2

GEOLOGICAL AND PALEOCLIMATIC EVOLUTION OF THE SOUTHERN OCEAN–ANTARCTIC SYSTEM

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2.1 Birth of Antarctica and the Southern Ocean

2.1.1 Tectonic Evolution

The Antarctic continent has been located at or very close to the South Pole since around 300 million years ago (Ma; Müller et al., 2019). Prior to ~180 Ma, Antarctica was part of the supercontinent Gondwana, merged together with the other Southern Hemisphere landmasses of South America, Africa, India, Australia and Zealandia. The Pacific portion of West Antarctica was the only sector of Antarctica that was not landlocked, remaining adjacent to an ocean basin. The break-up of Gondwana, and the gradual isolation of Antarctica, led to the evolution of the Southern Ocean (SO) and establishment of the Antarctic Circumpolar Current (ACC). At ~180 Ma, West Gondwana (South America and Africa) separated from East Gondwana (Antarctica, India, Australia and Zealandia), opening the first ocean basin. From ~136 Ma, South America, Africa and India continued to move roughly northwards away from Antarctica. Finally, separation between Australia, Lord Howe Rise, New Zealand and Antarctica occurred from ~90 Ma. The tectonic isolation of Antarctica continued as all these plates separated further, widening all sectors of the SO.

2.1.2 The Antarctica-SO System: From Eocene to Oligocene

Coeval with the progressive tectonic isolation of Antarctica, the global climate transitioned from hot 'greenhouse' conditions during the Late Cretaceous period (~65 Ma) to warm climates of the Paleogene and early Eocene (~65–45 Ma). This tectonic and climate transition was associated with cooler conditions and the first occurrences of small ephemeral Antarctic glaciers during the middle-late

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Eocene (~40 Ma; Carter et al., 2017; Gulick et al., 2017) and large-scale Antarctic glaciation at the Eocene/Oligocene boundary (~34 Ma; Bohaty et al., 2012; Zachos et al., 1994). The Early Eocene Climate Optimum (~53–51 Ma) atmospheric CO_2 concentrations were the highest over the past 66 Ma at ~1600 ppm (CenCO_2PIP Consortium et al., 2023).

The last land bridges connecting Antarctica with Gondwanan continents finally separated during the Eocene, allowing shallow water exchange through both the Drake Passage and Tasman Gateway (Figure 0.1) by around 50 Ma (Bijl et al., 2013; van de Lagemaat et al., 2021). Further deepening of the Tasman Gateway occurred between ~33.5 Ma and 30 Ma, but the Drake Passage remained constricted until <26 Ma (van de Lagemaat et al., 2021). These tectonic events preconditioned the SO for the inception of the ACC but were insufficient to initiate a deep-reaching, modern-style ACC until the Miocene (Sangiorgi et al., 2018; Evangelinos et al., 2024). Nevertheless, these tectonic events left Antarctica entirely isolated and meant that there was a clear circumpolar ocean pathway, which enabled a major reorganisation of SO currents. As the gateways deepened, the subpolar gyres shrank and could no longer transport warm water to the Antarctic coast, resulting in 2–4°C cooling of Antarctic surface waters (Sauermilch et al., 2021). The opening of the Drake Passage and Tasman Gateway, combined with decreasing atmospheric greenhouse gas concentrations, likely both contributed to the glaciation of Antarctica.

2.2 Gradual Cooling of the Antarctica–SO System: From EOT to Late Miocene

2.2.1 From EOT to Late Oligocene (34–23 Ma): Gradual Global Cooling

In Antarctica, ice likely first nucleated on elevated coastal regions and on high topography in the late Eocene (Baatsen et al., 2024). A rapid step-increase in Antarctic ice extent and volume occurred at the Eocene–Oligocene transition (EOT; 34.0–33.5 Ma). Hypotheses for the cause of this expansion include: (1) a large drop in atmospheric CO_2 levels; (2) the development of a modern-like ACC (Kennett, 1977); and (3) internal carbon-cycle feedbacks (Coxall et al., 2005). Numerical oceanic and ice-sheet simulations suggest that the tectonic changes in the main SO gateways alone cannot result in sufficient cooling to explain the ice expansion in Antarctica (e.g., Sauermilch et al., 2021; DeConto and Pollard, 2003). Those simulations show that a substantial drop in CO_2 was necessary to trigger large glaciations and that the opening of the ocean gateways strengthened the regional cooling over Antarctica. The atmospheric CO_2 threshold required to trigger large-scale glaciations over Antarctica remains undetermined. Numerical climate and ice sheet simulations suggest a range of atmospheric CO_2 thresholds from 500 to 900 ppm (Ladant et al., 2014) to glaciate East Antarctica.

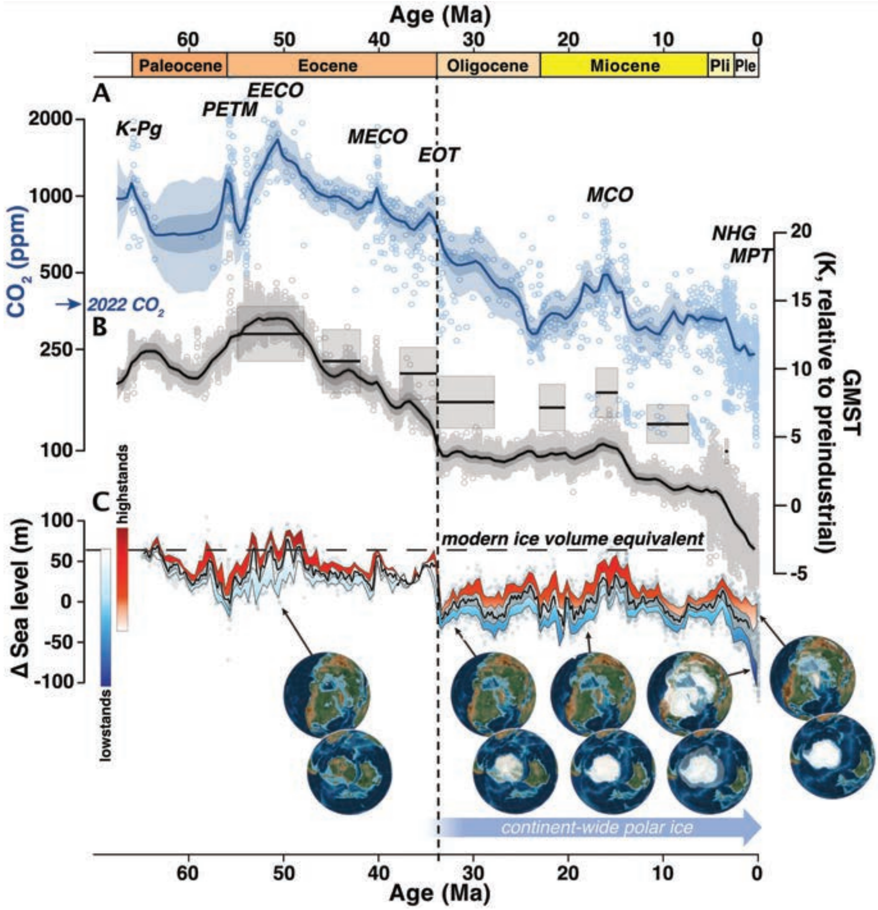

FIGURE 2.1 (A) Atmospheric CO_2 multi-proxy compilation showing a 500-kyr mean statistical reconstruction (median and 50% (dark blue) and 95% (light-blue) (CenCO_2PIP Consortium et al., 2023) and the climate events mentioned in this chapter: EECO, Early Eocene Climatic Optimum; MECO, Middle Eocene Climatic Optimum; EOT, Eocene/Oligocene Transition; MCO, Miocene Climatic Optimum; NHG, onset of Northern Hemisphere Glaciation; and MPT, Mid-Pleistocene Transition. (B) Global mean sea-surface temperature estimated from benthic $\delta^{18}O$ data (Westerhold et al., 2020); individual proxy estimates as symbols, and statistically reconstructed 500-kyr mean values shown as the continuous curve, with 50 and 95% credible intervals. Grey boxes show surface temperature estimates from Ring et al. (2022). (C) Sea level reconstruction (Miller et al., 2020); grey dots are raw data; the solid black line reflects median sea level in a 1-Myr running window. Paleogeographic reconstructions and the growing presence of ice sheets in polar latitudes are shown on the globes (Scotese, 2021). Figure modified from CenCO_2PIP Consortium et al. (2023).

Paleo sea-level records and benthic oxygen isotope records both suggest that Antarctic ice volume fluctuated significantly throughout the Oligocene (34–23 Ma) (Zachos et al., 2008; Rohling et al., 2022). The Cape Roberts sediment records (western Ross Sea) revealed numerous orbitally-paced cycles attributed to oscillations in East Antarctic Ice Sheet (EAIS) extent and volume between 34 Ma and 17 Ma (Naish et al., 2001; Barrett, 2008; Galeotti et al., 2016). Although the magnitude is still debated, estimates of eustatic sea level fluctuations over this period suggest that Antarctic ice volume varied between 15% and 60% of that of the modern ice sheet (Pekar et al., 2006). Coastal temperatures cooled progressively through the Oligocene (Barrett, 2008), although the cooling was subdued compared to elsewhere in the SO (Duncan et al., 2022), and plant species diversity and abundance remained relatively high (Prebble et al., 2006), suggesting that ice coverage remained limited.

The ice sheets of the Oligocene waxed and waned at orbitally-paced timescales and were characterised by a predominantly warm-based glacial regime, in contrast to the modern AIS, which lacks comparable meltwater (Hambrey et al., 1991). This led to selective erosion of the bed (Thomson et al., 2013), as evidenced by the large volumes of glaciogenic sediment deposited on the East Antarctic continental shelf and upper slope at this time (Hochmuth et al., 2020). Sedimentological evidence from marine sediments indicates that the EAIS reached the coast during the early Oligocene (Passchier et al., 2017). The extent of the West Antarctic Ice Sheet (WAIS) during the Oligocene is more uncertain and evidence is contradictory. Sedimentation rate records from the Weddell Sea and Ross Sea suggest increased erosion in West Antarctica following the EOT, implying the expansion of an ice sheet over West Antarctica, fostered by a paleotopography with a land area ~20% larger than today (Wilson et al., 2013; Paxman et al., 2019). By comparison, present-day West Antarctic bed topography is mostly below sea level (Morlighem et al., 2020; Figure 2.2). Other geological evidence contradicts this analysis and suggests the existence of shallow bathymetric troughs and inland seaways in West Antarctica that may have facilitated warm water intrusions, thus inhibiting the expansion of an ice sheet over West Antarctica during the early Oligocene (Coenen et al., 2020; Uenzelmann-Neben et al., 2022).

While the AIS extent remained limited, water-mass signatures retrieved from SO sediment records spanning the last 31 Myr clearly suggest that the ACC remained shallow during the Oligocene until the mid-Miocene (Evangelinos et al., 2024). These data also suggest that a water mass resembling Circumpolar Deep Water (CDW), as part of the deep layer of the ACC, already formed at that time, fostered by efficient deep water exchange between the Atlantic and Indian Ocean, but not between the Indian and Pacific Ocean.

2.2.2 Gradual Cooling toward Polar Conditions: The Miocene Period (23–5.3 Ma)

The early to mid-Miocene (23–14 Ma) provides insight into the behaviour of the AIS during a period when Antarctic paleotopography continued to evolve towards the

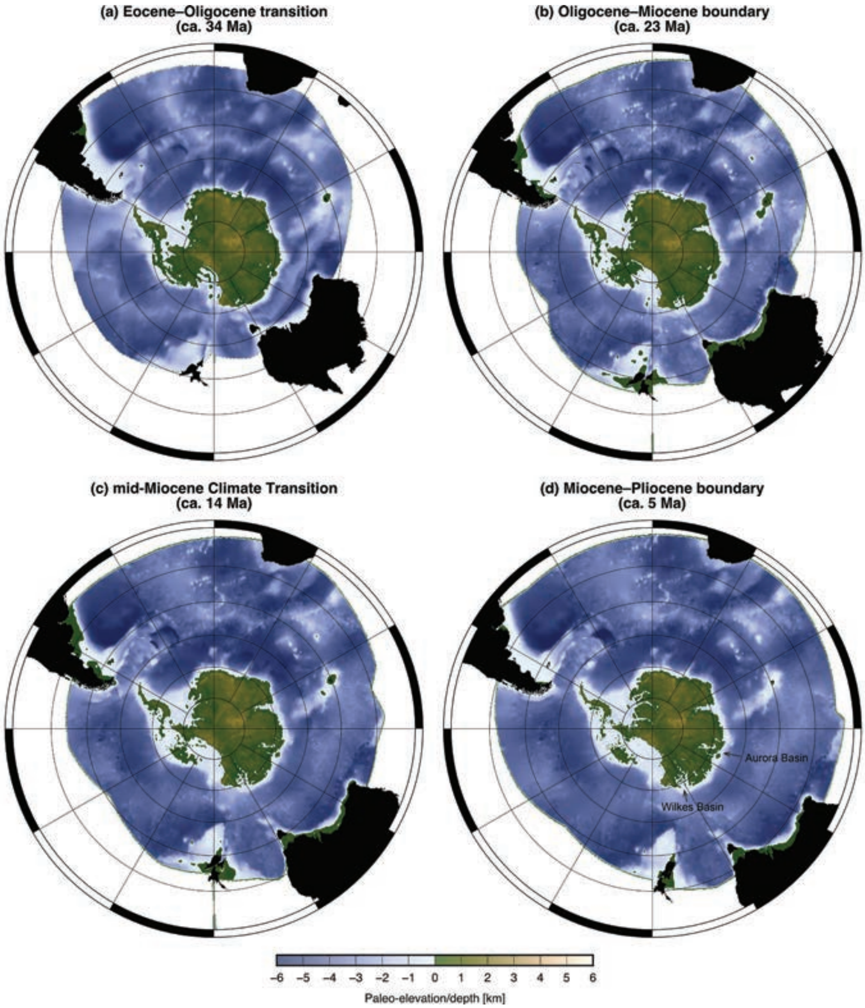

FIGURE 2.2 Reconstructed evolution of the paleogeography, paleotopography and paleobathymetry of Antarctica and the Southern Ocean from (a) the Eocene–Oligocene transition (~34 Ma) through (b) the Oligocene–Miocene boundary (~23 Ma) and (c) the mid-Miocene Climate Transition (~14 Ma) to (d) the Miocene–Pliocene boundary (~5 Ma) (Paxman et al., 2019; Hochmuth et al., 2020).

modern configuration and atmospheric CO_2 fluctuated between 200 and 400 ppm, with potential peaks at ~800 ppm during interglacial periods of the mid-Miocene Climatic Optimum (MCO; 17.0–14.8 Ma) (CenCO_2PIP Consortium et al., 2023). During MCO warm intervals, mean summer temperatures in the McMurdo Dry Valleys were 5–7°C, up to 20°C warmer than the present-day (Lewis et al., 2008;

Lewis and Ashworth, 2016). Sea-surface temperatures during the MCO peaked at around 11–17°C off the Adelie Coast (Sangiorgi et al., 2018) and 6–10°C in the Ross Sea (Levy et al., 2016). Warm interglacials of the MCO were characterised by the presence of temperate tundra vegetation (*Nothofagus*, shrubs, grasses and mosses) in coastal lowlands and warm oligotrophic waters in the SO (Warny et al., 2009; Sangiorgi et al., 2018). Neodymium isotope signatures suggest that the strengthening of Atlantic Meridional Overturning Circulation (AMOC) (Via and Thomas, 2006) fostered the inflow of Atlantic deep water masses into the CDW during the early Miocene (Evangelinos et al., 2024).

Far-field eustatic sea-level reconstructions and benthic oxygen isotope records suggest sea-level oscillations of 40–60 m (Lear et al., 2008; Miller et al., 2020), implying periods of near-complete loss of Antarctic land ice. Marine-based portions of the ice sheet repeatedly retreated inland, leaving open-water conditions in most of the marine-based sectors of Antarctica during warm intervals of the MCO (Naish et al., 2001, Sugden and Denton, 2004; Pierce et al., 2017). Geochemical and petrographic analysis of Ross Sea sediment records indicates that a larger-than-present WAIS expanded to cover most of the shallow continental shelf at around 17.8–17.4 Ma (up to 15 metres sea level equivalent compared to 4.5 m today; Marschalek et al., 2021). Terrestrial portions of the EAIS persisted during retreat phases, with the ice margins receded from the coastline and surrounded by tundra (Levy et al., 2016; Sangiorgi et al., 2018; Chorley et al., 2022). The AIS volume during the MCO is simulated to have been 85–90% of the modern-day EAIS (Halberstadt et al., 2021), with an associated sea-level contribution estimated at 30–36 m (Gasson et al., 2016). Total melting of the WAIS during peak warm intervals of the MCO combined with partial loss of the EAIS can explain the inferred magnitude of MCO sea-level oscillations.

The mid- to late-Miocene was characterised by the gradual establishment of an arid polar climate and a persistent continental-scale AIS. During the mid-Miocene Climate Transition (MCT; ~14.8–13.8 Ma), terrestrial evidence from the Transantarctic Mountains indicates cooling of 8°C (Lewis et al., 2008; Lewis and Ashworth, 2016). Although plant and animal fossil records indicate that coastal areas of Antarctica were still ice-free during MCT interglacials (Lewis and Ashworth, 2015; Sangiorgi et al., 2018), glacial intervals progressively intensified, causing the AIS to expand across the marine-based sectors (Shevenell et al., 2008; Holbourn et al., 2018). This cooling was accompanied by the establishment of widespread pan-Antarctic perennial sea-ice cover for the first time since the early Oligocene (Levy et al., 2016; Bijl et al., 2018; Sangiorgi et al., 2018; Halberstadt et al., 2021). By the end of the MCT, a modern-like CDW was established, as a result of the potential inflow of Pacific deep waters, as well as the strengthening of Atlantic and Indian inflows, into CDW via the Drake Passage (Evangelinos et al., 2024). Geological and geochemical evidence suggests major tectonic changes in the Drake Passage, such as the development and deepening of an oceanic gateway along the southern Scotia Ridge until after 12 Ma that would have fostered such an inflow of Pacific deep waters (Dalziel et al., 2013). Based on the neodymium isotope records, the emergence of a well-mixed CDW – as part of the ACC – connecting all three ocean basins was established at the end of the MCT ~12 Ma, but granulometry of the SO sediment records indicates that the ACC speed remained low until at least the late Miocene (~10 Ma; Evangelinos et al., 2024).

After the MCT and throughout the late Miocene, the terrestrial sectors of the ice sheet became increasingly cold-based and less-erosive (Sugden et al., 1999). The AIS is inferred to have stabilised during the late Miocene, with cosmogenic nuclide evidence suggesting that the EAIS draining into the Ross Sea has not retreated significantly onto land since ~8 Ma (Shakun et al., 2018). SO sediment records indicate that ACC depth and speed resembled the present-day by ~10 Ma suggesting that the establishment of polar conditions in the Antarctic since the MCT led to the steepening of equator-to-pole air and sea temperature and density gradients and, hence, to the strengthening of the westerly winds driving the ACC (Evangelinos et al., 2024).

2.3 The Emergence of a Bi-Polar World: The Plio-Pleistocene Period

2.3.1 The Pliocene (5.3–2.6 Ma): A Future Analogue of the SO–Antarctic System?

During the Pliocene, and especially the mid-Pliocene Warm Period (3.3–3.0 Ma), reconstructions suggest that atmospheric CO_2 concentrations were substantially higher ~370 ppm than pre-industrial value of 280ppm (de la Vega et al., 2020), leading to average global temperatures that were 2–3°C warmer. This is comparable to low-end emission scenarios for the end of the 21st century (shared socioeconomic pathways SSP2–2.6 to SSP2–4.5; Meinshausen et al., 2020). Furthermore, the tectonic boundary conditions were similar to present (Haywood et al., 2013), but with smaller Antarctic and Greenland Ice Sheets (Dutton et al., 2015), so this interval provides a useful geological analogue for ongoing anthropogenic climate warming.

The mid-Pliocene was characterised by globally weak meridional temperature gradients and reduced sea-ice concentrations relative to modern in both hemispheres (Whitehead et al., 2005; Knies et al., 2014). Warmer SO sea-surface temperatures, reduced sea ice (Escutia et al., 2009) and more southerly westerly winds relative to pre-industrial conditions (Li et al., 2015; Abell et al., 2021) would have facilitated access of warm CDW onto the continental shelves, likely also contributing to a more retreated AIS (Naish et al., 2009; Cook et al., 2013). Marine sedimentary and geochemical records indicate that there may have been episodic retreat and/or collapse of the marine-based portions in both West Antarctica (e.g., Ross Sea, Naish et al., 2009) and East Antarctica (Adelie/George V Land, Cook et al., 2013; Patterson et al., 2014. Wilkes Land, Williams et al., 2010). Data-model comparisons based on benthic carbon isotopes and simulated ocean ventilation ages are consistent with enhanced Antarctic Bottom Water (AABW) formation and increased ventilation of the SO, in agreement with mid-Pliocene proxy reconstructions showing weak SO stratification (Zhang et al., 2013).

Reconstructed sea-level estimates for the warm intervals of the mid-Pliocene (3.3–3.0 Ma) range from 5 to 25 m above present (Dumitru et al., 2019; Grant et al., 2019). Numerical simulations show a similarly wide range of Antarctic sea-level contributions, such as 6–21 m (Golledge et al., 2017a, DeConto et al., 2021). It is difficult to disentangle the magnitude of the AIS contribution to global mean sea-level rise at that time, due to proxy uncertainties and challenges related to glacio-isostatic adjustment and dynamic topography. The current WAIS hosts an ice volume of about 5.3 m sea level equivalent and the EAIS holds a volume of 52.2 m sea level equivalent (Morlighem et al., 2020), with a further 7.42 m sea level equivalent in Greenland (Morlighem et al., 2017). Even if the entire Greenland Ice Sheet was absent in the mid-Pliocene, significant ice loss from the marine-based margins of East Antarctica as well as loss of the entire WAIS would be required to explain the upper sea-level estimates (Miller et al., 2012; Grant et al., 2019).

2.3.2 Establishment of the Bi-Polar World (2.7 Ma–Present): A Two-Step Process

During the Pliocene-Pleistocene Transition, global climate abruptly cooled, which enabled the progressive expansion of the Northern Hemisphere ice sheets at ~2.7 Ma (Ravelo et al., 2004) and the establishment of the Pleistocene bi-polar world (Zachos et al., 2001). Numerous factors have been hypothesised to have contributed to the onset of cooling and Northern Hemisphere glaciation, including insolation changes (Maslin et al., 1998) and a decline in atmospheric CO_2 concentrations to a threshold of ~300 ppm between 2.8 and 2.5 Ma (DeConto et al., 2008; Lunt et al., 2008; Hönisch et al., 2009). Ice volume and sea-level fluctuations, which were previously dominated by advance and retreat of the AIS, became largely influenced by the periodic growth and decay of the Laurentide and Eurasian ice sheets (Rohling et al., 2022).

Gradual increases in the severity of glacial conditions (colder with larger ice sheets) occurred across the Mid-Pleistocene Transition (MPT) from 1.2 Ma to 800,000 years ago, when the periodicity of cold glacial to warm interglacial climate states switched from 41,000 years to 100,000 years (Ruddiman et al., 1989). Major changes in ocean circulation and carbon cycling in the SO likely contributed to cooling across the MPT via increased seasonal sea-ice extent (McKay et al., 2012a), increased water-column stratification south of the Polar Front (Sigman et al., 2004, Hasenfratz et al., 2019) and increased biological productivity north of the Polar Front associated with enhanced iron fertilisation (Cortese et al., 2004; Martínez-Garcia et al., 2011). For a review of the mechanisms associated with the aforementioned climate feedbacks, see Berends et al. (2021). New continuous million-year-old ice core records are anticipated in the coming years and will improve our understanding of the processes that occurred across the MPT.

2.3.3 Antarctic Ice Sheet Dynamics in the Late Pleistocene

The Pleistocene epoch has the best documented global and regional climate records, including high-resolution Antarctic ice core records going back 800 kyr, so climate forcing and changes in the ice sheet can be resolved on orbital (tens of thousands of years) to centennial timescales. Furthermore, global boundary conditions were similar to the modern-day, including AIS subglacial topography, continental configurations and global ocean circulation patterns, such that an understanding of the climate system behaviour during this period is useful for informing future climate states.

The marine-based portions of the AIS are susceptible to retreat due to atmosphere and ocean warming through a combination of ice-shelf thinning or collapse, and marine ice sheet instability (Jamieson et al., 2012). Pleistocene interglacials provide a good target to understanding climate thresholds because certain 'super-interglacials', such as Marine Isotope Stage (MIS) 5e (129–116 ka) and MIS 11 (424–395 ka), were warmer than the pre-industrial Holocene by ~0.5–2°C globally, and by up to 2–4°C for a few thousand years in Antarctica (Jouzel et al., 2007; Yin and Berger, 2015). Global mean sea-level reconstructions require ice loss from the AIS during peak warm conditions for MIS 5e (<5m, Dumitru et al., 2023; 6–9m, Dutton et al., 2015), beyond partial ice loss from Greenland alone. Asynchronous meltwater contributions from Greenland and the AIS have been proposed, including a dominant Antarctic sea-level contribution to the early MIS 5e sea-level peak at ~129 ka in response to ocean warming (Rohling et al., 2019; Barnett et al., 2023). These past changes are not direct analogues for present-day or near-future anthropogenic climate change but are invaluable for assessing millennial-scale ice-sheet behaviour and the processes and feedbacks involved. Notably, such behaviour cannot be determined from the relatively short record of satellite observations, since the ice sheet and ocean remain out of equilibrium with the climate due to their long response timescales.

Given the sensitivity of West Antarctic catchments, such as the Pine Island/Thwaites Glacier system (Amundsen Sea Embayment) and the Siple Coast (Ross Sea) to ocean warming in numerical models (Golledge et al., 2017b; Clark et al., 2020), a partial or full collapse of the WAIS has been suspected for recent warm interglacials such as MIS 5e and/or MIS 11. Collapse of the WAIS during MIS 11 has also been simulated in models, which suggest an overall Antarctic sea-level contribution during this interval of ~4–8 m (Mas e Braga et al., 2021). However, geological evidence supporting or refuting WAIS collapse remains equivocal. The presence of a trans-Antarctic seaway between the Weddell and Ross seas during at least one late Pleistocene interglacial has been suggested (Scherer et al., 1998; Barnes and Hillenbrand, 2010; Lau et al., 2023), while sediment cores from the Ross Sea shelf imply loss of the Ross Ice Shelf during MIS 5e or MIS 7, providing indirect evidence for WAIS deglaciation (McKay et al., 2012b). West Antarctic blue-ice records also provide evidence for regional climate changes expected from WAIS collapse (Steig et al., 2015), and for ice sheet changes in the Weddell Sea Embayment (Turney et al., 2020; Figure 2.3b).

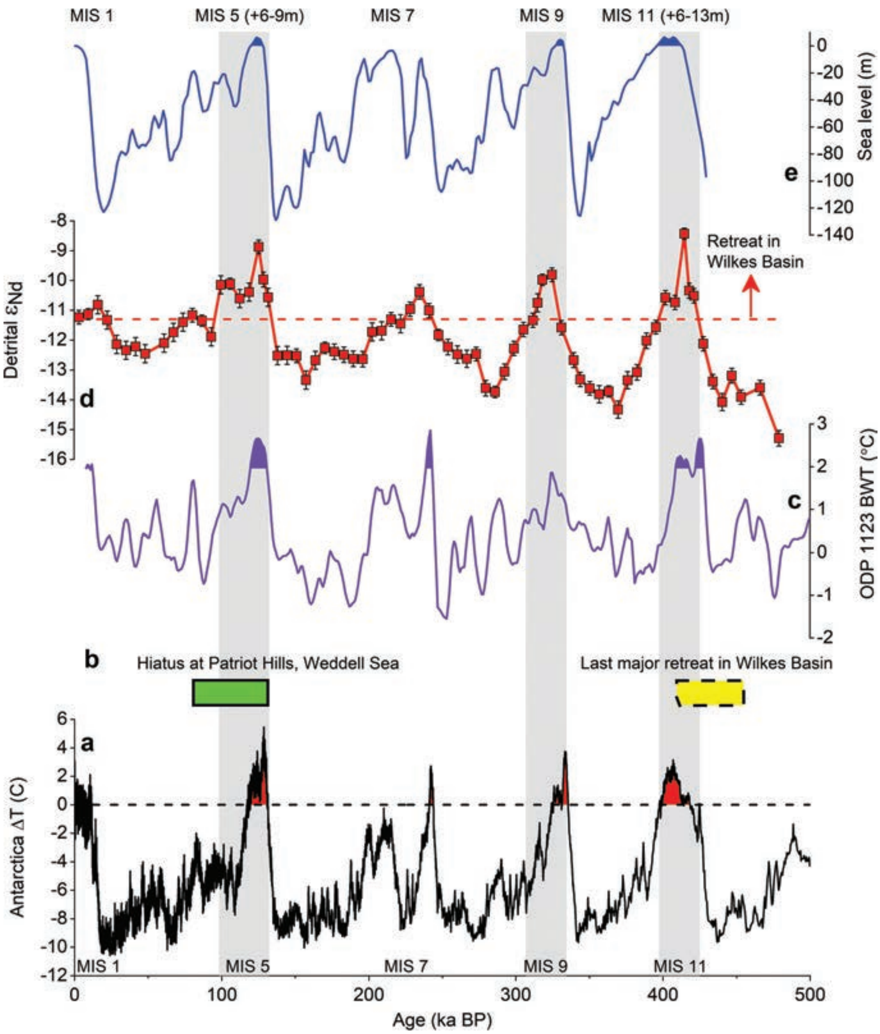

FIGURE 2.3 Late Pleistocene Antarctic Ice Sheet evolution. (a) Antarctic temperature change (ΔT) from δD in EDC ice core (Jouzel et al., 2007). (b) Timings of a hiatus at Patriot Hills, indicating grounding line retreat in the Weddell Sea sector of the WAIS (Turney et al., 2020), and the last major retreat in the Wilkes Subglacial Basin of the EAIS (95% confidence interval; Blackburn et al., 2020). (c) Southern Ocean bottom-water temperature (BWT) from benthic foraminiferal Mg/Ca at ODP Site 1123 (Elderfield et al., 2012). (d) Detrital sediment Nd isotopes at IODP Site U1361 (Wilson et al., 2018), indicating ice sheet retreat (grey bars) in the Wilkes Subglacial Basin. (e) Global sea-level proxy from benthic $\delta^{18}O$ (Waelbroeck et al., 2002), with marine isotope stages (MIS) and selected sea-level estimates (Dutton et al., 2015). Shading in (a, c, e) and red dashed line in (d) enable comparison to late Holocene values.

Much of the EAIS is terrestrial-based and these portions appear to have remained intact since the late Miocene (Shakun et al., 2018). However, nearly one-third of the ice in East Antarctica is located in marine-based catchments within the Wilkes, Aurora and Recovery Subglacial Basins (Figure 0.1), which may have experienced variability during the Pleistocene. There are only limited sedimentary records from these remote regions, of which the Wilkes Subglacial Basin is the best investigated due to the Integrated Ocean Drilling Program Expedition 318 (Escutia et al., 2011). Sediment provenance records from offshore of the Wilkes Subglacial Basin indicate ice-margin retreat during MIS 5e, MIS 9 and MIS 11 (Wilson et al., 2018; Iizuka et al., 2023) (Figure 2.3). Furthermore, differing responses to these warm interglacials compared to the Holocene and MIS 7 indicate that retreat may have occurred when Antarctic air temperatures were at least 2°C warmer than pre-industrial for ~2,500 years or more (Figure 2.3). Despite suggesting a contribution to late Pleistocene interglacial sea levels from the EAIS, those data are not able to quantify the extent of retreat or the sea-level contribution. Independent evidence from the geochemistry of subglacial opal and calcite precipitates in the Wilkes Subglacial Basin suggests a major ice-margin retreat during MIS 11 (Figure 2.3b), with a potential sea-level contribution of up to 3–4 metres, but only minor changes during subsequent interglacials (Blackburn et al., 2020). In addition, ice-core evidence for ice-sheet elevation at Talos Dome during recent interglacials also supports only modest retreat (rather than collapse) during MIS 5e and MIS 9, restricting sea-level contributions from the Wilkes Subglacial Basin to a maximum of ~0.5–1 metres at those times (Sutter et al., 2020; Crotti et al., 2022). MIS5e ice sheet modelling further suggests localised glacier acceleration and thinning enhanced inland erosion, coeval with sedimentary records (Wilson et al., 2018), with insufficient atmospheric warming given the known topography boundary conditions to allow ocean-driven inland retreat (Golledge et al., 2021).

Near-future changes are likely in the marine basins of Antarctica, due to their vulnerability to ocean-driven basal melting and run-away grounding line retreat. However, each catchment has a different sensitivity to climate and ocean forcing (Golledge et al., 2017a). New direct glaciological combined with geological evidence of past ice-sheet behaviour and regional ocean dynamics are needed to inform tipping points and thresholds on a sector-by-sector basis.

2.3.4 The Bipolar Seesaw and Atmosphere-Ocean-Ice Sheet Interactions in the Late Pleistocene

The bipolar seesaw invokes the interhemispheric redistribution of atmospheric and oceanic heat on centennial to millennial timescales via changes in the AMOC (Stocker and Johnsen, 2003) to explain the anti-phase temperature patterns observed in Greenland and Antarctic ice cores (Blunier et al., 1998; EPICA Community Members, 2006). In this hypothesis, a strong AMOC causes warming in the North Atlantic and cooling in the SO, whereas a collapse of AMOC and reduced

North Atlantic Deep Water formation leads to cooling in the North Atlantic and warming in the SO (Broecker, 1998; Stocker and Johnsen, 2003). The forcing generating this seesaw could originate from processes affecting deep-water formation in both the North Atlantic and/or the SO, while both oceanic and atmospheric processes probably played a role in transmitting such signals (WAIS Project Members, 2015; Buizert et al., 2018). In the case of a weakened AMOC, the build-up of ocean heat north of the ACC is transferred poleward across the ACC via eddies to Antarctica, which melts sea ice and sets up the ice-albedo feedback that results in further warming (Pedro et al., 2018). SO observations today show warming and freshening trends around the Antarctic margin (Bronselaer et al., 2020), impacting SO overturning circulation (Figure 2.4).

The bipolar seesaw can cause significant local warming of the ocean and the atmosphere in the vicinity of Antarctica that is above the expected ‘background’ levels for a given climate state (Holden et al., 2010), which could help explain peak Pleistocene interglacial Antarctic temperatures that were ~2–4°C warmer than pre-industrial conditions (Marino et al., 2015). Such warming could be crucial for driving both atmospheric and ocean mechanisms that influence ice-sheet stability (Clark et al., 2020). For example, Antarctic ice loss during early MIS 5e has been proposed during and/or following Heinrich Stadial 11 when the AMOC was perturbed by freshwater released by Northern Hemisphere ice sheets (Rohling et al., 2019; Turney et al., 2020; Figure 2.5), and several other climate states such as the last deglaciation (Golledge et al., 2014; Weber et al., 2014). The bipolar seesaw can act as a positive feedback mechanism for Antarctic ice mass loss, whereby enhanced upper-ocean stratification around Antarctica and southward shifts in Southern Hemisphere westerly winds (Menviel et al., 2018) arise in response to weakening of the AMOC, enhancing upwelling and incursions of warm CDW onto and across Antarctic continental shelves (Fogwill et al., 2014) and reducing AABW formation and abyssal ocean ventilation (Phipps et al., 2016; Figure 2.6). Such wind shifts have been observed over recent decades (Herraiz-Borreguero et al., 2022). The resemblance between processes that operated during MIS 5e and those characterising the present-day or near future suggests that inter-hemispheric coupling could play a major role in regulating the future of the Antarctic system.

These processes are consistent with those captured in recent observations and modelled outcomes for future ice-ocean feedbacks and AABW formation (Silvano et al., 2018; Bronsalear et al., 2018; Figure 2.5c–e). Furthermore, there is also evidence for AMOC instability within other recent interglacial periods (e.g., MIS 11; Galaasen et al., 2020; Glasscock et al., 2020), which suggests that it may be a persistent feature of the climate system, and one that could reoccur in the future. Current ice sheet simulations are typically run until 2100 (e.g., Golledge et al., 2019), preventing a full assessment of how bipolar seesaw mechanisms will impact Antarctic contributions to sea-level rise in the coming centuries.

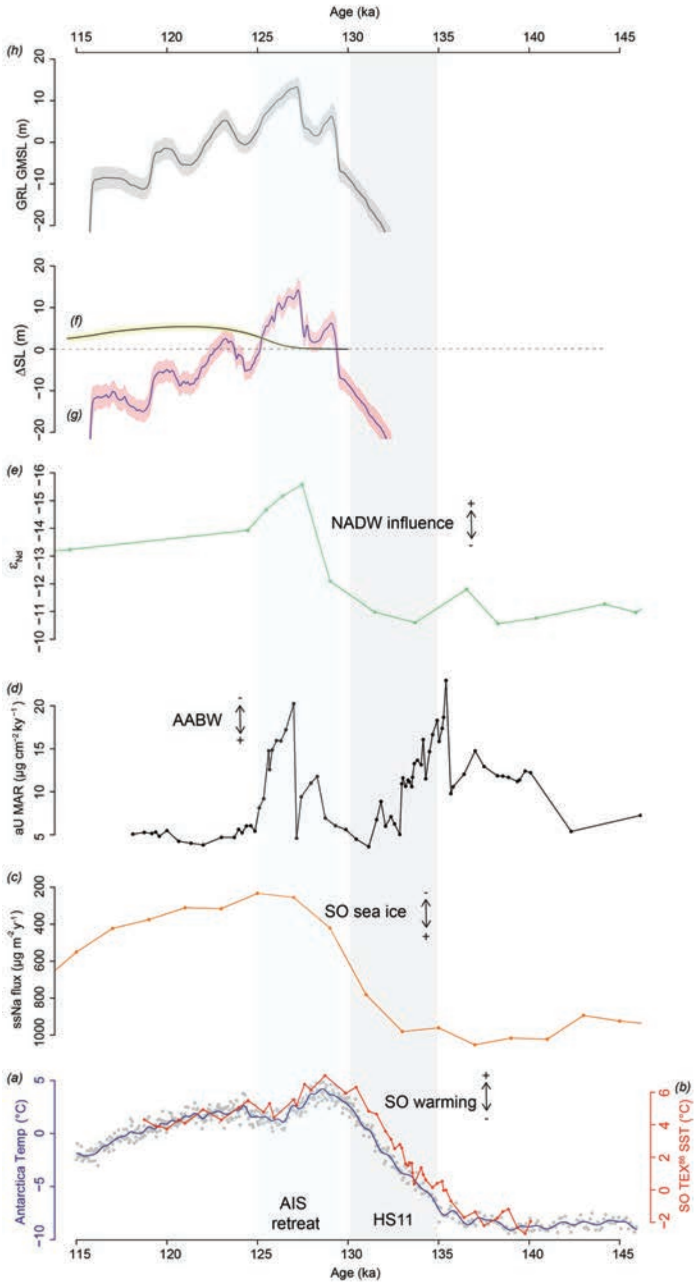

FIGURE 2.4 The Antarctic Ice Sheet contribution to MIS 5e sea-level rise following Heinrich Stadial 11 (130–135 ka). (a) Antarctic air temperature (Jouzel et al., 2007), (b) Southern Ocean TEX_{86}^{L} sea-surface temperature (Hayes et al., 2014), (c) Antarctic sea-ice extent inferred from sea salt sodium flux (ssNa) (Wolff et al., 2006), (d) authigenic U accumulation rate at ODP Site 1094 in the Southern Ocean (Hayes et al., 2014), (e) Nd isotopic composition tracing AMOC changes from ODP Site 1063 on the Bermuda Rise, NW Atlantic (Deaney et al., 2017), (f) Greenland Ice Sheet contribution to global sea level from the model-data assimilation of Yau et al. (2016), (g) AIS contribution to sea-level rise based on the difference between (h) the Red Sea KL11 global sea-level record and the Greenland sea-level contribution (Rohling et al., 2019). Shading shows 95% confidence intervals.

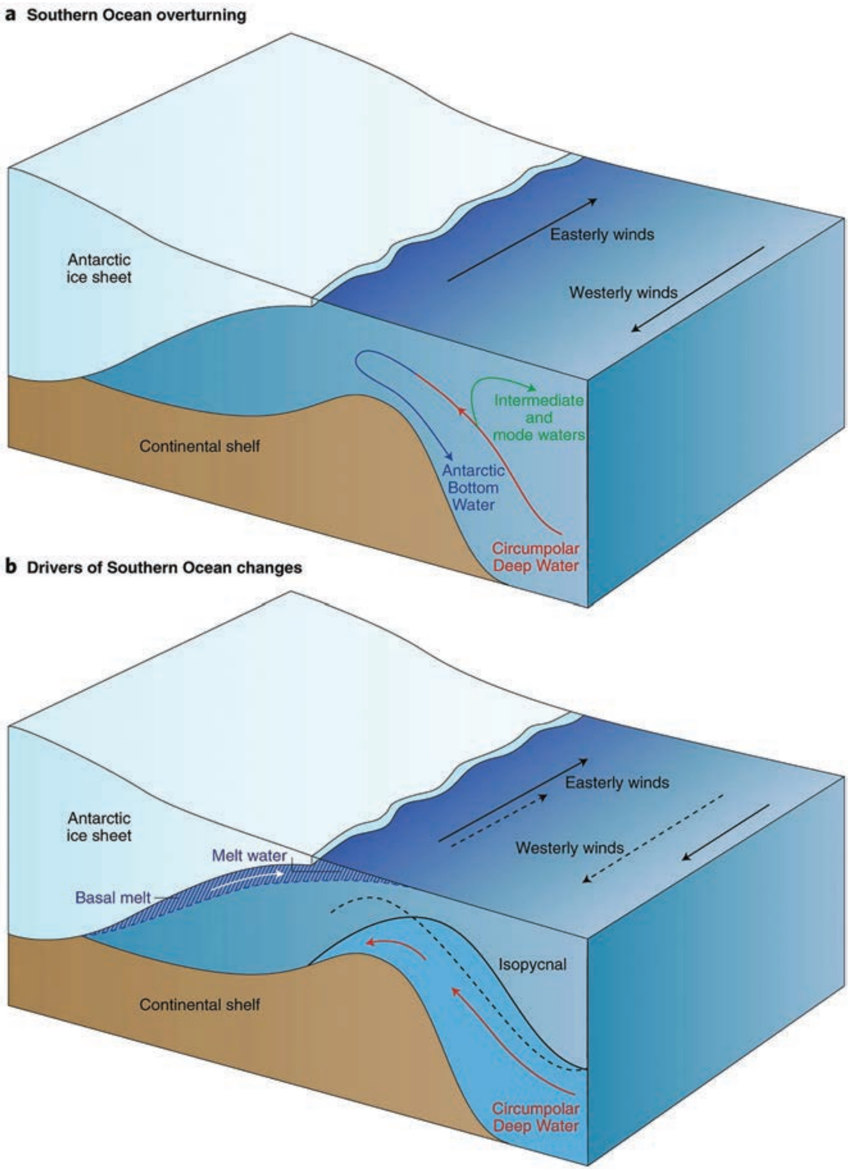

FIGURE 2.5 (a) Present-day conditions in the Southern Ocean showing the divergence between westerly and easterly winds that drives upwelling of CDW, which is transformed into either lighter intermediate and mode waters (upper overturning) or denser AABW (lower overturning); (b) A strengthening and poleward shift of westerly winds, combined with weaker easterly winds (dashed arrows; Bronselaer et al., 2020), causes isopycnals to shoal near the Antarctica, driving more warm water intrusions onto the continental shelf (dashed line). Increased meltwater discharge from the AIS enhances ocean stratification near the surface (figure from Silvano, 2020).

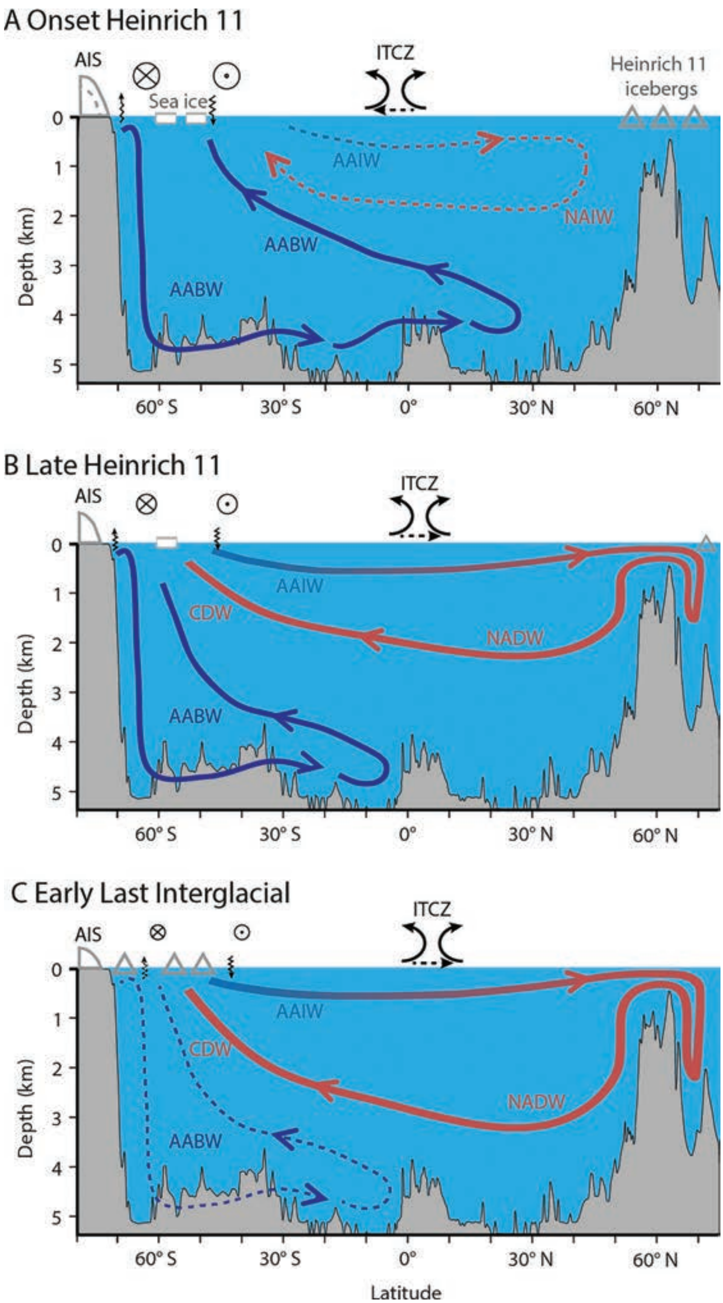

FIGURE 2.6 The bipolar seesaw at the end of the penultimate glaciation 130–135 ka ago, showing the change in AABW formation in response to iceberg discharge (Heinrich stadial 11) in the North Atlantic, which disrupted the AMOC (A–B) and resulted in a build-up of heat in the Southern Hemisphere. (C) Ice core evidence shows substantial ice mass loss from the Weddell Sea sector of Antarctica in response to ocean heat transfer via CDW to the Antarctic margin during the Last Interglacial (figure modified from Turney et al., 2020).

2.4 Summary and Future Directions

The geological record provides observations of the Antarctic Ice Sheet (AIS) and SO during remarkably different climate settings and over a range of timescales. These observations provide constraints for modelling studies to test our understanding of the processes that are relevant to predictions of future Antarctic and SO change. The processes and rates governing AIS mass loss continue to contribute large uncertainties to future sea-level rise projections (Oppenheimer et al., 2019). These processes include those governing ice dynamics, such as marine ice cliff instability and hydrofracturing, the role of subglacial hydrology/hydrogeology, the solid-Earth response to changes in ice mass and feedbacks associated with meltwater and sea ice that can act to moderate oceanic and atmospheric warming.

Ice-sheet growth and retreat since the establishment of the AIS around 40 Ma has increased the extent of interaction between the ice sheet and the ocean. This vulnerability developed through repeated erosion of the Antarctic continent during warm periods across the Oligocene and Miocene, resulting in low-lying topography in West Antarctica and large subglacial basins in East Antarctica (e.g., Wilkes Basin, Aurora Basin; Figure 2.2). The growth of continental shelves and associated expansion of the ice sheet as the climate cooled during the Plio-Pleistocene modified the interaction of the AIS with the ocean. Glacial expansion of the ice sheet calved deep troughs seaward, which today, under modified atmospheric and oceanic conditions, help to facilitate the cross-shelf transport of CDW to the grounding lines of glaciers draining low-lying basins (e.g., Thwaites and Pine Island Glaciers).

Paleo-archives from the Mid-Miocene Climate Optimum, with peaks in atmospheric CO_2 of ~800 ppm (CenCO2PIP Consortium et al., 2023), provide insight into the worst-case SSP5–8.5 future climates, with some differences due to the modern ice sheet being more sensitive to ocean forcing and runaway retreat than for the more stable Miocene Antarctic topography. The CO_2 forcing of the Mid-Pliocene Warm Period (367 ppm; de la Vega et al., 2000) has already been surpassed today. However, the warm Mid-Pliocene provides insight into an equilibrated climate state with smaller ice sheets associated with more southerly SO SST gradients and reduced sea ice relative to present. More recent warm interglacial periods of the Pleistocene, particularly MIS 11 and MIS 5e, experienced strong SO heat build-up associated with the collapse of AMOC and are likely the best short-term (centennial to millennial scale) analogues to current and future anthropogenic climate forcing.

Further international collaboration is necessary across the Antarctic-SO science community to develop new geological archives to understand which vulnerable sectors of Antarctica will contribute to sea-level rise in the near-term, the regional climate and solid-Earth thresholds associated with atmospheric and oceanic forcing, and estimates of the rate of ice sheet change. Other feedbacks requiring further research include the impact of changes in sea ice and meltwater on wider SO ecosystems and climate feedbacks related to changes in the global overturning circulation system.

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