence of sea ice are shown for different Antarctic sectors, RS (Ross Sea), WS (Weddell Sea), WL (Wilkes Land margin), PB (Prydz Bay). Open blue squares indicate intermittent sea ice cover during the period, solid blue squares indicate seasonal sea ice (mostly no sea ice during austral summer) and question mark correspond to absence of information for the sector. For MCO - **Sangiorgi et al. (2018), Levy et al. (2016), Hannah (2006);** for mPWP - **Burckle et al. (1990), Whitehead et al. (2005), McKay et al (2012a), Taylor-Silva and Riesselman (2018);** for MIS 31 - **Bohaty et al. (1998), Scherer et al (2008), Villa et al. (2008), Beltran et al. (2020);** for MIS 11 - **Kunz-Pirrung et al. (2002), Wolff et al. (2006), Wilson et al. (2018), Escutia et al. (2011);** for MIS 5 - **Kunz-Pirrung et al. (2002), Wolff et al. (2006), Konfirst et al. (2012), Presti et al. (2011), Hartman et al. (2016), Wilson et al. (2018)**. **e.** Ranges for global mean sea level changes (GMSL, dark blue) relative to today from data and models (dotted grey). For mWPW, the proxies indicate a sea level rise up to + 40 m above present that is indicated by the transparent blue line on the plot. For MIS 31, GMSL data are uncorrected from GIA and dynamical topography. Simulated ranges of Antarctic Ice Sheet melting contributions are shown in light blue. Range from recent GMSL reconstruction based on benthic δ^{18}O records from **Miller et al. (2020a)** is shown with dotted purple line. See **Figure 1c and 1d** and the main text for references.

Figure 7

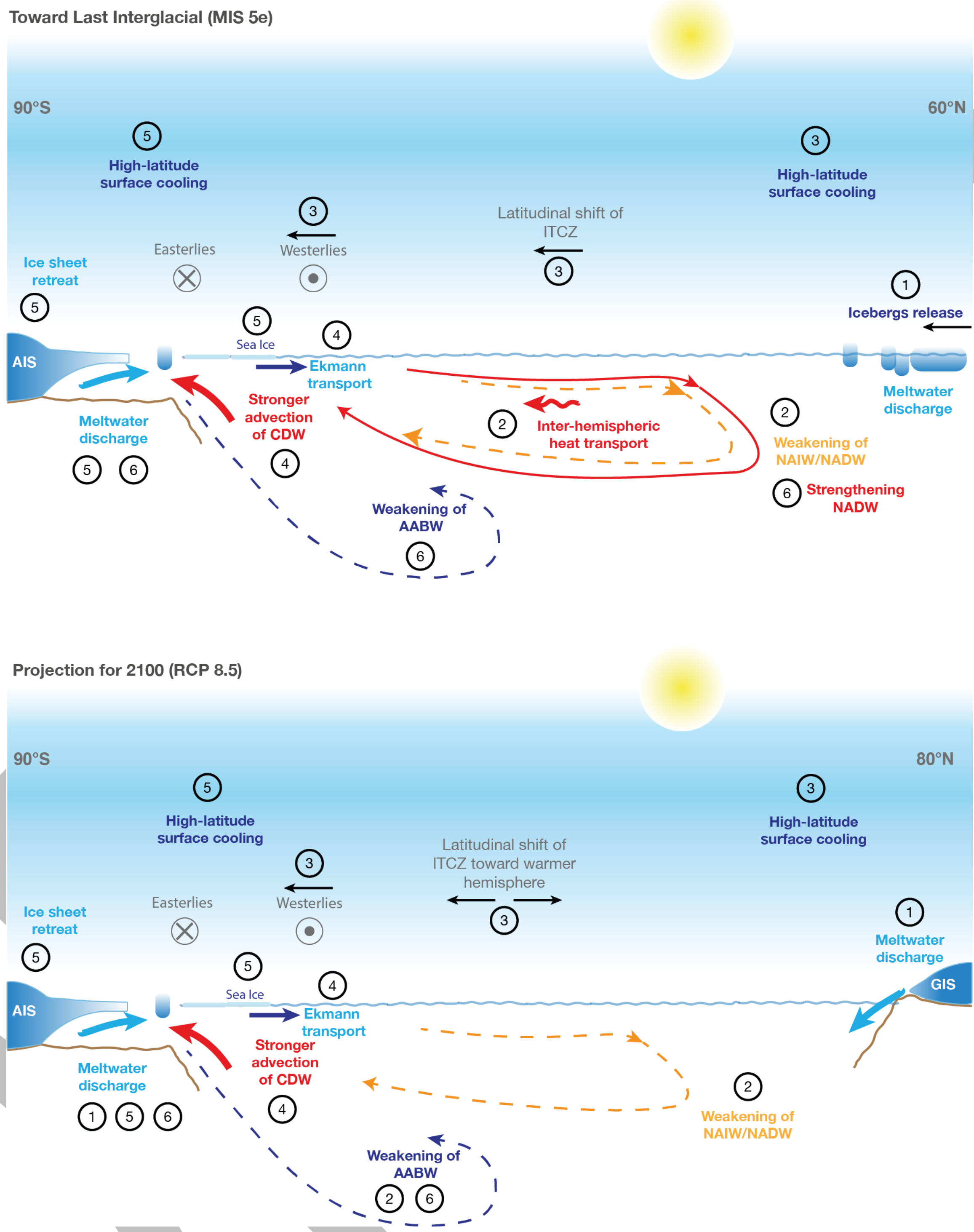


Figure 7: Cartoons for the impact of freshwater release and associated feedbacks loop at global scale and on Antarctica. Top cartoon is based on **Turney et al. (2020)** and describes the evolution of those feedback as a consequence of Heinrich event 11 (~135 ka) and subsequent evolution until MIS 5e (time frame of few millennia). Numbers indicate the order of the sequence. Bottom cartoon shows similar feedback but as projected until 2100 for the RCP 8.5 high-emission scenario (time frame of few decades) based on **Golledge et al. (2019)**. AABW: Antarctic Bottom Water; CDW: Circumpolar Deep Water; ITCZ: Inter-Tropical Convergence Zone; NAIW: North Atlantic Intermediate Water; NADW: North Atlantic Deep Water.

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